Crustal structure and dynamics in the Rhine Graben and the Alpine foreland

Marc-André Gutscher

Geophysikalisches Institut der Universität Karlsruhe, Hertzstrasse 16, D-76187 Karlsruhe, Germany

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SUMMARY

The Rhine Graben rift stretches for 300 km from Basel to Frankfurt. The axis of maximum subsidence switches from west in the southern graben, to east in the northern graben. Subsidence in the south began in the upper Eocene and was interrupted in the mid-Miocene by a broad uplift of 1–1.5 km in the Vosges–Schwarzwald (Black Forest) region, whereas subsidence in the north proceeded continuously from the lower Oligocene to the present.

A combined geophysical and geological interpretation of the Rhine Graben is presented in the form of a 3-D gravity model and 2-D flexural plate modelling from the Alps across the Alpine foreland to the Rhenish Massif. Modelling is based on crustal structure revealed by geologic, gravity and seismic data, including ECORS, DEKORP and EGT profiles.

The 3-D gravity modelling supports a modest density contrast of 0.3 g cm\(^{-3}\) at the crust–mantle boundary. Since there is no evidence of a large-scale upper-mantle anomaly, a dense lower crust (2.9–3.0 g cm\(^{-3}\)) is proposed as the cause of this modest density contrast. The presence of a dense, mafic, lower crust in south-west Germany is corroborated by lower-crustal xenoliths, elevated P-wave velocities (6.7–7.2 km s\(^{-1}\)) and high Poisson's ratios (0.26–0.29).

Residual gravity and magnetic data reveal strong crustal heterogeneities, particularly from the dense, magnetic Saxothuringian Terrane to the light, non-magnetic Moldanubian Terrane. The polarity of the asymmetric Rhine Graben reverses at this Variscan tectonic boundary. Light graben fill can account for nearly the entire 20–40 mGal, short-wavelength gravity minimum, but the presence of intrusive bodies ranging from granitic to mafic composition can also strongly influence the local gravity field.

2-D modelling of the flexure of the European lithosphere along seven 1000 km long profiles indicates a flexural bulge of 1–1.5 km in the Alpine foreland of Southern Germany, which corresponds to an ENE-trending Moho uplift of the same magnitude and to a band of free-air gravity highs. Longer flexural wavelengths in the east indicate increasing lithospheric rigidity and increasing elastic thickness towards the Bohemian Massif. The Vosges–Schwarzwald dome is interpreted as the intersection of the Oligocene Rhine Graben rift and associated rift flank uplift with the mid-Miocene flexural bulge.

Key words: Alps, crustal structure, dynamics, Germany, gravity, lithosphere, uplift.

INTRODUCTION

The Rhine Graben, one of the best-studied continental rifts in the world (Illies & Fuchs 1974; Fuchs et al. 1987; Prodehl et al. 1992), belongs to the Central European Rift system located in a broad belt across the Alpine foreland. The Rhine Graben is 30–40 km wide and stretches 300 km from Basel to Frankfurt (Fig. 1). Graben subsidence began in the south in the upper Eocene, with maximum sediment thicknesses (up to 2 km) in the west of the graben (Doebel & Olbrecht 1974). Deposition was interrupted in the
mid-Miocene by a broad uplift of c. 1.5 km affecting both rift floor and flanks. To the north, subsidence proceeded uninterrupted from the Oligocene to the present with a maximum accumulation of 3 km of fill in the eastern graben (Villemin, Alvarez & Angelier 1986).

Seismic reflection experiments across the Rhine Graben (Brun et al. 1991; Wenzel et al. 1991), in adjacent areas such as the Schwarzwald (Black Forest) (Lüschen et al. 1987), as well as through the Alps (Valasek et al. 1991), offer a detailed picture of the deep crustal structure of this region. Refraction seismic investigations, for example from the European Geotraverse (EGT) (Prodehl & Giese 1990; Aichroth, Prodehl & Thybo 1992) have provided independent data on crustal thickness and more detailed information about the P- and S-wave velocity distributions in the lithosphere.

The primary goal of this study is to combine this detailed information on crustal structure with newly compiled gravity data on north-east France and south-west Germany (Plaumann 1987; Rousset & Bayer 1990; Grosse & Conrad 1990) (see Fig. 2) in order to construct a consistent 3-D model of the crust and upper mantle in the study area. Remaining anomalies are then to be interpreted in their geological context. Such integrative studies (Plaumann, Groschopf & Schädel 1986; Setto & Meissner 1987; Müller 1988) provide better constrained crustal models than any single method.

Geodynamic models of the Rhine Graben in the past have often included a mantle diapir below the southern portion of the rift (Illies 1972; Kahle & Werner 1980; Villemin et al. 1986). One of the most commonly cited pieces of evidence for the diapir model is the 1.5 km Miocene uplift.

A further goal of this study is to investigate an alternative mechanism for the broad Miocene uplift through flexural plate modelling of the European lithosphere in the Alpine foreland of southern Germany. The causal relationship between lithospheric flexure, orogenic belts and foreland basins with accompanying flexural bulges has already been demonstrated in combined seismic and gravity studies in the Western Alps (Guellec et al. 1990; Bergerat et al. 1990) and the Pyrenees (Desegaulx, Roure & Villein 1990).

3-D GRAVITY MODELLING

A 3-D gravity model for the northern Alps, Rhine Graben and southern Germany was developed using an interactive 3-D program: IGAS (Götze & Lahmeyer 1988). The primary objective of the gravity modelling was to identify the contribution to the Bouguer gravity from (1) light Tertiary sediments and (2) variations in Moho depth. Since these are the interfaces with the greatest density contrasts (0.3–0.5 g cm$^{-3}$) and shallower depths in the lithosphere, they will influence the gravity field the most strongly. If these effects can be adequately modelled, the remaining anomaly will be a combination of very long-wavelength mantle, and short-wavelength crustal, heterogeneities. A necessary first step is ascertaining the appropriate density contrast to use at the crust–mantle boundary.

Density contrast at the crust–mantle boundary

Gravity studies performed by previous investigators have included density contrasts as high as 0.5 g cm$^{-3}$ (Kahle & Werner 1980). In the southern Rhine Graben area, the crust is only 25 km thick, compared to the regional average of 28–30 km; one would therefore expect a long-wavelength gravity maximum of the order of 60–100 mGal using such a large density contrast. Since this is not observed (the so-called 'gravity paradox'), it was postulated that a low-density mantle body could be responsible for negating the gravity effect of the uplifted Moho boundary (Kahle & Werner 1980; Rousset & Bayer 1990; Rousset et al. 1993).

However, there is evidence from seismic velocity studies which suggests that the lower crust is composed of dense, mafic material and that the ensuing crust–mantle density contrast is much lower than previously assumed. P-wave velocities obtained from seismic experiments for the lower crust below 21 km range from 6.7 to 7.2 km s$^{-1}$ in the Rhine Graben (Edel et al. 1975; Zucca 1984) and in the adjacent Schwarzwald (Lüschen et al. 1987) (see Figs 3a–c). Such velocities correspond to densities of 2.9–3.0 g cm$^{-3}$ based on velocity–density relationships (Kahle & Werner 1980), taking the appropriate pressure into account (see Fig. 3d).

Combined P- and S-wave studies (Holbrook et al. 1988) have revealed widespread regions of elevated Poisson's ratio (from 0.26 to 0.29) in the lower crust in south-west Germany. Petrologically, these are interpreted to correspond to rocks of gabbroic to dioritic composition which have densities in the 2.9–3.0 g cm$^{-3}$ range (Plaumann 1987). Such a dense lower crust would imply a modest density contrast, of the order of 0.3 g cm$^{-3}$, at the crust–mantle boundary.

Furthermore, xenolith studies from Cenozoic volcanic provinces in the study area (Wedepohl 1986; Sachs 1988; Mengel et al. 1991) document the presence of high-density components in the lower crust. Near the southern graben, xenoliths from the lower crust (23–28 km depth) consist of ultramafic meta-pyroxenites and hornblendites (Urach) and mafic spinel–pyroxene granulites (Hegau) with minor occurrences of salic metasedimentary granulites, and exhibit P-wave velocities in the 6.5–7.8 km s$^{-1}$ range (Mengel et al. 1991; Glahn, Sachs & Achauer 1992). Xenoliths from the north Hessen Depression and Eifel volcanic fields, near the northern end of the graben, reveal a lower crustal layer (27–33 km depth) of mafic granulites and meta-pyroxenites to meta-hornblendites, with P-wave velocities of 6.8–7.5 km s$^{-1}$ (Wedepohl 1986; Mengel et al. 1991).

Model geometry and densities

A simple 3-D model was constructed with the following five litho-tectonic units:

1. Rhine Graben sediments,
2. Molasse Basin sediments,
3. crystalline basement of the upper crust,
4. a dense lower crust and
5. the upper mantle (see Fig. 4).

Tertiary sedimentary thicknesses in the Rhine Graben were taken from a previous compilation (Doebi & Olbrecht 1974), and for the Molasse Basin from the European Atlas of Geothermal Resources (Haenel & Staroste 1988).

A lateral density contrast of $-0.40$ g cm$^{-3}$ between the graben fill and adjacent basement provided the best fit to the short-wavelength, 20–40 mGal dropoff at the rift flank and is in line with previous 2-D studies (Plaumann et al. 1991).
Figure 2. Bouguer gravity map of study area (after Plaumann 1987; Rousset & Bayer 1990) (5 mGal contours). Location of Figs 7 (a)–(d) and profiles from Fig. 9 are marked.
Figure 3. (a)--(c) P-wave velocity distribution in the Black Forest–Rhine Graben area (Edel et al. 1975; Zucca 1984; Lütschen et al. 1987). (d) Density–velocity relationship (Kahle & Werner 1980).
A lateral density contrast of $-0.30 \text{ g cm}^{-3}$ was used for the Molasse basin. The average density of the crystalline upper crust is taken to be $2.67 \text{ g cm}^{-3}$, which is the reduction density used in creating the Bouguer gravity maps (Plaumann 1987; Roussel & Bayer 1990) and consistent with basement samples from the study area (Plaumann et al. 1986; Edel & Fluck 1989). A density contrast of c. $0.2-0.3 \text{ g cm}^{-3}$ is presumed to exist between the upper crust and lower crust. This boundary may represent a gradual transition, and, since scarcely any information is available on its geometry and composition, it was assigned a constant depth and thus provides no lateral density contrast which could affect the gravity field.

The Moho topography used is shown in Fig. 5 and is a modified form of pre-existing maps (Zeiss, Gasewski & Prodehl 1990; Prodehl et al. 1992), with the inclusion of near-vertical seismic reflection data from the ECORS-DEKORP Rhine Graben profiles (Brun et al. 1991; Wenzel et al. 1991), the DEKORP 1-C line (Meissner & Bortfeld 1990), the DEKORP 2-S line (DEKORP Research Group 1985), the Schwarzwald profiles (Lüschen et al. 1987) and the Urach profiles (Bartelsen et al. 1982). Moho depths in northern Switzerland are based on data from the Swiss portion of the EGT (Schwendener & Mueller 1990). The model dimensions are $300 \times 360 \text{ km}$ (from $47.5^\circ$ to $50.2^\circ$ N, and from $6.4^\circ$ to $11.2^\circ$ E) (see Fig. 2), with an additional $50 \text{ km}$ to the north, where crustal thickness varies little, and $100 \text{ km}$ on all other sides to eliminate edge effects. The observed Bouguer gravity was gridded at a total of 525 stations, with a $15 \text{ km}$ spacing. The first station along the x-axis is at $15 \text{ km}$. The calculated Bouguer gravity (based on the 3-D model geometry and densities) is then gridded at

Figure 4. Geometry of the 3-D gravity model displayed along four E-W profiles from north (a) to south (d).

Figure 5. Moho depth map of study area (after Zeiss et al. 1990) with the inclusion of seismic refraction data from the EGT (+) and other regional studies (×) and the addition of seismic reflection data from the ECORS-DEKORP Rhine Graben study, DEKORP 2-S, and the Black Forest and Urach studies (○).
Crustal structure in the Rhine Graben

Results of gravity modelling

The gravity effect from the 3-D model was calculated with density contrasts of 0.3, 0.4 and 0.5 g cm\(^{-3}\) at the crust–mantle boundary. The observed and calculated gravity along two 360 km E-W profiles, just north of Basel and through the Odenwald (N–S 20 and 245 km respectively on Figs 7 and 8), are shown in Fig. 6. The two profiles are chosen because they represent two tectonically different regions of the Rhine Graben and surrounding Mesozoic platform. The north has a maximum 3 km sediment thickness and a west-dipping master fault in a region unaffected by Miocene uplift and where the gravity field is not influenced by a strong gradient in crustal thickness, and the south has a 1–1.5 km deep basin of opposite polarity, in a region strongly affected by Miocene uplift, and further east the same profile crosses the Molasse Basin and Alpine Front where crustal thickness increases laterally to the SE.

In the southern profile, the local 20–30 mGal misfit below the Molasse Basin is probably the result of too low a density assigned to the deepest sediments in the basin (2–5 km depth). This minor effect aside, the 0.5 g cm\(^{-3}\) contrast at the crust–mantle boundary produces a misfit of 60 mGal, the 0.4 g cm\(^{-3}\) contrast a misfit of 30 mGal and the 0.3 g cm\(^{-3}\) contrast parallels the observed data nicely. When the effect of the ‘light Moldanubian crust’ (see next section) is taken into account, the fit to the observed data is excellent (Fig. 6b).

The northern E-W profile through the Odenwald and northern Rhine Graben also shows the effect of too high a density contrast at the Moho (Fig. 6c). The thicker crust west of the rift produces too low a gravity signal (–35 to –25 mGal versus the –15 to –5 mGal observed) for high-density contrasts. Variscan granites (see next section) produce the local minimum east of the eastern rift flank.

The calculated Bouguer gravity for the entire study area is shown in Figs 7(b)–(d) for density contrasts of 0.3, 0.4 and 0.5 g cm\(^{-3}\), which produce calculated gravity effects ranging from –116 to +18, –136 to +23 and –156 to +28 mGal, respectively. The observed Bouguer gravity (Fig. 7a and Fig. 2) ranges from –120 to +20 mGal. A value of 0 mGal is chosen as the common reference level, which means that large misfits are located in the Alpine region where there is a rapid increase in crustal thickness to the south-east. Conversely, if –100 mGal is chosen as the reference level, then the misfit is shifted to the southern Rhine Graben in the form of +60 mGal maximum.

Additionally, a 3-D density inversion was performed using the interactive 3-D IGAS gravity program. The previously described geometry was held fixed and the density contrasts between all bodies were set free. The inversion yielded an optimum fit for a density contrast of 0.28 g cm\(^{-3}\) at the crust–mantle boundary, 0.43 g cm\(^{-3}\) between the crystalline basement and graben fill, and 0.27 g cm\(^{-3}\) between the basement and Molasse Basin fill. A similar 3-D inversion of crustal and mantle densities in the Rhenish Massif–Rhine Graben area (Drisler & Jacoby 1983) yielded a density contrast of 0.25 g cm\(^{-3}\) at the Moho.

One can thus conclude that a moderate density contrast of 0.3 g cm\(^{-3}\) provides the best fit to the observed gravity, utilizing the known Moho topography. Additionally, EGT gravity studies crossing the Alpine orogenic belt, where crustal thickness locally exceeds 60 km (Cassinis, Cassano & Capelli 1990; Holliger & Klempener 1991), have also demonstrated excellent agreement between observed and calculated gravity for a modest density contrast of 0.3 g cm\(^{-3}\).
Figure 7. (a) Observed Bouger gravity, with Rhine Graben (light shading) and Molasse Basin (dark shading) indicated; (b)–(d) gravity maps calculated with density contrasts of 0.3 g cm$^{-3}$, 0.4 g cm$^{-3}$, and 0.5 g cm$^{-3}$ at the Moho (contour interval 10 mGal, borders in km).
Crustal structure in the Rhine Graben

Figure 7. (Continued.)
Lastly, a global gravitational potential study, utilizing spherical harmonics up to degree and order 30 (Martinec 1994) yielded a density contrast at the Moho of 0.28 g cm$^{-3}$ under continental areas.

**Alternative models**

The selection of a higher density contrast at the crust–mantle boundary (Kahle & Werner 1980) requires the addition of bodies at mantle depth to resolve the apparent 'gravity paradox'. The presence of an asthenospheric diapir (mantle plume) beneath the southern Rhine Graben has been proposed from surface-wave studies (Panza, Mueller & Calganile 1980) and crustal tomography (Koch 1993). Alpine parallel asthenospheric upwelling has also been suggested as a means of compensating the positive gravity contribution of the uplifted Moho and explaining the topographic uplift in the Alpine foreland (Lyon-Caen & Molnar 1989; Rousset et al. 1993).

However, neither the results of a small-scale teleseismic investigation over a $100 \times 200$ km area in the southern Rhine Graben area (Glahn et al. 1993) nor those of large-scale tomographic studies (Babuska, Plomerova & Sileny 1984; Spakman 1988; Babuska, Plomerova & Grant 1990) confirm a broad region of reduced seismic velocities in the mantle, to depths as great as 150 km. Finally, the counter argument, that a modest density contrast exists due to a dense lower crust, is supported by the crustal seismic velocity studies and petrologic investigations discussed above.

**Crustal Structure and Composition and the Moldanubian–Saxothuringian Suture**

The difference between the observed and calculated gravity (Fig. 8) reveals the effect of lateral variations in crustal composition plus a long-wavelength contribution from possible mantle and lower crustal heterogeneities. A very marked trend is the 10–20 mGal trough, trending NE from Heidelberg, attributable to a series of Variscan granitic plutons (as known from the Heidelberg granite). The extension of this trend across the Rhine Graben is indicated by the bull’s-eye residual gravity low, 50 km north of Strasbourg. Granitic basement is present in this locality, just north-east of the Vosges, known from the Soultz borehole (Jung 1990). Local gravity maxima, such as the Saverne–Saarbourg maximum (+20 mGal), Kraichgau anomaly (+10 mGal) or the Odenwald maximum (+20 mGal), are common in the Saxothuringian Terrane. The latter correlates with surface outcrops of schist, diorite, amphibolite and gabbro (in the Frankenstein Massif, east of Darmstadt) (Gutscher 1991a) and these rocks are possible candidates for the origin of the other local maxima (Plaumann et al. 1986; Edel & Fluck 1989; Rousset et al. 1993).

Short-signal wavelengths (with ‘half-widths’ of 5–11 km) and strong horizontal gravity gradients (0.5–1.0 mGal km$^{-1}$) indicate that the depth to the centre of these bodies is not greater than 5–11 km and that they are thus located within the upper crust (see Fig. 9). This is derived from the basic

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**Figure 8.** Difference between observed and calculated Bouguer gravity (in mGal), with Saxothuringian–Moldanubian terrane boundary (dashed line), Variscan granite trend (dotted line), Rhine Graben (light shading) and Molasse Basin (dark shading) indicated (borders in km).
Figure 9. 2-D gravity profiles, with ‘half-width’ (maximum depth to centre) indicated (for location, see Fig. 2).

Thus, the potential field data paint a picture of two sharply contrasting crustal types: the heavy and magnetic Saxothuringian domain (Heinrichs 1986) versus the light and non-magnetic Moldanubian (Edel & Fluck 1989). The Saxothuringian Terrane can thus be visualized as a ‘high plateau’ in the Bouguer gravity field, cut by the NE-trending Variscan granites and cut more deeply by the NNE-trending light sedimentary fill of the Rhine Graben.

Upper crustal seismic velocities in the Saxothuringian Terrane are slightly higher than in the Moldanubian Terrane (6.1–6.2 km s⁻¹ versus 5.9–6.0 km s⁻¹ at 5 km depth) (Aichroth et al. 1992). Whereas the Saxothuringian Terrane is marked by a 10 km thick, high-velocity (6.6–7.0 km s⁻¹) lower crust, the Moldanubian Terrane displays a layer with a maximum thickness of 5 km and only moderately high velocities (6.4–6.8 km s⁻¹) in the lower crust (Prodehl & Giese 1990). The Variscan suture is also marked by a region of high electrical conductivity in at least three NW–SE profiles in southern Germany (ERCEUGT-Group 1990).

Rift geometry and the Variscan suture

Particularly striking is the fact that the polarity of the Rhine Graben reverses at the Saxothuringian–Moldanubian boundary; to the north the main subsidence axis and master normal fault is in the east, whereas to the south it lies in the west (Doebl & Olbrecht 1974; Brun et al. 1992) (Fig. 10). Furthermore, the results of the seismic reflection profiles
Figure 10. Topography in the study area and the base Tertiary/top Jurassic horizon (in metres below sea level) (after Doebl & Olbrecht 1974; Haenel & Staroste 1988).
DEKORP-ECORS 9-N and 9-S (Wenzel et al. 1991; Brun et al. 1991) reveal deep crustal structures interpreted as a westward-dipping shear zone in the northern graben and an eastward-dipping shear zone in the southern graben. In both cases, the shear zones cross the upper crust, affect the lower crust, and appear to offset the Moho, although a direct ‘simple shear’ continuation through the lower crust (Wernicke 1985) can neither be established nor ruled out. Therefore, this Variscan tectonic zone, known as the Lalaye-Lubine-Baden-Baden Zone, appears to have played a dominant role in the formation and geometry of the Rhine Graben (Brun et al. 1992). This late Palaeozoic suture (Behr et al. 1984) seems to have served as an accommodation or transfer zone (Bosworth 1985; Pinet & Coletta 1990) (Fig. 11). Scaled models of the crust and lithosphere using sand and silicone (Tron & Brun 1991) demonstrate that such a pre-existing discontinuity striking obliquely to the direction of extension can produce the alternating asymmetric geometry observed in the Rhine Graben.

**FLEXURE OF THE ALPINE FORELAND**

**Regional setting**

Direct observations of the bathymetry and free-air gravity of the oceanic lithosphere at subduction zones reveal a trench–‘outer rise’ pair, separated by several tens to a few hundred kilometres (Watts & Talwani 1974; Bodine, Steckler & Watts 1981). In similar fashion, loading of the continental lithosphere produces flexural basins, for example the Alberta Basin in the foreland of the Canadian Rockies (Beaumont 1981) or the Appalachian Basin, whose continued existence c. 300 Ma after loading suggests that lithospheric flexure is a long-lived phenomenon (Karner & Watts 1983). Flexure studies in the Appenine–Adriatic area (Moretti & Royden 1988) and in the Himalaya and Alps (Karner & Watts 1983) indicate that lithospheric flexure is caused primarily by subsurface loads (i.e. obducted crustal blocks or dense, sinking lithospheric slabs) and is not merely caused by the emplacement of thin-skinned nappes or other surficial, topographic loads.

The proximity of the study area to the Alpine Orogenic belt raised suspicions that the Cenozoic Alpine collision may have affected the foreland of southern Germany. Thus, the geometry of the European lithosphere was analysed along seven, 1000 km long, 2-D profiles starting in the Po plain and crossing perpendicular to the Alps (Fig. 12), in order to determine the magnitude and location of possible vertical deflections due to lithospheric flexure.

The deep crustal structure of the Alps as seen in the Swiss seismic-reflection profiles (Valasek et al. 1991) clearly shows the European lithospheric plate descending below the Adriatic plate (see Fig. 13). The Molasse Basin is a classical example of a continental foreland basin or foredeep (Ziegler 1982; Mugnier & Ménard 1986). The gentle 1° to 2° SSE dip of Mesozoic strata in southern Germany (Swabian Jura) increases below the Molasse basin to 5° to 6° (Haenel & Staroste 1988; Lyon-Caen & Molnar 1989).

**Flexural Modelling**

The depth to the top of the European plate below the Alps (Fig. 14, filled diamond) was obtained from an Alpine EGT study (Roeder 1990). For the entire Alpine foreland region, the uppermost Jurassic limestone was chosen as a stratigraphic marker (Fig. 14, open diamonds) to indicate subsequent changes in elevation. For each profile, the theoretical curve describing the flexure of a semi-infinite (broken) elastic plate, under an applied end load (eq. (A1), Appendix), was fit to the deflection of the stratigraphic marker (Fig. 14) by inserting the appropriate maximum deflection $w_0$ (from 16 to 22 km) and origin $x_0$ and by setting $\alpha$, the flexural wavelength, free. After several iterations, the least-squares fit function converged and yielded the best-fit flexural wavelength (see Table 1).

**Discussion**

As can be seen by comparing the sections (Fig. 14), the flexural wavelength (distance from the plate edge to the bulge) decreases from east to west while the height of the bulge increases from east to west. The flexural rigidities (calculated from eq. (A3), Appendix), $D = 10^{22}$–$10^{24}$ N m and effective elastic thicknesses (calculated from eq. (A4), Appendix), $T_e = 25–50$ km (Table 1) are generally consistent with previous studies in the area (Karner & Watts 1983; Lyon-Caen & Molnar 1989) and with others from the Pyrenean (Desegaulx et al. 1990) and Appenine forelands (Moretti & Royden 1988).

Equation (A5) (Appendix) permits a calculation of the applied end load $F$. The total weight per unit length, due to the overlying Adriatic crustal block, applied to the end of the European plate was found to be $2–5 \times 10^{13}$ N m$^{-1}$, while the buoyancy force is $3–5 \times 10^{12}$ N m$^{-1}$. For comparison, the topographic load from the Alps is only of the order of $3–4 \times 10^{12}$ N m$^{-1}$, roughly equivalent to the buoyancy force, providing an indication that the Alps are largely isostatically compensated, as suggested by previous authors (Lyon-Caen & Molnar 1989). A simple computer program based on eq. (A2) (Appendix) was used for calculating the flexure of a distributed load $F(x)$ (lithospheric wedge and topography) on a broken elastic plate. The previously derived flexural rigidities were employed, together with the known geometry of the Adriatic block. The flexural wavelengths and flexural
Figure 12. Location map of flexure profiles, with depth to top of European plate below the Alps, crystalline Massifs shown by '+', and the location of the crest of the flexural bulge shown by a dashed line.

Figure 13. Schematic lithospheric cross-section of the Alpine collision, based on EGT studies (after Valasek et al. 1991), and showing the location of the foreland flexural bulge.
bulge heights calculated by the program were the same as those obtained from the graphical least-squares-fit approach.

The regional decrease in flexural wavelength from east to west corresponds to a decrease in the flexural rigidity (and effective elastic thickness) of the European plate from east to west (Table 1), as already noted in previous investigations (Karner & Watts 1983; Lyon-Caen & Molnar 1989). The decrease in rigidity correlates with a general westward increase in heat flow from the central Bohemian Massif (60 mW m\(^{-2}\)) to the Swabian Jura, Schwarzwald–Vosges area (80–90 mW m\(^{-2}\)) (Cermak & Rybach 1979). A local increase over 100 mW m\(^{-2}\) in the Rhine Graben is most likely due to near-surface circulation effects.

The Vosges–Schwarzwald dome and the flexural bulge versus the mantle plume model

The modelled flexural bulge can account for a maximum uplift of c. 1500 m in the southern Schwarzwald (Black Forest), diminishing to c. 1000 m in the Vosges and northern
of Moho uplift (as a passive marker in the flexed lithosphere) is also consistent with the bulge height calculated. Therefore, the Vosges–Schwarzwald dome can be interpreted as the intersection of the Oligocene Rhine Graben rift and associated rift flank uplift with the later, superimposed, Miocene flexural bulge.

### Free-air gravity and uncompensated topography

The Alpine foreland of southern Germany is marked by a broad region with mean elevations of 1000–700 m trending ENE from the Black Forest to the Swabian Jura (see Fig. 10); in other words, 600–300 m above the regional base of c. 400 m in the southernmost Rhine Graben, western Molasse Basin and the northern Mesozoic platform.

This modest plateau is expressed as a belt of free-air gravity highs (50–75 mGal) extending from the south-west corner of the study area ENE across the Swabian Jura, locally reaching 75–100 mGal in the regions of greatest topography in the southern Vosges and Schwarzwald (see Fig. 15). The regional base in the Rhine Graben and Molasse basin is 0 to −25 mGal, while to the north and west, on the Mesozoic platform, it ranges from 10 to 40 mGal.

This belt of c. +50 mGal free-air gravity highs represents isostatically uncompensated topography and crustal masses. The crustal thicknesses in this flexural bulge region are known from extensive seismic profiling (see Fig. 5) to be slightly thinner than in surrounding areas, i.e. instead of a crustal root there is an ‘anti-root’. This provides incontrovertible proof that the excess masses are not in local isostatic equilibrium, but rather are supported by so-called ‘regional isostasy’, i.e. the flexural strength of the lithosphere.

This free-air gravity signal also argues strongly against a mantle diapir model, which predicts a circular region of isostatically compensated high topography, (i.e. free-air gravity = c. 0 mGal). This is in conflict with the observed 50–90 mGal free-air gravity maxima in the Vosges–Schwarzwald area indicating uncompensated uplift.

### Objections to the flexural model and Alpine parallel upwelling model

Some flexural plate modellers questioned the applicability of a flexed elastic plate model to the Alps and claimed that ‘an absurd set of parameters’ and a ‘contrived physical explanation’ were required in order to match the current geometry (Lyon-Caen & Molnar 1989), and that the current state of near-isostatic equilibrium in the Alps ‘argues against the application of a substantial end load’.

However, these authors only considered the topographic load of the Alps ‘above sea level’ (roughly 2 km) and neglected the load of the 20 km thick overlying Adriatic block in their calculations (Lyon-Caen & Molnar 1989). It is true that the Alps themselves are roughly in isostatic equilibrium, with the buoyancy of the crustal root roughly balanced by the excess topography as discussed above, but the Adriatic block resting on the flexed European plate exerts a net downward force ten times greater than the topographic load. Additionally, regions in the northern Alps and Alpine foreland are alternatively below (c. −25 to

### Table 1. Flexural modelling parameters.

| Profile | Location        | α (km) | D (10^23 Nm) | T_e (km) | Height of bulge (m) | w_H (km) |
|---------|-----------------|--------|--------------|--------|---------------------|--------|
| 1       | Udine-Potsdam   | 109    | 10.6         | 53     | 1050                | 16     |
| 2       | Venice-Celle     | 107    | 9.8          | 52     | 1200                | 18     |
| 3       | Venice-Osnabrück | 106    | 9.6          | 51     | 1200                | 18     |
| 4       | Padua-Philadelphia | 94    | 5.8          | 43     | 1350                | 20     |
| 5       | Verona-Bergamo  | 79     | 2.9          | 34     | 1400                | 21     |
| 6       | Bergamo-Brussels| 66     | 1.4          | 27     | 1470                | 22     |
| 7       | Milan-Normandy  | 59     | 0.9          | 23     | 1470                | 22     |
-50 mGal in the Molasse basin and at the Alpine front) or above (c. 50 to 90 mGal in the Schwarzwald–Schwäbische Jura belt) isostatic equilibrium (see Fig. 15), in classical agreement with the free-air gravity signal from a ‘trench–outer rise pair’. Finally, the authors used an elevated density contrast of 0.45–0.5 g cm\(^{-3}\) at the crust–mantle boundary to calculate the expected Bouguer gravity for a given flexed geometry. This density contrast is in conflict with the recent findings from the EGT program (Casinis et al. 1990; Holliger & Klempener 1991) and with the results discussed in the above sections.

An alternative model has been proposed to explain the ENE-trending uplift in southern Germany as a zone of upwelling of hot, light, asthenospheric material in a belt across the Alpine foreland (Lyon-Caen & Molnar 1989; Rousset et al. 1993). It is argued that this ENE-elongated, upper-mantle anomaly creates a regional gravity minimum of 40–60 mGal (centred near the southern Rhine Graben) and that the associated buoyancy is responsible for the topography and uplift observed (Rousset et al. 1993).

An Alpine parallel upwelling model successfully predicts the Alpine parallel uplift and present-day topography. However, the following arguments speak against this alternative model.

(1) Selection of a higher density contrast of 0.4 g cm\(^{-3}\) leads the authors to the conclusion that a light upper-mantle body centred near the southern graben is required to resolve a moderate ‘gravity paradox’ of 60 mGal amplitude. The lower density contrast of 0.3 g cm\(^{-3}\) is shown by the same authors to produce a residual ENE-trending minimum of only 30–40 mGal (Rousset et al. 1993).

(2) 30–40 mGal of the postulated regional Bouguer gravity minimum is in fact due to lateral variation in crustal composition between the Saxothuringian and Moldanubian terranes. The same authors demonstrate the shallow nature of the Saverne–Saarebourg body (which also causes a 40 mGal signal) and discuss the substantial difference in crustal composition between the two Variscan terranes, yet ignore the substantial gravity contribution of this crustal discontinuity on a broad, regional scale (Rousset et al. 1993).

(3) Local isostatic modelling used to explain the uplift due to the postulated light upper-mantle body ignores the loading (just 200 km to the south) due to the Alpine crustal wedge (Rousset et al. 1993).

(4) An uplift caused by a hot, light mantle body in isostatic equilibrium predicts a zero signal in the free-air gravity, since the topography is isostatically compensated. This is in contradiction to the observed 50–90 mGal free-air gravity maxima in the Alpine foreland.

(5) Finally, the authors propose a body at 40–100 km depth, exhibiting a density reduction of 0.07 g cm\(^{-3}\). However, neither extensive seismic refraction profiling in southern Germany (Prodehl & Giese 1990) nor large-scale European tomographic studies (Babuska et al. 1984; Spakman 1988; Babuska et al. 1990) document broad regions
with low seismic velocities, and thus provide no evidence of an anomalously light upper mantle in the Alpine foreland of Germany.

CONCLUSIONS

(1) The observed Bouguer gravity in the Rhine Graben and Alpine foreland can be satisfactorily modelled using the known Moho topography and a modest density contrast of 0.30 g cm\(^{-3}\) at the crust–mantle boundary and does not require the addition of light bodies in the upper mantle (e.g. mantle plume).

(2) The presence of a high-density (2.9–3.0 g cm\(^{-3}\)), mafic lower crust is supported by the gravity modelling as well as by elevated P-wave velocities (6.6–7.2 km s\(^{-1}\)), high Poisson's ratio (0.26–0.29) and mafic lower-crustal xenoliths.

(3) A major crustal discontinuity exists between the Saxothuringian and Moldanubian zones, marked by an anomalously light upper mantle in the Alpine foreland of Germany.

(4) The Oligocene Rhine Graben rift reverses polarity at 3000 m elevation) to a magnetically quiet zone across southern Germany.

(5) A 1–1.5 km, ENE-trending, flexural bulge is imaged in topographic/stratigraphic data and in free-air gravity data and coincides with a 1–1.5 km, ENE-trending, Moho uplift across southern Germany.

(6) A shorter flexural wavelength and higher amplitude bulge in the Voges–Schwarzwald area indicate a reduced lithospheric 'elastic thickness' compared to regions further east (e.g. the Bohemian Massif), most likely due to a warmer, thinner lithosphere near the rift.

(7) The Voges–Schwarzwald dome appears to have formed due to the intersection of the NNE-trending Oligocene Rhine Graben and associated rift flank uplift with an ENE-trending Miocene flexural bulge.

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**APPENDIX: FLEXURE THEORY AND ANALYSIS**

The function describing the flexure of a broken elastic plate due to an applied end load is (Turcotte & Schubert 1982)

\[ w(x) = w_0 \exp \left[ \frac{-(x - x_0)}{\alpha} \right] \cos \left[ \frac{(x - x_0)}{\alpha} \right], \]  
(A1)

which is a solution of the fourth-order differential equation describing the vertical deflection \( w \) of an elastic plate overlying a fluid substratum, under an end load \( F_i \):

\[ D \left( \frac{d^4w}{dx^4} \right) + (\rho_m - \rho_f)gw = F_i. \]  
(A2)

In the case of lithospheric flexure, the flexural wavelength is related to the flexural rigidity according to (Turcotte & Schubert 1982):

\[ \alpha^4 = \frac{4D}{(\rho_m - \rho_f)g}. \]  
(A3)

Flexural rigidity is related to the effective elastic thickness by:

\[ D = \frac{ET^2}{12(1 - \nu^2)}. \]  
(A4)

The flexural wavelength \( \alpha \) and rigidity \( D \) are in turn related to the applied load \( F_i \) (Turcotte & Schubert 1982) according to:

\[ \alpha^2 = \frac{F_i}{(4DW_0)}. \]  
(A5)

**Constants and parameters used in flexural analysis**

- \( w \): vertical deflection
- \( w_0 \): maximum deflection at end of plate
- \( x \): horizontal distance
- \( x_0 \): distance from the origin to the end of the broken plate
- \( \alpha \): flexural wavelength
- \( \rho_m \): density of the elastic plate
- \( \rho_f \): density of the fluid substratum
- \( g \): acceleration due to gravity = 9.8 m s\(^{-2}\)
- \( \rho_m \): mantle density
- \( \rho_f \): crustal density
- \( E \): Young's modulus = 8 \times 10^{10} \text{ N m}^{-2}
- \( \nu \): Poisson's ratio = 0.25
- \( F_i \): applied end load
- \( D \): flexural rigidity
- \( T_e \): effective elastic thickness