The role of subduction in the evolution of the Apennines foreland basin

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Abstract

Seismic tomographic images of the Mediterranean region lead to the formulation of an hypothesis involving detachment (or alternatively break-off) of subducted lithosphere, propagating horizontally in the along strike direction: lateral migration of slab detachment [Wortel, M.J.R., Spakman, W., (1992). Structure and dynamics of subducted lithosphere in the Mediterranean region. Proc. K. Ned. Akad. Wet. 95, 325-347]. On the basis of this hypothesis variations in vertical loads are expected along strike of the convergent plate margin involved. We investigate whether such variations in vertical loads can be inferred from the Plio-Pleistocene foreland basin along the eastern side of the Apennines. Our aim is to test the hypothesis of a laterally migrating slab detachment beneath Italy. We employ a finite difference method to obtain the flexural downbending of the Adriatic plate resulting from known loads, such as topography and sediments. The unknown slab pull is inferred by constraining the model to fit the horizon that is assumed to reflect flexural downbending in two-dimensional sections perpendicular to the strike of the Apennines. Our method forms a consistent treatment of all loads involved in two-dimensional lithospheric flexure. If we assume that the downgoing Adriatic plate behaves purely elastically, we conclude that there is a trade-off relation between slab pull and elastic plate thickness for sections in north Italy. For these sections gravity anomaly data cannot constrain a best elastic model. If we make the more realistic assumption of a continental lithosphere with a depth-dependent rheology, we obtain a good match for sections in the north. In the south, however, we find that a match cannot be obtained. This may imply that the data reflect more than purely flexural processes — probably thrusting. The conclusion is that a trend in vertical loads along strike of the Apennines cannot be resolved from Plio-Pleistocene deflection data. Therefore, the hypothesis of lateral migration of slab detachment can neither be shown to be consistent nor inconsistent with the observations. © 1998 Elsevier Science B.V. All rights reserved.

Keywords: continental flexure; slab detachment; subduction; foreland basin; Italy

1. Introduction

The Mediterranean region is the locus of convergence between Africa and Europe. This results in a complex pattern of subduction- and collision zones, part of which is formed by the Tyrrhenian–Apennines system (Fig. 1). The origin of this system, with compression in the Apennine thrust belt adjacent to extension in the Tyrrhenian Sea, is related to the westward subduction of the Adriatic plate beneath Italy. In Calabria, in the south, subducted lithosphere is delineated by the hypocentra of earthquakes below the Tyrrhenian basin (e.g.,
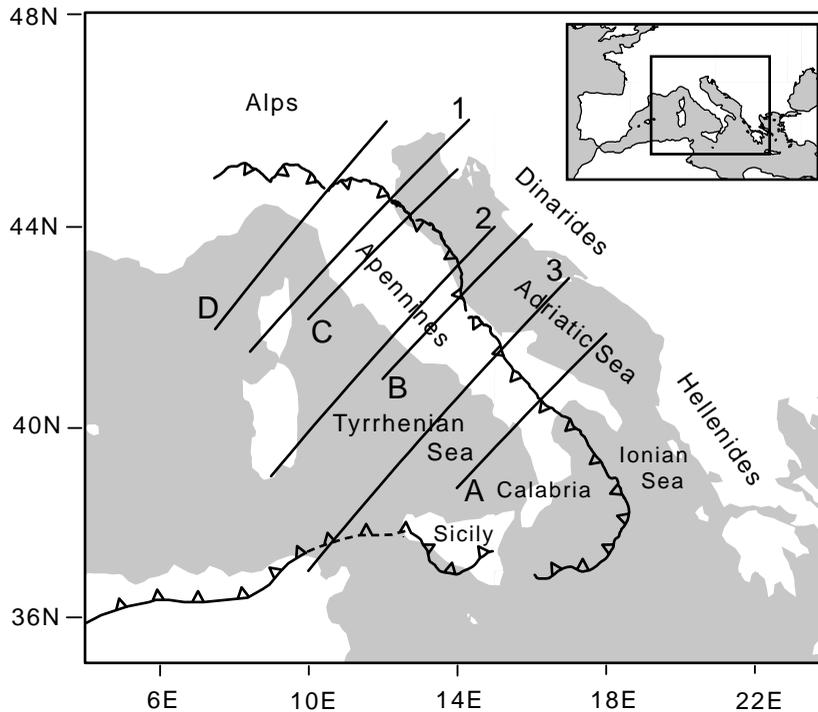

Fig. 1. Map of Italy, barbed line: outermost thrust front, straight lines: the modelled sections (*1,2,3*: this study, *A–D*: data from Royden (1988)).

Ritsema, 1972; Anderson and Jackson, 1987; Giardini and Velona, 1991), but the presence of several hundred kilometers of subducted lithosphere can be inferred along all of the Italian–Adriatic plate boundary (Spakman, 1990; De Jonge and Wortel, 1990; Amato et al., 1993; Spakman et al., 1993). The extensional Tyrrhenian basin is created by roll-back of the subduction zone (Boccaletti et al., 1976; Malinverno and Ryan, 1986; Laubscher, 1988; Doglioni, 1991, 1993). The extensional effect of the roll-back process can be intensified by second order processes, such as the convergence of Africa and Europe (Boccaletti et al., 1982; Mantovani et al., 1993) and gravitational collapse of the post-collisional Alpine crustal wedge (Channell and Mareschal, 1989; Faccenna et al., 1996).

The foreland basin of the Apennine mountains lies along the eastern side of the mountain belt. In this paper the term 'foreland basin' is used to denote the sedimentary basin that lies on the subducting Adriatic plate in front of the advancing Apennine thrust belt (Fig. 2). Although a foreland basin is usually not associated with a subduction zone setting,

this terminology is in agreement with existing literature on this region. The basin is a rather deep basin adjacent to a moderately high mountain belt: the Plio-Pleistocene foreland basin locally contains up to 9 km of sediments, whereas the Apennine mountains are at most 2 km high. Previous studies have shown that the topographic load is insufficient to cause such a deep basin and that additional subsurface loads must be present (Karner and Watts, 1983; Royden and Karner, 1984; Royden et al., 1987). The subsurface loads are most likely associated with subducted Adriatic lithosphere. We consider the Apennine foreland basin to result from flexural downbending of the subducting Adriatic lithosphere, combined with the subsidence caused by the surface loads, i.e. the load of the thrust belt, water and sediments. Variations, in space and/or time, in the load associated with the subducted plate (the slab pull) will, therefore, affect the configuration of the foreland basin. The hypothesis of lateral migration of slab detachment of Wortel and Spakman (1992) predicts such variations in slab pull along strike of the Apennines. This hypothesis is based on tomographic studies of the Mediterranean
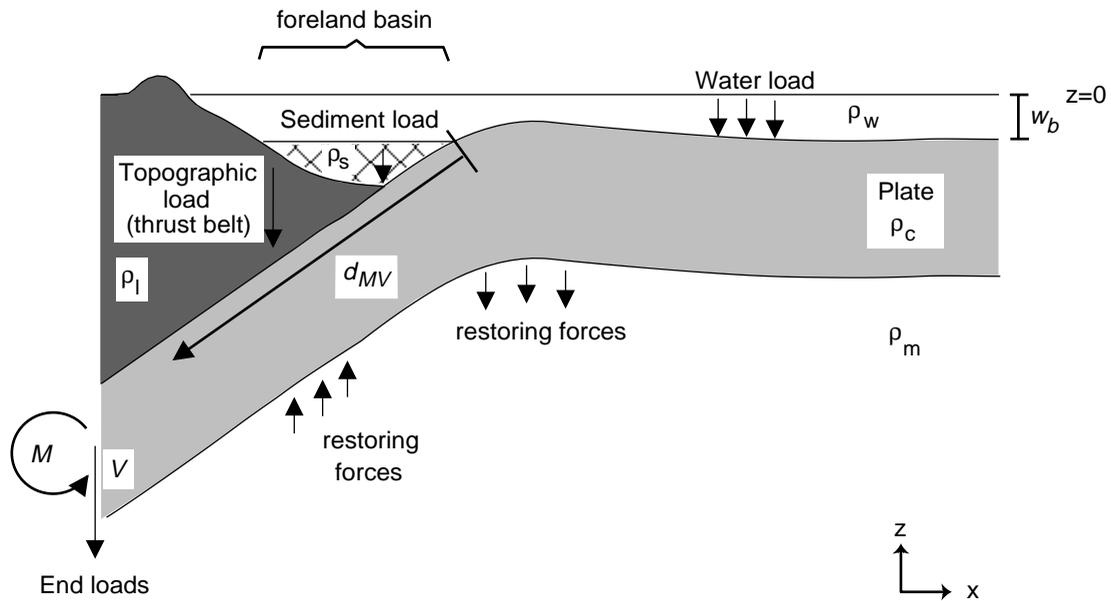
Fig. 2. Loads involved in plate flexure; densities of the different materials are indicated: m mantle, w water, c crust, s sediment, l topography. w_b is the height of the layer above the surface of the flexing plate, in this figure this layer is a water layer, but it can also partly or completely consist of sediments. d_{MV} denotes the location for which values for the end loads are reported. Figure is not to scale.

region (Spakman, 1990), which show evidence for a gap in the subducted Adriatic lithosphere at 100–200 km depth. According to Wortel and Spakman (1992), this detachment of the slab started in the north and migrated south/southeastward with time, causing temporal and spatial variations in the slab pull. The process is illustrated in Fig. 3. As detachment proceeds, the gravitational pull of the detached part of the slab is transferred to the undetached part. This leads to an increase in the effective slab pull exerted by the undetached slab. A deepening of the foreland basin and a more pronounced outward migration of the trench, relative to the situation where only the roll-back process is active, are, therefore, expected above the undetached slab. Where and when the slab is detached, the slab pull decreases, leading to a shallowing of the foreland basin.

It is the aim of this paper to test whether the effects of a laterally propagating tear in the subducting plate are in accordance with observations from the Plio-Pleistocene foreland basin. In particular, we study variations in the vertical loads along the convergent plate margin of Italy. To this purpose the flexural response of the foreland basin is studied in two-dimensional sections perpendicular to the strike of the Apennines (Fig. 1). In this way, NW–SE trends can be investigated. By incorporating published geophysical data, for example, crustal thickness and topography, and constraining the model to fit deflection and gravity anomaly data, the unknown slab pull is quantified for each section. Rather than an attempt to discriminate between various proposed tectonic scenarios for the Italian region, our study should be viewed as an attempt to test the predictions of one of these scenarios, the hypothesis of lateral migration of slab detachment. The implication of this approach is that the hypothesis will be rejected if the results of our work fail to match the predictions. If, however, the hypothesis survives our test, we recognize the possibility that alternative scenarios may also explain the observations we use. In that case, further tests will have to be developed to discriminate between competing hypotheses.

In a similar study, Royden (1988) matched results from a two-dimensional, elastic model to observations of plate flexure and gravity anomalies, for four sections across the Apennine foreland basin. The values reported by Royden (1988) for the deep load
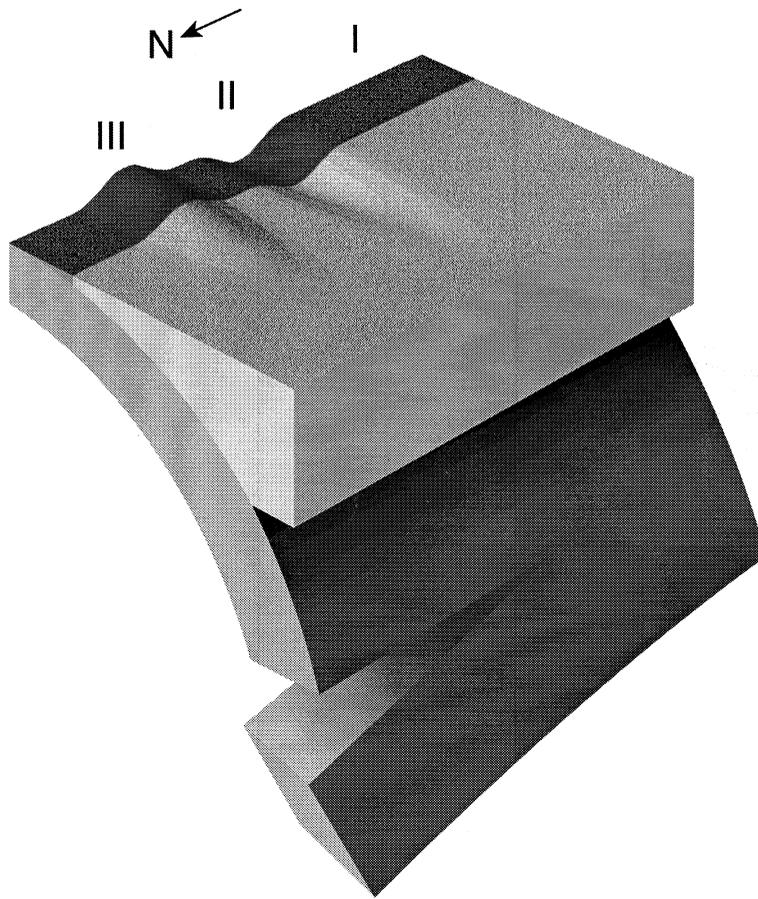
Fig. 3. Illustration of vertical motions resulting from lateral migration of slab detachment. A tear in the slab propagates from north to south. (I) Undisturbed situation; (II) The tip of the tear. The weight of the detached part of the slab is transferred to the undetached part, this leads to an increase in the effective slab pull. This is reflected in a deeper deflection of the surface. (III) Detached slab. The slab pull decreases, leading to a rebound of the surface.

can, however, not be compared between the different sections, since the load was not measured at the same point along the deflection profiles. As such, these results give no indication of spatial variations in slab pull. We will, therefore, re-evaluate the models of Royden (1988), making improvements and extensions on the modelling approach.

The main conclusion that can be drawn from the results of the present study is that the hypothesis of lateral migration of slab detachment cannot be shown to be consistent or inconsistent with the flexural response recorded in the Plio-Pleistocene foreland basin.

2. Modelling method

2.1. Geometry

The origin of the Apennine foreland basin is related to the flexural downbending of the Adriatic lithosphere, which is maintained by the deep load associated with the subducting plate in combination with the load of material overlying the plate. To obtain a quantification of all vertical loads, we employ a two-dimensional model (Fig. 2) to sections perpendicular to the strike of the Apennines. Along strike variations can then be detected by comparing the results for the different sections.

There are various arguments for a continental character of the Adriatic lithosphere. It has a crustal thickness of approximately 30 km (Nolet et al., 1978; Meissner et al., 1987; Ravnik et al., 1995) and a lithospheric thickness of about 100 km (Calcagnile and Panza, 1981; Suhadolc and Panza, 1989). Also, the low to normal heat flow values are in agreement with a continental lithosphere (Čermák, 1979; Loddo and Mongelli, 1979). The continental character is also supported by it probably originating from the former passive continental margin of Africa (Channell et al., 1979). There are indications that the continental lithosphere is involved in the Adriatic subduction. From tectonic reconstructions for the Mediterranean region (Dewey et al., 1989), we deduce that the depth of the ocean–continent boundary is well below our model. Subduction of a continental Adriatic lithosphere may have taken place by a delamination process, in which the upper crustal parts and the sediments are sheared off to form thrust sheets, while the lower crust with the lithospheric part of the mantle subducts (Bally et al., 1986; Panza and Suhadolc, 1990; Serri et al., 1993). In agreement with this scenario, the Apennines are for the greatest part composed of continental rocks, only the uppermost units (Ligurides) are oceanic.

The Adriatic plate is covered by Mesozoic–Cenozoic sediments of variable thickness; on average the sedimentary layer is 10 km thick (Morelli et al., 1969; Nolet et al., 1978; Calcagnile et al., 1982; Mostardini and Merlini, 1986). In this study these old sediments are considered to be part of the crust.

2.2. Lithospheric flexure

For each section, the deflection of the Adriatic lithosphere is calculated from the one-dimensional flexure equation:

$$\frac{\mathrm{d}^2}{\mathrm{d}x^2} D \frac{\mathrm{d}^2 w}{\mathrm{d}x^2} + (\rho_\mathrm{m} - \rho_\mathrm{o})gw = q \tag{1}$$

in which w is the deflection, q the vertical load, D the flexural rigidity, ρ_m and ρ_o the densities of the mantle and of material overlying the lithosphere, respectively, and g the gravity acceleration. In our frame of reference, forces are positive upward and moments are positive counterclockwise. The equation is solved with a finite difference method. By using this equation, we implicitly assume that the length scale of variations in loads and rheology out of the plane of our section is significantly larger than the horizontal extent of the section. Also, the thickness of the model plate has to be much smaller than the wavelength of the flexure (thin plate approximation). The deflection of the top of our model plate is fitted to the base of the Plio-Pleistocene foreland basin of the Apennines. Model parameters (e.g. loads, densities, rheologies), which are not closely constrained by observations are being varied systematically in the modelling. A grid search technique is adopted to identify those model parameters which minimize the rms-difference between observed and computed deflections. We report the resulting best-fitting model parameters. We also show results of additional experiments where we fix all parameters, except one, to give the reader a sense of the sensitivity of our results to that specific parameter.

2.3. Loads and restoring forces

Fig. 2 summarises the vertical loads involved in lithospheric flexure at a convergent plate boundary. The surface loads are the topography, the sediments filling the basin and a water and/or sediment layer. It is assumed that these loads are completely supported by the subducting plate, which implies that an overriding plate is either not present in the studied region or mechanically too weak to support vertical loads. The boundary between the thrust belt and sediments is often diffuse, since sediments of former foreland basins can be incorporated in the advancing thrust belt. The deflected surface is covered by a layer of water and/or sediments of thickness w_b. Whether the sediments are deposited after or prior to flexure does not make a difference in our end-situation modelling (Fig. 4). The subsurface load (or deep load) is associated with the subducting plate. To limit the size of the model, this load is simulated by applying a bending moment M and a vertical force V at an artificial slab end; M and V are here referred to as 'end loads'. The unknown end loads are constrained in the modelling procedure. Besides vertical loading, the flexing plate can also be subject to (changes in) horizontal tension or compression (Cloetingh et al., 1985; Peper, 1993). For reasonable values of intraplate stress (tens of MPa) (Bott, 1991, 1993;
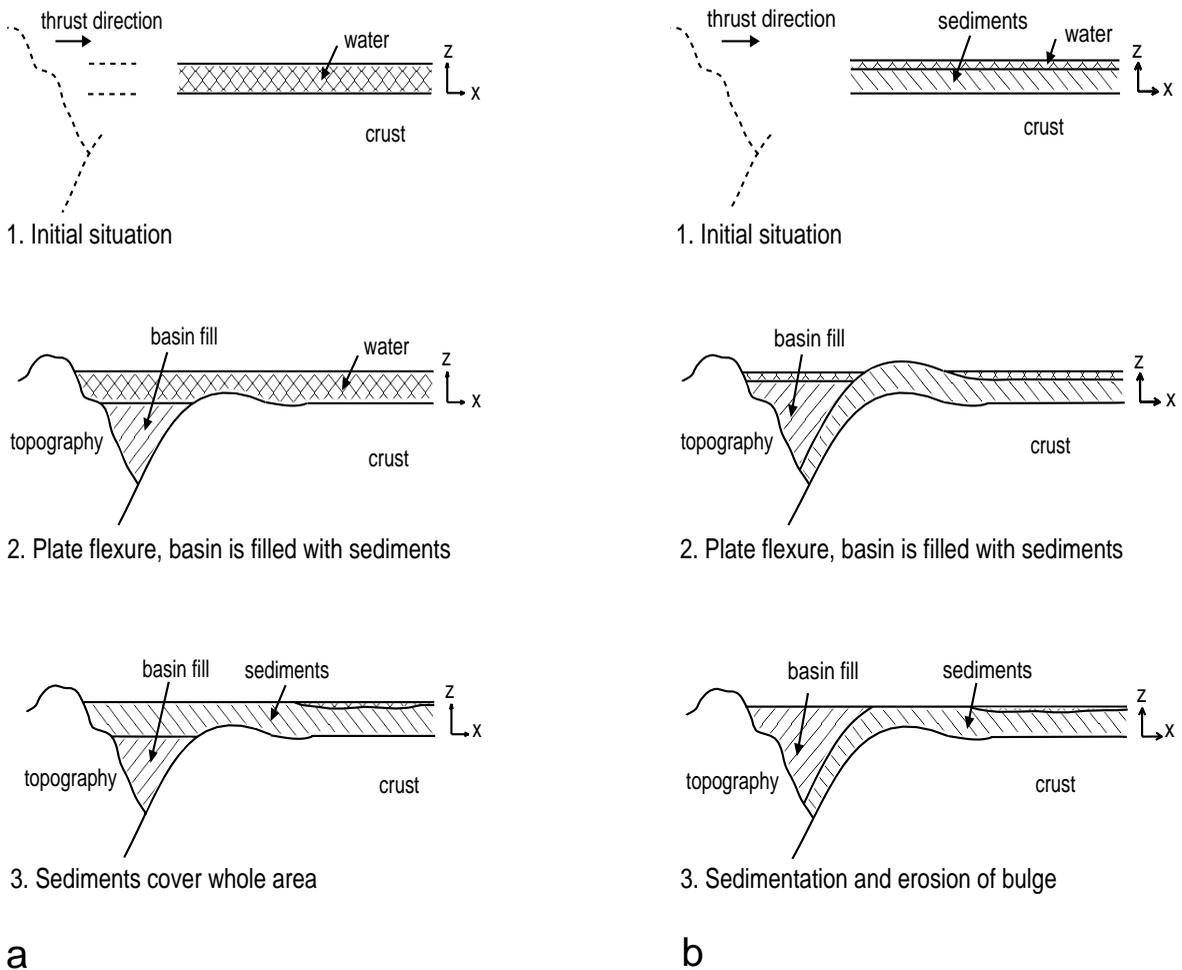
Fig. 4. The flexing plate is covered by a layer of sediments, these can be deposited after flexure (a) or prior to flexure (b). If it is assumed that the densities of the sediments are equal on average, the end situation of both figures is the same.

Wortel et al., 1991) and thick plates, the effect on the calculated deflection is, in general, not very large. In this study intraplate stress is, therefore, not taken into account.

Restoring forces arise from buoyancy, resulting from the replacement of mantle material underlying the plate by material of a lower density overlying the plate ($(\rho_m - \rho_o)gw$ in Eq. 1).

The main difference between the elastic model of Royden (1988) and the elastic model employed in this study is expressed in the way densities or density differences are handled (Fig. 5). Royden (1988) assumes that, initially, the top of the plate lies at a depth w_0 and that the plate is covered with water.

The water overlying the plate is taken into account in the flexural calculations. Therefore, her restoring force equals $(\rho_m - \rho_w)gw$ and the topographic and sedimentary contributions to the loading of the plate below sea level are determined by the density difference with water (e.g. $(\rho_s - \rho_w)$). However, according to her Fig. 1 and the description of the load of the infilling material, Royden (1988) employs the density difference with water for material between sea level and w_0, and not for material below w_0 (Fig. 5). This approach is not consistent. It results in a higher density for the infilling material and, therefore, a larger load. Likewise, it is assumed in the study of Royden (1988) that the plate is cov-
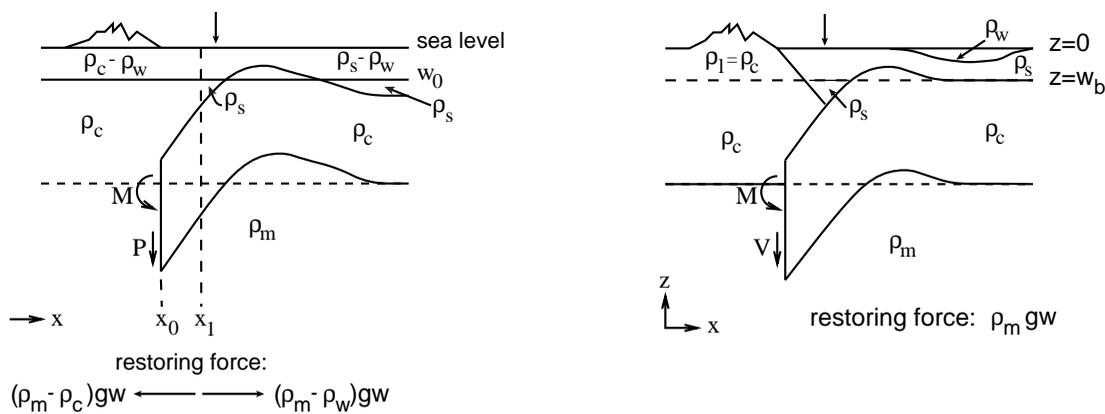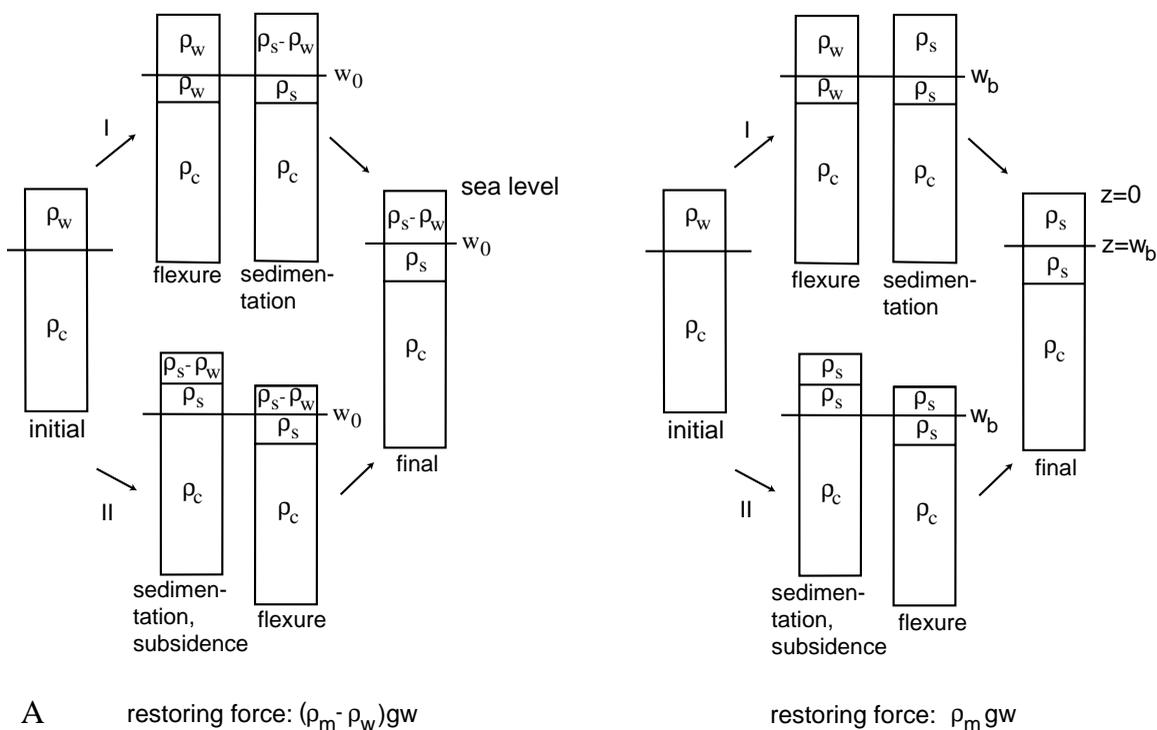
Fig. 5. Schematic representation of two models, left: Royden (1988), right: this study. (a) Densities in the initial and final situation. The two scenarios (I and II) lead to the same final situation. The last column includes the subsidence caused by sediments added in the previous step. The left figure shows that Royden (1988) is not consistent with the densities used in determining the loads and the restoring force. (b) The models. The arrow at the top indicates the location of the column shown in (a). The left figure is taken from Royden (1988). Densities of the different materials are indicated: m mantle, c crust, s sediment, l topography, w water.

ered with material with a crustal density ρ_c towards the subduction zone. In this case the restoring force equals $(\rho_m - \rho_c)gw$ and the loads on the plate are determined by the density difference with the crustal material. For material between sea level and w_0, Royden (1988), however, still uses the density difference with water (see also Kruse and Royden, 1994). This again results in a load which is too large. In the model employed in this study we do not incorporate the material covering the plate through the restoring force, but explicitly include the load of all material above the plate in the loading term. In our case, the restoring force equals $\rho_m gw$. In this way we avoid an inconsistent treatment of the loads.

2.4. Rheology

We first model the configuration of the foreland basin with an elastic plate rheology; this facilitates a comparison with the results of Royden (1988). Models employing an elastic plate can, in general, match first order characteristics of observed deflection profiles (e.g., Caldwell et al., 1976; Jordan, 1981; Karner and Watts, 1983). Combined with the relatively simple calculations, this explains the frequent usage of an elastic plate model. However, the representation of continental lithosphere by an elastic plate of constant thickness means a substantial approximation, although relations between effective elastic plate thickness and lithospheric parameters can be derived (Burov and Diament, 1992; Ranalli, 1994).

Therefore, we approximate the rheology for the continental Adriatic lithosphere (hereafter referred to as 'continental rheology') by an elastic–plastic rheology (Fig. 6). The yield stress in the brittle regime follows Byerlee's law, ductile yielding is described by power law creep. The deflection of a plate with a continental rheology is obtained by iterative solution of the flexure equation (Eq. 1) with an effective elastic flexural rigidity:

$$D_{\text{eff}}(w) = \frac{M_{\text{tot}}}{-w''} = \frac{\int_H \sigma(z)z\,\mathrm{d}z}{-w''} \qquad (2)$$

where $-w''$ is curvature, $\sigma = \sigma_{\text{horz}} - \sigma_{\text{vert}}$, z is depth and H is the thickness of the mechanically strong part of the lithosphere (e.g. Burov and Diament, 1992; see also Bodine et al., 1981; McAdoo et al.,
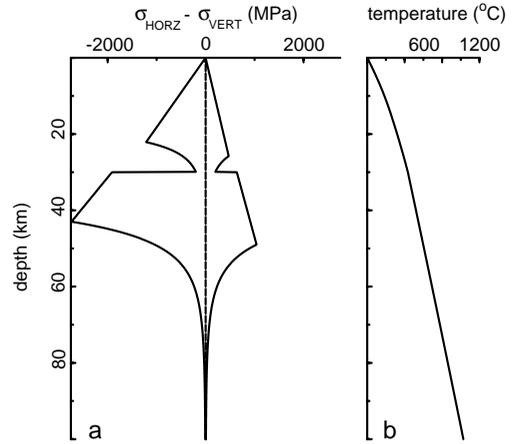
Fig. 6. (a) Continental yield rheology for $\dot{\epsilon} = 10^{-14}$ 1/s and geotherm as in (b). Compression is negative. (b) Continental geotherm following Chapman (1986) (surface heatflow 55 mW/m², Moho depth 30 km).

1985 for the case of an oceanic rheology). This is purely a modelling parameter representing the rigidity of an elastic plate which adequately simulates the observed bending behaviour. In general, D_{eff} will vary laterally. The effective elastic flexural rigidity depends on the variation of yield stress with depth, and therefore, on the thermal state, crustal thickness and strain rate, and on plate curvature and stress (Karner et al., 1983; McNutt et al., 1988; Burov and Diament, 1992, 1995). Increased yielding through an increase in curvature or horizontal stresses will lead to a decrease in effective rigidity. Mechanical decoupling of the crust and mantle at substantial amounts of bending (De Rito et al., 1986; McNutt et al., 1988; Burov and Diament, 1995) is implicitly incorporated in our simulations. Rheological parameters that were used in this study are listed in Table 1.

2.5. Gravity anomalies

An additional constraint on the different models can be provided by requiring calculated gravity anomalies to fit observational gravity data. Bouguer gravity anomalies are calculated using the polygon method of Talwani et al. (1959). As a reference model, the undisturbed situation, i.e. the foreland in the models, is used. We do not a priori incorporate slab detachment in our models and, therefore, we use a continuous subducted plate in the calculation of

Table 1
Parameters defining lithospheric rheology

Parameter	Symbol	Value	Unit
Moho depth [a]		30	km
Surface heat flow [b]	Q_s	55	mW m^{-2}
Young's modulus	E	10^{11}	Pa
Poisson's ratio	ν	0.25	
Strain rate	$\dot{\epsilon}$	10^{-14}	s^{-1}
Coefficient of friction	μ	0.577	
Pore-fluid factor	λ	0	
Densities			
Crust	ρ_{crust}	2800	kg m^{-3}
Mantle	ρ_{mantle}	3250	kg m^{-3}
Universal gas constant	R	8.31	J K^{-1} mole^{-1}
Power law creep–crust [c]			
Pre-exponent [d]	A_{pl}	1.23×10^{-23}	s^{-1} Pa$^{-3.2}$
Activation energy	Q_{pl}	238.6	kJ mole^{-1}
Power	n	3.2	
Power law creep–mantle [e]			
Pre-exponent [d]	A_{pl}	2.87×10^{-25}	s^{-1} Pa$^{-4.48}$
Activation energy	Q_{pl}	498	kJ mole^{-1}
Power	n	4.48	

[a] Nolet et al., 1978; Meissner et al., 1987; Ravnik et al., 1995.
[b] Čermák, 1979.
[c] Dry anorthosite (Shelton and Tullis, 1981).
[d] Modified to ensure validity of PL creep for a general state of stress (Ranalli, 1987) and for a thin flexing plate.
[e] Wet Aheim dunite (Chopra and Paterson, 1981).

gravity anomalies. In this aspect our method differs from Royden (1988), where a fit to the gravity data is obtained by varying the location of an effective end of the subducted lithosphere. The contribution to the gravity anomaly of slab material that is possibly present beyond the effective slab end is neglected in Royden's approach.

3. Data

3.1. Model input data and constraints

The calculated deflection is fitted to the base of the Plio-Pleistocene foreland basin. The thickness of the Plio-Pleistocene sediments is taken from the map 'The neotectonic model of Italy' (Ambrosetti et al., 1987) for sections 1, 2 and 3. The data are averaged over an interval of 0.2 degrees and recalculated to depth below the surface $z = 0$ (Fig. 2). The loads of the thrust belt and the water layer are constrained by topographic and bathymetric data, respectively (ETOPO5), while the sediment load is constrained by the base of the Plio-Pleistocene foredeep. For sections A through D, deflection and gravity data from Royden (1988) are used.

3.2. Regional considerations

It is assumed that the horizon that at present forms the base of the Plio-Pleistocene foredeep was initially deposited horizontally. In this case, the present shape of the base of Pliocene level is the result of processes acting from the Early Pliocene to present. If the process of slab detachment would be occurring at present, differences in slab pull along strike of the Apennines could be expected. However, it is possible that the slab has already been detached along most of Italy from Late Miocene to present (Wortel and Spakman, 1992). Also, detachment possibly occurred very recently (Quaternary) beneath the southern Apennines and Calabria (Westaway, 1993; Hippolyte et al., 1994). Therefore, depending on the timing of slab detachment, the base of

Pliocene level at different locations along the coast of Italy, was deposited before, during or after slab break-off. The base of Pliocene level experienced a different tectonic history in these different settings, which influences the present configuration of this level. Therefore, different loads are necessary to reproduce a post- or a pre-slab detachment deposited level. Along strike differences in slab pull and/or surface loads (i.e. sediments, topography and water) are therefore expected. Since slab detachment probably occurred relatively recent in south Italy, it can be expected that in the south the base of Pliocene level has been deposited before slab break-off, while in the north this same level may have been deposited after slab detachment occurred.

The Apennine thrust belt-foreland basin system migrates eastward with time, the amount of migration increases towards the south (Bally et al., 1986; Boccaletti et al., 1990a). The cause of roll-back could be an increase in slab pull, in which case the effects of changes in vertical loads and of horizontal migration are coupled, or it could be escape tectonics (Mantovani et al., 1993). Outward migration will cause a deepening of the sediment horizons in the basin. If this effect is not taken into account, it can be expected that a larger amount of migration will be reflected in a larger modelled loading. Since we do not include roll-back in our models, we have to bear this in mind when interpreting the results.

We also have to keep in mind that in the modelling it is implicitly assumed that the base of Pliocene is only deformed by flexure. It is, however, possible that this level is also disturbed by thrusting (Ricci Lucchi, 1986; Ori et al., 1986).

4. Results

4.1. Elastic rheology

4.1.1. Deflection

To determine the best fit to the base of the Plio-Pleistocene foredeep for each of the seven sections shown in Fig. 1, the following quantities were varied: elastic plate thickness, sediment density, the depth of the baseline, and the end loads, M and V. In most of the models a sediment density of 2300 kg/m^3 gave the best results. This agrees with the average den-

Table 2
Fixed input parameters

Parameter	Symbol	Value	Unit
Moho depth		30	km
Gravity acceleration	g	9.81	m s^{-1}
Densities			
Crust	ρ_c	2800	kg m^{-3}
Mantle	ρ_m	3250	kg m^{-3}
Water	ρ_w	1000	kg m^{-3}
Topography	ρ_l	2800	kg m^{-3}
Sediment	ρ_s	2300	kg m^{-3}

sity found for Plio-Pleistocene sediments, which is around 2250–2350 kg/m^3 (Mostardini and Merlini, 1986). Therefore, all figures will show results for a sediment density of 2300 kg/m^3. Values of fixed input parameters are shown in Table 2. The results of modelling with a completely elastic plate are shown in Fig. 7. For these best fits, the values of plate thickness, baseline, the end loads and the surface load are given in Table 3. In this study, these end loads are always measured at the same location to allow a comparison between different model results. Values are given for a point located at a fixed distance, measured along the plate, from the first point of zero deflection (d_{MV} in Fig. 2). Since the magnitude and sign of bending moment and shear force vary with distance along the bending plate, it is possible that the sign of the measured values differs from that of the applied end loads. A measure for the surface load is obtained by integrating the loads of topography, sediments and water over plate length.

For sections 1, C and D a reasonable fit to the base of the foredeep can be obtained with different elastic plate thicknesses. From the values in Table 3 it can be seen that a trade-off relation between elastic plate thickness and the end loads exists. Thus, for these sections the deep load cannot be constrained from modelling with an elastic plate. In the next section it is demonstrated that this situation does not improve by incorporating gravity data in the models. For sections 2 and B the calculated deflection fails to fit the high in the data, even for a very low value of elastic plate thickness of 6 km. The issue of the sharp highs will be considered in more detail in Section 5. The calculated deflection well fits the data for sections A and 3.
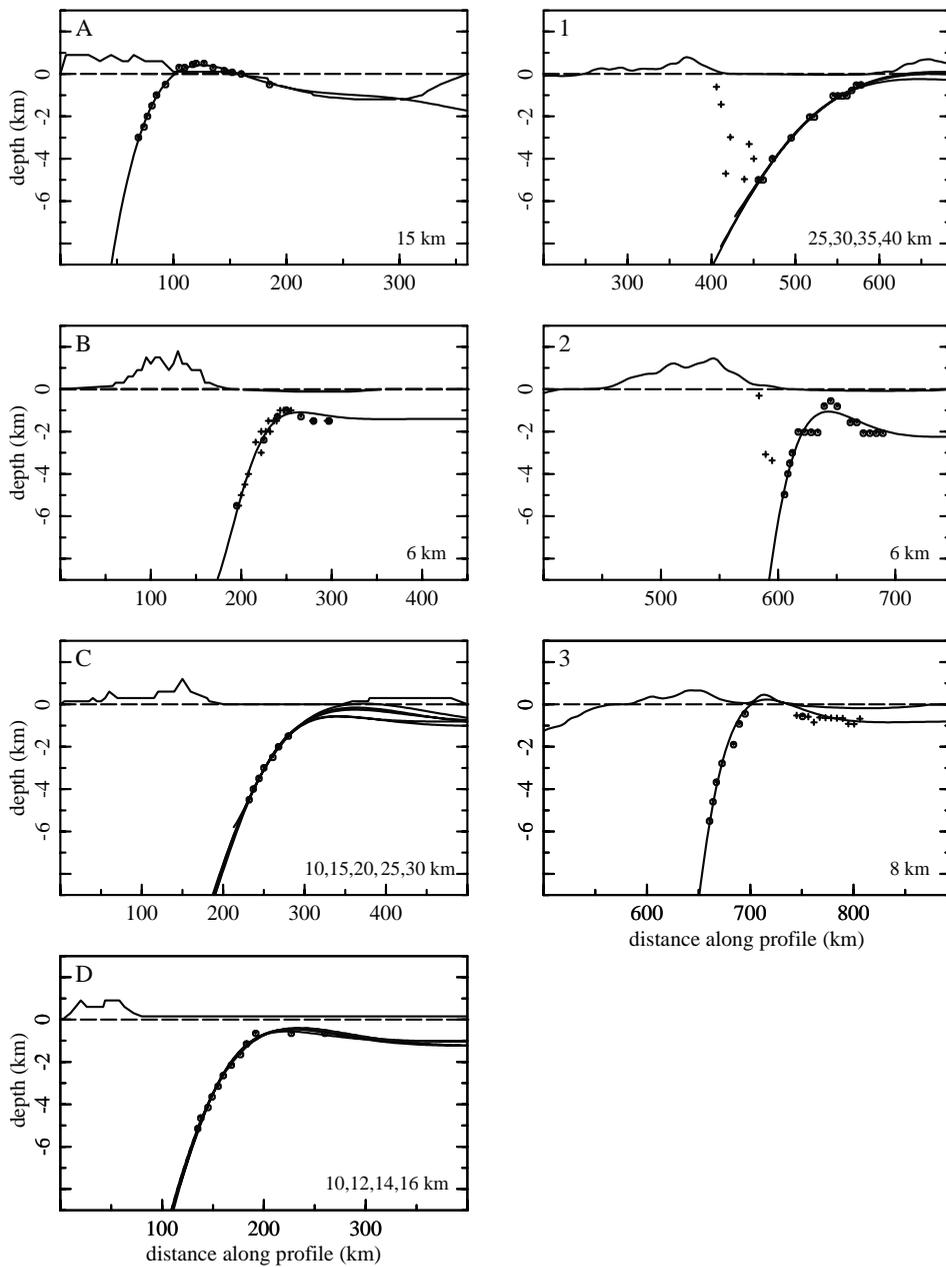
Fig. 7. Solutions with an elastic rheology for each of the sections shown in Fig. 1. The values for elastic plate thickness are shown in each frame. The base of the Plio-Pleistocene foredeep is indicated with small circles. The crosses also indicate base of Pliocene: on section B for parallel profiles, on 1 and 2 the overthrusting side and on 3 values which were not used in calculating the rms-difference with the modelled deflection.

The best fitting model for each plate thickness is determined by the lowest value for the rms-difference between the calculated deflection and the base of Pliocene level. This method allows a good de-

termination of the values of the end loads (Fig. 8). The values for the rheological parameters had to be adopted in the modelling ($E = 1 \times 10^{11}$ Pa, $v = 0.25$). In general, variation of these values either

Table 3
Results for modelling with an elastic plate (V and M measured at $d_{MV} = 60$ km, see Fig. 2)

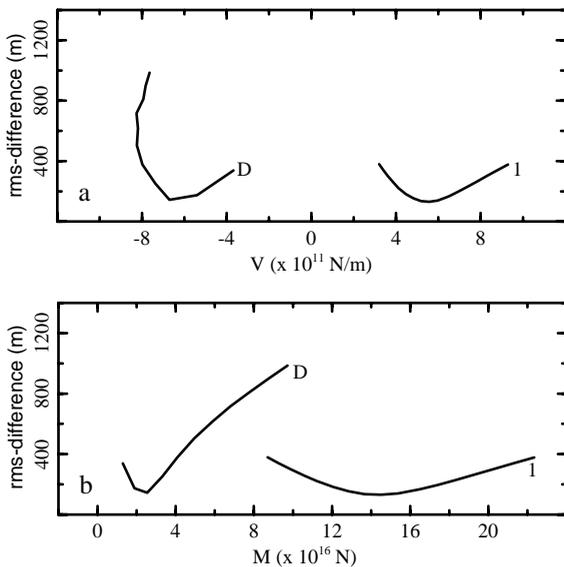| Section	Elastic plate thickness (km)	Baseline (m)	V (10^{11} N m^{-1})	M (10^{16} N)	Surface load, $\int q_{surf}\, dx$ (10^{13} N m^{-1})	rms-difference, $\left[\dfrac{1}{n}\sum\limits_{n}(y_n - y_{dn})^2\right]^{\frac{1}{2}}$ (m)
A	15	2200	33.4	35.5	−3.78	99
B	6	1400	−6.39	−0.48	−0.52	232
C	10	800	−4.41	0.54	−0.42	74
	15	1000	−2.91	3.34	−0.84	63
	20	800	−0.41	6.79	−1.51	57
	25	1000	3.49	14.6	−2.62	56
	30	1000	7.77	23.0	−3.26	58
D	10	1000	−7.80	1.06	−0.37	152
	12	1000	−6.70	2.54	−0.43	144
	14	1200	−5.34	5.02	−0.59	148
	16	1200	−3.68	6.90	−0.73	147
1	25	600	1.35	6.16	−0.65	133
	30	400	2.91	9.01	−0.75	133
	35	400	5.56	14.5	−1.03	131
	40	600	8.72	22.4	−1.79	130
2	6	2200	−5.92 [a]	2.01 [a]	−0.52	338
3	8	800	0.80	6.50	−1.78	320

[a] $d_{MV} = 45$ km.

Fig. 8. Minimum in rms-difference between calculated and observed deflection for two elastic models: *D* (plate thickness 12 km) and *1* (35 km). The values for the end loads, *M* and *V*, are measured at $d_{MV} = 60$ km. (a) rms-difference against shear force *V*, (b) rms-difference against moment *M*.

hardly affects the fit or leads to a worse fit. The results are most sensitive to the choice of Young's modulus *E*.

4.1.2. Gravity anomalies

By two-dimensional modelling of gravity anomaly data besides deflection data, it may be possible to distinguish between different elastic models that fit the deflection data (e.g. sections 1, C and D). While the data describing the base of the foredeep constrain a laterally limited portion of the calculated deflection profile, observed gravity anomalies are available for a wider region, thus possibly allowing a further constraint on the model. On average the range of the measured bouguer anomalies is from −80 to 40 mGal. If we assume an uncertainty in the order of 20 mGal (due to errors in the density for the bouguer correction, instrumental and reading errors) (see also Wessel and Watts, 1988; Keating, 1995) the data contain a significant signal. The bouguer gravity signal over this subduction zone is for the greatest part determined by the bending of the subducting plate. This bending determines the sedimentary basin
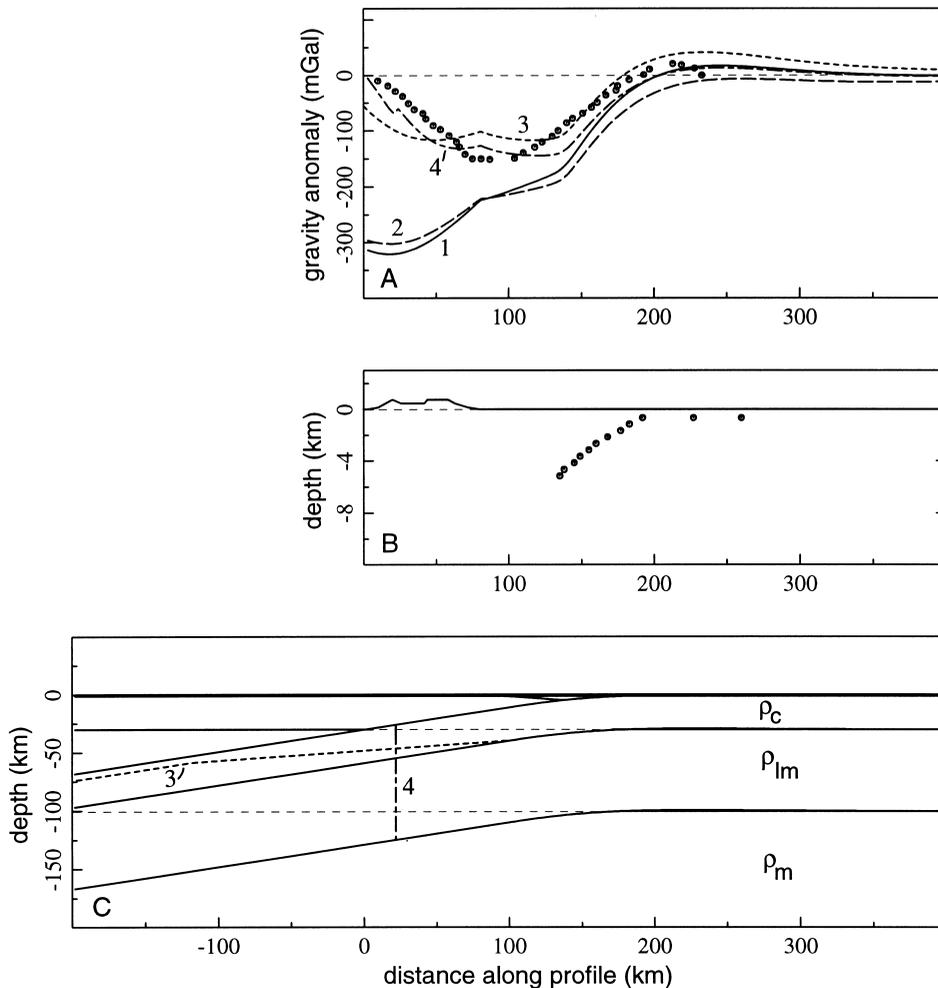
Fig. 9. Effect of variations in crustal thickness, extrapolation and density of the lithospheric mantle on the bouguer gravity anomaly for section D. (a) Bouguer gravity anomaly, small circles: data from Royden (1988), lines: calculated (elastic thickness 14 km, crustal thickness 30 km). $\rho_c/\rho_{lm}/\rho_m$: densities of crust/lithospheric mantle/mantle. (1) $\rho_c/\rho_{lm}/\rho_m = 2800/3450/3250$ kg/m^3, (2) $\rho_c/\rho_{lm}/\rho_m = 2800/3250/3250$ kg/m^3, (3) decreasing crustal thickness, (4) effective end of subducted lithosphere. (b) Deflection data. (c) The (extrapolated) models.

and causes lateral density differences (e.g. between crust and mantle, sediment and crust). Similar deflection profiles will, therefore, lead to similar gravity anomalies. For example, for profile 1 the difference in calculated bouguer anomaly between the models with an elastic plate thickness of 30 and 40 km is at most 10 mGal. Assumptions concerning the values of densities, thickness of the crust and lithosphere and extrapolation of the deflection profile to greater depths have a much larger effect on the calculated anomaly than small differences in deflec-

tion. Fig. 9 illustrates this argument for variations in crustal thickness and extrapolation. Royden (1988) obtained a fit to the gravity data by varying the location of the effective end of the subducted lithosphere; slab material that is possibly present beyond the effective slab end does not contribute to the calculated anomaly. In the case of our example (Fig. 9) this approach leads to a reasonable fit. However, the presence of subducted material down to large depths in this region is apparent from tomographic studies (Spakman, 1990; Spakman et al., 1993; Amato et

al., 1993). It is not clear why slab material beyond the slab end should not contribute to the bouguer anomaly. The calculated bouguer gravity anomalies include the effect of the relatively thin crust of the Tyrrhenian Sea (related to its extensional origin). This positive contribution was calculated using the depth of the Moho from the map 'The neotectonic model of Italy' (Ambrosetti et al., 1987).

Straightforward calculation of bouguer gravity anomalies from the obtained deflection models does not lead to a fit to the observed gravity data. We examined reasonable variations in different parameters: sediment density, crustal thickness, density of the lithospheric mantle, extrapolation of the deflection profile to greater depths and location of the boundary between thrust belt and sediments. Since similar deflection profiles lead to similar gravity anomalies, their difference is insignificant compared to the much larger effect of these variations. Therefore, it is concluded that including gravity data in the analysis does not help to discriminate between the different elastic models that fit the deflection data.

4.2. Continental rheology

From the results of modelling with an elastic rheology we concluded that for some sections a fit to the deflection data can be obtained for a range in elastic plate thicknesses, while for other sections a fit can only be obtained for a very thin plate. Employing a more realistic continental rheology may possibly further constrain the solution and allow us to quantify the end loads. For each of the seven sections the best fit to the base of Pliocene level for a model with a continental rheology is shown in Fig. 10. The modelling parameters are given in Table 4. Overall it can be stated that the plate with a continental rheology gives a worse fit to the deflection data than an elastic plate. For the large curvature values associated with the base of the foreland basin the effective elastic plate thickness locally rapidly reduces, this leads to an increase in bending. For smaller amounts of bending a flexing continental lithosphere is mechanically stiff. The continental model can, therefore, not reproduce the sharp highs on some of the sections (e.g. B, 2 and 3). A good fit is obtained for those sections which also allowed multiple elastic solutions (i.e. C, D and 1). However, due to the large difference in

the quality of the fit to the data between the different sections (Table 4) a comparison of the obtained loads is not very meaningful. This implies that it cannot be shown whether laterally propagating slab break-off beneath Italy is compatible or incompatible with Plio-Pleistocene deflection data.

Our solutions are sensitive to the choice of values for the different rheological parameters, since these determine the variation of yield stress with depth (and therefore affect the effective elastic flexural rigidity). We considered variations in geothermal gradient (determined by surface heatflow), crustal thickness and ductile flow laws. The plate behaves stronger for an higher ductile yield stress or a smaller crustal thickness (the latter through a reduction of the lower crust ductile zone). This results in a worse fit in all cases. For a number of sections an improved fit could be obtained with a weakened plate. This is the case for sections 2, A, B and D with an higher surface heatflow and for sections B and D with a lower ductile yield stress. This agrees with our observation that the continental plate behaved too stiff to simulate the deflection data of central Italy. Considering the best fits from these variations does not significantly change the results of Fig. 10. Therefore, we conclude that the hypothesis of lateral migration of slab detachment cannot be shown to be consistent nor inconsistent with the flexural response recorded in the Plio-Pleistocene foreland basin.

5. Discussion

5.1. Causes for discrepancies between model results and data

The sharp highs observed in the deflection data for three of the sections (2, 3 and B) pose a problem in the modelling. For models with an elastic rheology a more or less reasonable fit could only be obtained for very thin plates; however, a small elastic plate thickness is not immediately expected on the basis of the continental character of the Adriatic lithosphere. The bending behaviour of the continental plate model is too stiff to simulate an high and sharp bulge. Nevertheless, a sharp high is observed in these sections and different explanations are possible. One explanation is that an high flexural bulge is main-
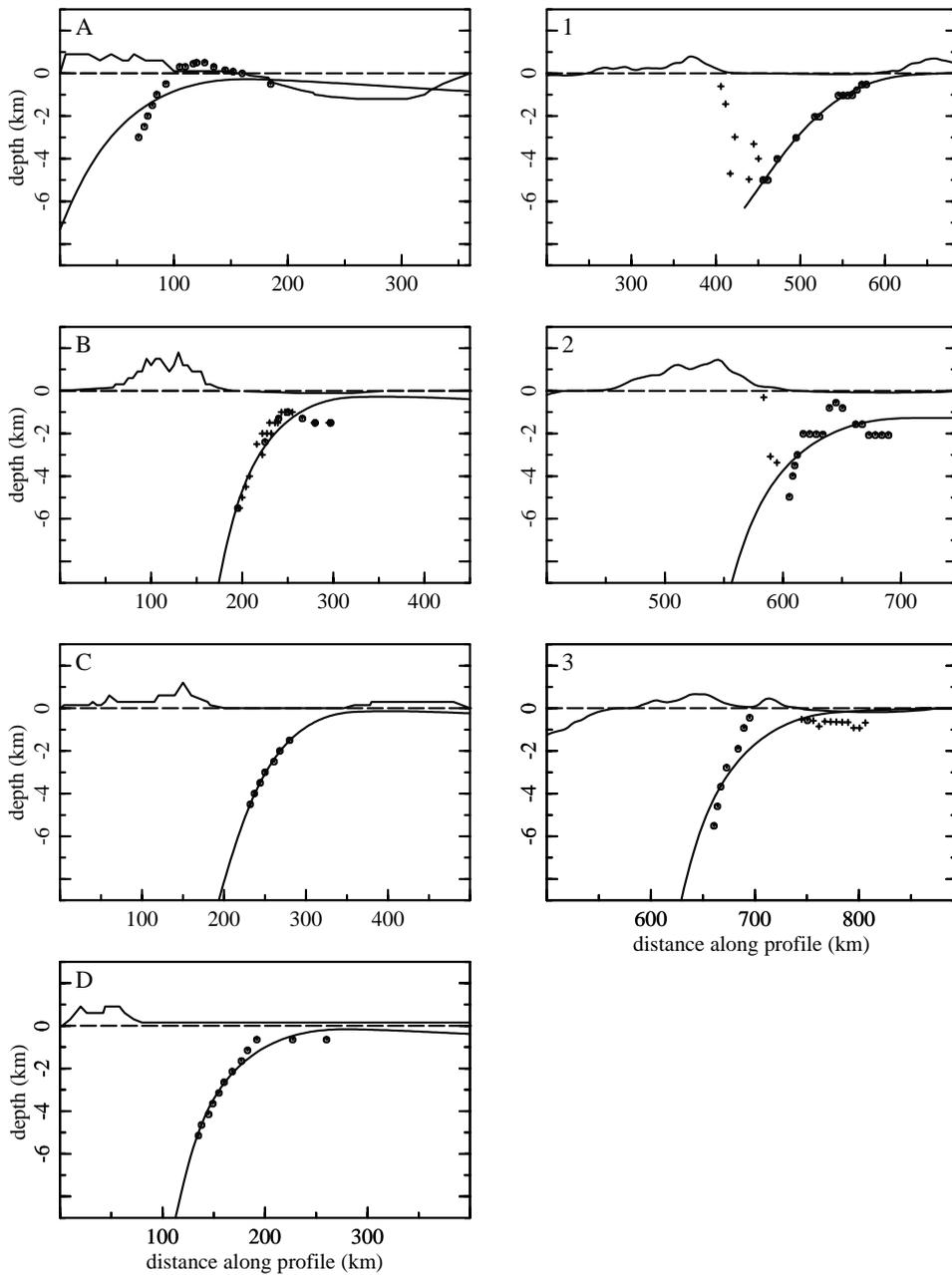
Fig. 10. Solutions with a continental rheology for each of the sections shown in Fig. 1. Symbols as in Fig. 7.

tained by intraplate compressive stresses. For purely elastic plates reasonable intraplate stresses [order of magnitude tens of MPa (Bott, 1991, 1993; Wortel et al., 1991)] can lead to higher sharp bulges if the plate thickness and, therefore, the flexural wavelength is small. The flexural wavelength of models with a con-

tinental rheology is too large to lead to such small bulges. Also, intraplate compression only needs to be invoked for the sections through mid Italy, it is not clear what could be the mechanism that causes compression only in this region. Alternatively, a local weak zone in the foreland may localize and uplift

Table 4
Results for modelling with a continental plate (V and M measured at $d_{MV} = 60$ km, see Fig. 2)

Section	Baseline (m)	V (10^{11} N m^{-1})	M (10^{16} N)	Surface load (10^{13} N m^{-1})	rms-difference (m)
A	800	2.67	21.3	−3.52	687
B	800	5.86	21.7	−3.42	533
C	600	5.52	20.4	−1.93	53
D	800	6.51	21.9	−2.01	277
1	200	3.78	12.7	−0.75	131
2	1800	5.84	21.8	−4.11	788
3	200	8.16	5.34	−2.76	891

the flexural bulge (Waschbusch and Royden, 1992). For example, for section 2 a reduction in elastic plate thickness by a factor of 2 over a few tens of kilometers gives an acceptable fit to the high in the data. The same holds for section B (fig. 9 of Waschbusch and Royden, 1992). For a plate with a continental rheology a local weak zone can also uplift the flexural bulge, but the plate remains too stiff to form such an high and small bulge as suggested by the data for sections 2, 3 and B. Therefore, we do not prefer introducing the undetermined feature of local weak zones in the models, but instead prefer a third explanation of the sharp highs in the data. This explanation is related to the implicit assumption in the modelling that the base of Pliocene level is only deformed by flexure and that the present shape of this level can thus be represented by a flexural model. However, the Plio-Pleistocene Adriatic foreland basin is affected by thrust deformation (Ricci Lucchi, 1986; Ori et al., 1986; Bally et al., 1986). It is, therefore, very likely that in some regions the base of Pliocene level has been distorted by thrusting. The structure that has been implicitly interpreted as a flexural bulge on sections 2 and B is probably caused by a major fault structure (like the Struttura Costiera or the Contessa or Colosseo faults (Ori et al., 1986, 1991)), and is not, or not completely, a flexural bulge. This would explain the difficulties that were met in modelling these sections. Also in the case of section 3 it is likely that the high is not only caused by flexure related to the Adriatic subduction zone. At the present location of the high an high structure has existed since at least the Oligocene (Boccaletti et al., 1990b). This cannot have been a flexural bulge of the Adriatic subduction zone at that time, since

the outermost thrust front (and hence the trench) was then still located further to the west.

In this study only the westward directed subduction of the Adriatic plate beneath Italy has been considered; the Adriatic plate, however, also subducts eastward beneath the Dinarides and Hellenides (Fig. 1). These opposite vergent subduction systems can potentially influence one another (Moretti and Royden, 1988). This may play a role in the southern Adriatic Sea, where the distance between the Apennine/Calabrian and the Hellenides subduction zones is decreasing southward (to approximately 300 km for section A). The deflection data for the different sections do, however, not show indications for influence of the eastern subduction system, not even for the southernmost section (A). Moretti and Royden (1988) modelled three sections through South Italy with an elastic plate subject to double sided subduction. Their fit for section A (their section 1) was obtained for a slightly thinner plate than in Royden (1988).

5.2. Comparison with Royden (1988)

For sections A through D data from Royden (1988) were used and, therefore, a comparison of results from the modelling with an elastic rheology to the results of Royden should, in principle, be possible. This is, however, not the case for sections C and D, since it was shown in this study that for these profiles a trade-off relation between elastic plate thickness and the end loads exists, which cannot be eliminated by the inclusion of gravity data. For both sections A and B the fit obtained in this study is for a slightly thinner plate than in Royden (1988). A

comparison of the values obtained for the end loads is therefore not very meaningful, even if these had been measured at the same position along the flexing plate in both studies. The visual fit for these two sections is quite similar.

5.3. Lateral migration of slab detachment

Remarkably, the three northernmost sections (D, 1 and C) can be modelled quite well, both with an elastic and a continental rheology, while the sections to the south (2, B, 3 and A) pose more difficulties. This may be caused by differences in the time at which slab detachment occurred in relation to the time at which the horizon that at present forms the base of the foredeep was deposited. Independent evidence in support of lateral migration of slab detachment below the Italian region, is the fact that the predicted lateral component in the migration of the deepest part of the foreland basin, the depocenter, has been found from stratigraphic analyses (Van der Meulen et al., 1998). From Late Miocene to Pleistocene the depocenter shifted from north to south Italy. If this depocenter shift is accepted as the surface effect of a migrating slab detachment, it follows that for sections in the north the base of the Plio-Pleistocene foredeep was deposited after slab detachment (and rebound), while for sections in the south this same horizon was deposited before detachment. Consequently, the base of Pliocene level on these southern sections experienced the effects of slab detachment: increased subsidence followed by rebound. These sections are most difficult to explain with the purely flexural model used in this study. The base of Pliocene level appears more distorted, an indication that these regions experienced more, post-base Pliocene, structural activity. This might be related to effects of slab detachment.

If it is accepted that slab detachment had already occurred in the north before the horizon that at present forms the base of the Plio-Pleistocene foreland basin was deposited, the question arises why this horizon at present lies as deep as 6–8 km. An explanation in the form of an increase in slab pull with time is in contradiction with the postulated timing of the slab detachment. A different explanation is that the outward migration of the trench system since the early Pliocene has caused the observed deep location of the horizon. We found that for an outward migration of approximately 60 km (Boccaletti et al., 1990b), an initially horizontal sediment level can be forced to the observed depths. The eastward migration of the Apennine thrust fronts and the foreland basin increases from north to south (Bally et al., 1986; Boccaletti et al., 1990a). However, since the base of the Plio-Pleistocene foredeep is slightly younger towards the south (mid E-Pliocene for sections 3 and A), the amount of migration that the base of the foredeep experienced is approximately equal from north to south. It is uncertain what drives the observed eastward migration if the subducted plate in north Italy is detached and the remaining short plate is continental. It is possible that the roll-back in the north is related to the roll-back in south Italy, as suggested by Wortel and Spakman (1992) and indicated by numerical modelling by Negredo et al. (1997).

6. Conclusions

On the basis of the hypothesis of lateral migration of slab detachment (Wortel and Spakman, 1992) variations in loads are expected along strike of the Apennine mountains. From the results of modelling with an elastic plate no conclusions related to the pattern of the loads along strike of the Apennines can, however, be drawn. This is due to two different causes. First, for three sections (1, C and D) a trade-off relation between elastic thickness and the deep load exists. The inclusion of gravity data in the modelling process does not make an effective discrimination between the different models possible. This contrasts with the findings of Royden (1988). She, however, employed an hypothetical slab end, beyond which slab material does not contribute to gravitational anomalies. We do not consider this approach in agreement with the presence of several hundreds kilometers of subducted lithosphere (Spakman, 1990; Spakman et al., 1993; Amato et al., 1993). Second, for the sections with a sharp high in the deflection data (2, 3 and B), a more or less reasonable fit could only be obtained for very thin plates. A small elastic plate thickness is, however, not reasonable for the continental Adriatic lithosphere.

The results of modelling with a continental rheology also did not allow a comparison of the values of loads between the sections, due to the large difference in the quality of the fit to the data between the different sections. A good fit was obtained for those sections (1, C and D) for which a good fit could also be obtained for different elastic thicknesses in the modelling with an elastic plate. It especially proved difficult to obtain a reasonable fit to the deflection data for the profiles with a sharp high (2, 3 and B), due to the stiff flexural behaviour of the continental plate model.

The sharp highs in the data for sections 2 and B, which posed difficulties in the modelling with both an elastic and a continental rheology, are most probably not (only) caused by flexure, but likely are a consequence of thrust activity, which is not included in the flexural models. The base of Pliocene level appears more distorted on the southern, than on the northern sections.

Along strike variations in vertical loads could not be resolved from the Plio-Pleistocene foreland basin of the Apennines by modelling deflection and gravity data in two-dimensional sections. We conclude, therefore, that the hypothesis of lateral migration of slab detachment can neither be shown to be consistent nor inconsistent with the observations. Evidence in support of the hypothesis can be found in the lateral component in the depocenter shifts (Van der Meulen et al., 1998). It is likely that the base of Pliocene level in the south experienced the effects of slab detachment. In the north this level was probably deposited after slab detachment occurred, the present (deep) flexure is in this case for the greatest part caused by the outward migration of the trench system.

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