Teleseismic wave front anomalies at a Continental Rift: no mantle anomaly below the central Upper Rhine Graben

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SUMMARY

The deep structure of the Upper Rhine Graben (URG), a continental rift in SW Germany and E France, is still poorly known. This deficit impedes a full understanding of the geodynamic evolution of this prominent rift. We study the lithosphere–asthenosphere structure using teleseismic waveforms obtained from the passive broad-band TIMO project across the central URG. The recovered, crust-corrected traveltime residuals relative to the iasp91 earth model are tiny (mostly less than 0.2–0.3 s). The average measured slowness (<1 s deg−1) and backazimuth (<5°) deviations are also very small and do not show any systematic wave front anomalies. These observed perturbation values are smaller than expected ones from synthetic 3-D ray tracing modelling with anomalies exceeding 2–3 per cent seismic velocity in the mantle. Thus there is no significant hint for any deep-seated anomaly such as a mantle cushion, etc. This result means that the rifting process did not leave behind a lower lithospheric signature, which could be clearly verified with high-resolution teleseismic experiments. The only significant traveltime perturbation at the central URG is located at its western side in the upper crust around a known geothermal anomaly. The upper crustal seismic anomaly with traveltime delays of 0.2–0.3 s cannot be explained with increased temperature alone. It is possibly related to a zone of highly altered granite. In the west of our network a traveltime anomaly (0.6–0.7 s delay) related with the Eifel plume is confirmed by the TIMO data set.

Key words Body waves; Continental tectonics: extensional; Dynamics of lithosphere and mantle.

1. INTRODUCTION

Extension of continental lithosphere or continental rifting are major ingredients of plate tectonics and generate impressive depressions on the Earth’s surface, for example, the East African or the European Cenozoic rift systems (Olsen 1995). Eventually rifting can lead to the development of a passive continental margin and a new ocean basin. The mechanisms that drive rupture and extension of up to 200 km thick continental lithosphere are not completely understood and appear diverse for different tectonic settings, depending on the specific stress field in a region, the far-field stresses, the history of the lithosphere and the underlying mantle processes (Olsen 1995; Ziegler & Cloetingh 2004). Two key rifting models, which describe the mechanical process of lithospheric extension, are pure shear extension versus simple shear extension (Fig. 1).

Pure shear extension (McKenzie 1978) can be described as about symmetric extension of crust and subcrustal lithosphere. This model assumes a rise of the asthenosphere into the thinned lithosphere. Depending on the amount of extension and hence asthenospheric upwelling, decompressional melting of asthenospheric material may start and lead to rift-related volcanism. Extension may be due to far-field stress or active mantle upwelling (Fig. 1c). Rise of the crust–mantle boundary (Moho) and the lithosphere–asthenosphere boundary (LAB) beneath the rift are direct consequences of pure shear extension (Fig. 1b). The simple shear model (Wernicke 1985) assumes an asymmetric extension of the lithosphere along a detachment fault or fault system (Fig. 1a). This mechanism should hardly influence the Moho and LAB depths. The distinction between active and passive rifting mainly describes the role of mantle processes and related magmatism. Active rifts (e.g. Kenya, Ethiopian rifts) are driven by hot, upwelling mantle flow or deep reaching plumes. Such rifts are characterized by extensive volcanism. At passive rifts (Cenozoic European rifts) far field stresses control the extension and volcanism plays a minor role.

The analysis and interpretation of the teleseismic wavefield are key studies to understand rifting mechanisms, because they are the most reliable methodologies to determine the subcrustal structure. Seismic images of the upper mantle structure below rifts (Davis et al. 1993) can be used to test between pure shear and simple shear models, challenge modifications to these concepts and provide
evidence for new conceptional improvements. Moho and LAB depths are revealed using receiver function modelling (Yuan et al. 2006), traveltime measurements and subsequent seismic tomography allow one to derive 3-D images of the lithosphere–asthenosphere system (Evans & Achauer 1993), and SKS-splitting analysis allows one to determine anisotropy pattern and textural interpretations (Silver 1996).

In this study we analyse teleseismic wave front anomalies at a recently acquired data set from the Upper Rhine Graben (URG), a branch of the European Cenozoic Rift System (ECRIS). Key questions in the central URG are (1) upper mantle structure and its physical properties which are important input boundary conditions for numerical geodynamic modelling, (2) the crustal structure which is a decisive ingredient for seismic hazard assessment concerning maximum potential fault dimensions, etc. and (3) the deep origin of a high heat flow anomaly which is increasingly harvested for geothermal energy. Illies (1972) and the Rhine Graben Research Group for Explosion Seismology (1974) proposed a mantle cushion which may be a hint for a pure shear rifting mechanism. This model was not confirmed by a later teleseismic tomography experiment (Glahn et al. 1993), but the mantle cushion hypothesis is sometimes still referred to and recent publications still speculate about the deep driving forces for the rift initiation including mantle plume components.

Here for the first time not only traveltime residuals are studied but also slowness and backazimuth anomalies are determined in detail at a rift structure. As we concentrate on the subcrustal lithosphere and asthenosphere, crustal traveltime anomalies are removed as far as possible based on existing seismic models. The interpretation of the wave front analysis allows us to constrain lateral structural variations of the lower lithosphere and asthenosphere as well as speculate about yet unexplained crustal traveltime residuals. Our results indicate that pure shear extension is unlikely as rifting mechanism for the URG, instead a simple shear detachment seems to be more likely to explain the subcrustal results.

2. SETTING AND TIMO EXPERIMENT

The URG is a branch of the ECRIS north of the Alpine collision zone (Dézes et al. 2004). The URG stretches about 320 km with NNE–SSW strike and 30–45 km width. The initial rifting phase occurred in early Oligocene time with east–west extension which was followed by a transtensional phase in Oligocene and Miocene time.

Volcanism played a very minor role in the URG and it was mainly concentrated in the south at the Kaiserstuhl volcanic complex about 15–20 Ma.

The shallow sedimentary structure of the URG is well known from numerous seismic reflections surveys for hydrocarbon reservoirs (Doebel & Olbrecht 1974), recent geothermal prospection lines and scientific deep reflection imaging (Brun et al. 1991; Wenzel et al. 1991; Mayer et al. 1997). These reflection surveys indicate alternating half grabens with listric faulting (Meier & Eisbacher 1991; Brun et al. 1991; Wenzel et al. 1991; Derer et al. 2005). There are reliable crustal models based on seismic refraction data for the southeastern shoulder (Black Forest), however, inside the URG and on the western rift shoulder region the observational constraints are sparse (Prodehl et al. 1976) and inconsistently interpreted (Edel et al. 1975; Zucca 1984). Potential field data (Rotstein et al. 2006) were used to discriminate tectonic structures at different scales.
Concerning the upper mantle structure, the hypothesis of a mantle cushion (Illies 1972) was based on early seismic refraction work (Mueller et al. 1967; Rhine Graben Research Group for Explosion Seismology 1974). This anomalous zone was interpreted as updoming mantle, which would thus favour a pure shear rifting mechanism. Later specific seismological experiments for analysis of the subcrustal structure gained teleseismic traveltime data in the southern part of the URG (Glahn et al. 1993). Surprisingly Glahn et al. (1993) and Achauer & Masson (2002) could not find a significant anomaly in the upper mantle. Especially, no upwarp of the asthenosphere below the URG was found (Davis et al. 1993) in contrast to earlier studies (Rhine Graben Research Group for Explosion Seismology 1974, Wenderoth 1978).

Our specific study area is in the central to northern part of the URG (Fig. 2) were information on the subcrustal lithosphere and upper mantle is missing yet. There the URG is about 35–40 km wide and the half graben has its main boundary fault at the eastern side where the thickest sediments (~4.8 km) occur. In the western part of the rift, in the area around Landau, there is an elevated heat flow with up to 150 mW m⁻² (Hurtig et al. 1991). This area is currently a focus of geothermal energy exploration. The deep origin of this geothermal anomaly is not completely known, however upper crustal fluid flow seems to be crucial (Clauser et al. 2002).

The TIMO (Tiefenstruktur des Mittleren Oberrheingrabens—Deep Structure of the Central Upper Rhine Graben) experiment was conducted from 2004 December until 2006 May to measure seismological waveforms for studying the deep structure of the URG (Ritter et al. 2008). A portable network of broad-band stations from the Karlsruhe Broad band Array (KABBA) was deployed between permanent broad-band and short period stations (Fig. 2). The KABBA stations were placed along two lines: a main east–west line from Stuttgart (STU) to Walferdange (WLF) in Luxembourg. East of the URG this main line is placed on the transition between the Kraichgau depression in the north and the Variscan Black Forest Mountains in the south. West of the URG the main line crosses the Pfälzer Bergland, the Saar-Nahe depression and the Rhenish Massif. The second, shorter SW–NE line reaches from the Landau area towards Heidelberg and into the northern Kraichgau depression as well as the southern Odenwald Mountains. For this study we use recordings from 31 KABBA stations, four broad-band permanent stations (BFO, ECH, STU and WLF) and four short-period permanent stations (DUP, IMS, PEB and RIVT). More specific information on the network and the recording can be found in Ritter et al. (2008).

The epicentre distribution of the studied earthquakes is shown in Fig. 3. The 92 events have a minimum distance of 30° to the TIMO network. The epicentres in Fig. 3 are relatively evenly distributed around the TIMO network, including for example, rare earthquakes in Africa (Mozambique, Tanzania) or the Sumatra 2004–2005 sequence. Earthquakes below the Tonga-Fiji region (Fig. 3b) provide steeply incident core phases (PKP) to the data set.

3. METHOD

3.1 Traveltime residuals

We measure waveforms of teleseismic earthquakes and determine traveltime residuals as well as slowness and backazimuth deviations relative to the standard earth model *iasp91* (Kennett & Engdahl 1991). Especially relative traveltime residuals are calculated which
are corrected for known crustal traveltime effects (see Section 5). A traveltime residual $r_{ij}$ is defined by

$$r_{ij} = t_{ij}^m - t_{ij}^t.$$  

(1)

This means that the measured arrival time $t_{ij}^m$ of a teleseismic phase $j$ from an event at station $i$ is subtracted with the theoretical $t_{ij}^t$ arrival time derived from an earth model (e.g. iasp91). This residual can be affected by source uncertainties (origin time, hypocentre location), which cause a systematic bias to $r_{ij}$. Furthermore, source side heterogeneity can contribute to $r_{ij}$ what is a traveltime unwanted portion, if one wants to derive information underneath the receiver network. Thus relative residuals are commonly used for which the average value of all measurements ($n$) of a phase $j$ is subtracted:

$$r'_{ij} = r_{ij} - \frac{1}{n} \sum_{j=1}^{n} r_{ij}.$$  

(2)

Physically this means that the residual part is eliminated which is common in all measurements. Then that portion of the traveltime residual is preserved which is caused by heterogeneity below the

Figure 3. Distribution of earthquake epicentres, which are used in this study. The triangle corresponds to the location of the TIMO network; (a) 84 events with mantle phases as first arrivals, (b) 8 events with core phases as first arrivals.
receiver network when the ray bundle splits up on its way to the individual stations. This normalization produces an equal amount of positive and negative residual times per phase with a zero mean. To account for variations in signal-to-noise ratio (s/n) of the picked phases, a weighting scheme can be introduced for tomographic inversions (Evans & Achauer 1993). The weights \( w_{ij} \) are selected while picking the arrival times and estimating their time uncertainties \( \sigma_i \) and their s/n. In our case four different weights \( A, B, C \) and \( X \) are chosen: \( A = 1.0 \) for \( \sigma_i \leq 0.05 \ s, \ B = 0.1 \) for \( 0.05 \ s < \sigma_i \leq 0.1 \ s, \ C = 0.04 \) for \( 0.1 \ s < \sigma_i \leq 0.2 \ s \) and \( X \) for \( \sigma_i > 0.2 \ s \). Picks with quality \( X \) are excluded from the further analysis. Our final data set includes 479 \( A \) picks, 873 \( B \) picks and 1160 \( C \) picks and the weights are not accounted for in the data plots.

As we are interested in relative residuals, we need not to pick the absolute first arrival phase, which may be partly hidden in the background noise. Instead we pick the first phase (peak or trough), which is clearly visible across the whole station network. Picking is done by hand at high magnification on the computer screen, allowing us to be precise as a sample. Our lowest sampling rate is 0.02 s and we aim at a picking with a precision of at least 0.05 s. More details on the picking and waveform analysis including record sections can be found in the electronic supporting information. The relative residuals have an error of mostly less than 0.1 s, which is less than the following crustal corrections.

The \( r_{ij}^{\text{crust}} \) are commonly used as input data for teleseismic tomography (Evans & Achauer 1993). If one wants to determine a high-resolution mantle model, heterogeneous crustal structures may cause complications, because the inversion of residuals along steep teleseismic rays can produce unwanted smearing effects (Evans & Achauer 1993; Lippitsch et al. 2005). This problem can be mitigated if there is an appropriate crustal seismic velocity model available. Then one can correct for the crustal traveltimes \( r_{ij}^{\text{crust}} \) relative to a background earth model

\[
t_{ij}^{\text{mcorr}} = t_{ij}^{\text{m}} - r_{ij}^{\text{crust}}.
\]

In this case crust corrected traveltimes (eq. 3) are inserted in eq. (1). Our crustal correction is outlined in chapter 5.

The normalization in eq. (2) has a profound influence on the residuals and their interpretation. In Fig. 4 we show three station network configurations above a seismic low velocity anomaly. The configurations have a completely different effect on the calculated relative residuals: in a background medium \((v_p = 8.0 \text{ km s}^{-1})\) an anomaly with \(-3\) per cent \(v_p\) \((7.76 \text{ km s}^{-1})\) is placed. Its vertical extension is 50 km, and for simplicity we assume vertical wave propagation. Stations above the anomaly are delayed by an absolute residual of \(+0.1930\) s compared to stations outside the anomaly (absolute delay \(0.0\) s). In Fig. 4(a) four stations are above the anomaly and four stations are outside. Thus the relative residuals are \(-0.0965\) s and \(+0.0965\) s. This means that the traveltimes anomaly is equally distributed across the network. The advanced arrivals outside the anomaly are of course not due to an increase of velocity relative to the background, but they occur only due to the way the relative residuals are calculated! The network in Fig. 4(b) covers the anomaly with only one station at which a relative residual of \(0.1654\) s is determined, whereas residuals of \(-0.0276\) s are calculated at the six surrounding stations. This residual pattern nicely indicates the low velocity anomaly at depth. The opposite situation is assumed in Fig. 4(c): four stations on the anomaly have a small relative residual of \(0.0386\) s and the only station outside the anomaly has \(-0.1544\) s. This finding may lead to the wrong interpretation that the one station outside the low velocity region may be situated above a high-velocity heterogeneity while the other four station are situated above the ‘normal’ background. These examples demonstrate that relative residuals must be treated with care. The only meaningful interpretation is the contrast between different stations with is constantly \(0.1930\) s in all three cases outlined above.

The measured waveforms from the TIMO network are bandpass filtered from 0.5 Hz to 2.0 Hz with a fourth-order Butterworth filter to improve the s/n. This narrow high frequency band allows us to study the uppermost mantle without caring about finite frequency effects (Nolet & Dahlen 2000). The diagrams in Fig. 5 present the residuals at two TIMO stations as function of slowness \(p\) \((0.10–10\ \text{deg}^{-1}\) along the radial axis) and backazimuth \(RAZ\). Steeply arriving phases with low \(p\) values appear closer to the centre, more inclined arriving phases are plotted closer to the edge of the diagram. At station TM011, on the western rift shoulder close to the edge of the URG, small residuals of \(-0.1\) s to \(+0.1\) s are found. Slightly faster arrivals are systematically observed for waves approaching from north, whereas arrivals from west and east are systematically slightly delayed. However these variations are mostly within the error bars \((\pm 0.14\) s) for the crustal correction (see Section 5) and lack any significant trend. In contrast at station TM010, inside the rift, the arrivals are delayed from all \(RAZ\) with a positive small residual amplitude of just \(0.15–0.30\) s.

### 3.2 Slowness and backazimuth residuals

In addition to the traveltimes residuals, two more parameters can be used to analyse the teleseismic wave front. First, the slowness \(p\) can be measured:

\[
p = \frac{r \sin(i)}{v}
\]

with \(r\) distance from the centre of the Earth, \(i\) incidence angle \((i = 0^\circ\) for vertical incidence) and \(v\) seismic velocity. \(p\) is constant along a ray path from the source to the receiver. Thus changes in \(v\) cause changes in \(i\), for example, a velocity reduction leads to a decrease in \(i\) or a steeper ray path. Secondly, the backazimuth, the angle between north and the epicentre, can be analysed. Again
changes in $v$ cause deviations of the wave front from the expected $BAZ$ (Krüger & Weber 1992; Bokelmann 1995). To measure the $p$ and $BAZ$ deviations we use array-processing techniques. Both parameters are determined by fitting the seismic arrival times to a plane wave front (Filson 1975).

The TIMO station network is divided into subarrays to be able to recover variations in $p$ and $BAZ$ across the area. Several station configurations were tested for the subarrays. Stable results were achieved by selecting three subarrays: SW (all stations on the western shoulder), SG (all stations inside the URG) and SE (all stations on the eastern shoulder). The subarrays are outlined in Fig. 2. The accuracy of our $p$ and $BAZ$ measurements is about 0.2 s deg$^{-1}$ and 2$^\circ$ outside the URG and about 0.2 s deg$^{-1}$ and 4$^\circ$ inside the URG, respectively. Slowness deviations are interpreted relative to a background earth model ($iasp91$ in our case) and $BAZ$ deviations are interpreted relative to the theoretical $BAZ$ between the source location and the subarray centre.

4. SYNTHETIC ANOMALIES

Synthetic traveltime modelling is used to determine the effect of a mantle anomaly on a teleseismic wave front. The same source and receiver configuration, and hence ray distribution, is taken as for the TIMO experiment. The background seismic velocity model corresponds to $iasp91$ (Kennett & Engdahl 1991). The crust has a $v_p$ gradient from 5.8 km s$^{-1}$ at the surface to 6.5 km s$^{-1}$ at the base of the crust. In the mantle $v_p$ increases slightly with depth starting with 8.04 km s$^{-1}$ below the Moho. The input anomaly (Fig. 6) is placed into the lithospheric mantle between 35 km depth (Moho) and 100 km depth (approximate depth of the LAB) with a lateral extension of 50 km in east–west direction and 35 km in north–south direction. The centre of the anomaly is placed underneath the western part of the rift (just east of station TMO19 in the case of the TIMO experiment). Inside this anomaly the velocity perturbation is variable for different scenarios: model SM0 has a reduction of $-3$ per cent $v_p$ at the top and $-1$ per cent $v_p$ at the bottom of
the anomaly. SM0 represents a very small perturbation in the lower lithosphere, compatible with about 100 °C temperature increase, for example, due to a small asthenospheric upwelling. It may also represent some fluid content or a disturbed lithospheric mantle due to mechanical reworking during the rift process. Models SM1–SM5, SM7 and SM10 have anomalies with homogeneous perturbations of −1 per cent $V_p$ to −10 per cent $V_p$, respectively. The borders of the anomalies are treated as 5 km wide horizontal and 2 km wide vertical gradient zones between the background and anomaly seismic velocity. These models represent more pronounced lower lithospheric anomalies and/or asthenospheric upwellings as known from the Rio Grande rift with −5 per cent $V_p$ (Gao et al. 2004) or the Kenya rift with −6 per cent $V_p$ to −12 per cent $V_p$ (Achauer et al. 1994).

3-D ray tracing through the synthetic models is performed using the ray bending method with a simplex algorithm by Steck & Prothero (1991). The ray tracing grid has horizontal extensions of ±1000 km (from the centre of the model) and extends to 300 km in vertical direction. The lateral grid points are at variable distances: 5 km distance in the inner ±300 km of the model, 10 km distance from 300 to 600 km, 20 km distance from 600 to 860 km and two grid point rows at 910 km and 1000 km. Horizontal grid point layers are at 0 km, 34 km, 36 km, 99 km, 101 km and 175 km depth. Between these grid point layers the seismic velocity is linearly interpolated. The step length for calculating the ray paths is 1 km. Traveltime, slowness and backazimuth perturbations are calculated relative to the *iasp91* background model without any anomaly.

In Fig. 7 the synthetic traveltime residuals for input model SM0 are plotted at the station sites as function of slowness and backazimuth in the same way as in Fig. 5. The residuals for the complete network in Fig. 7(a) have contrast values of mostly 0.2–0.5 s, the maximum contrast reaches 0.85 s. Outside the anomaly the values are mostly close to 0 s. A more detailed map in Fig. 7(b) displays variations close to the anomaly below the assumed URG. The azimuthal distribution of the residuals reflects the length of their ray path inside the anomaly: for example, stations in the south of the anomaly have traveltime delays for northern ray paths but hardly any delay for southern ray paths. Stations on the western rift shoulder, close or atop the anomaly, show traveltime delays for eastern ray paths but no delay for western ray paths. The contrasts of the residuals increase with increasing velocity perturbation $\Delta V_p$ reaching more than 1 s for $\Delta V_p$ exceeding 9 per cent (model SM10).

The synthetic slowness values are shown in Fig. 8 for the three subarrays and velocity perturbations ranging from 0 to −10 per cent (models SM1–SM10). The input wave front has a backazimuth of 87° (corresponding to a wave from Sumatra arriving at the URG). The synthetic slowness is about 5.4 s deg$^{-1}$ at the eastern subarray SE (diamond symbols in Fig. 8) and thus it is not affected relative to the value of 5.4 s deg$^{-1}$ of an undisturbed wave front. This test demonstrates that the waves did not travel through the anomaly, which is further west in the models. Stations west of the anomaly (subarray SW, triangles in Fig. 8) record a decrease in $\rho$, which decreases further with increasing negative $\Delta V_p$ amplitude inside the anomaly. This result can be explained with steeper ray paths (decreasing incidence angle) at subarray SW compared to a scenario without anomaly. In this way the rays, approaching from southwest, bypass the low velocity anomaly according to Fermat’s principle.

Atop the anomaly at subarray SG a clear effect on $\rho$ is shown in Fig. 8 (rectangles): for the used geometry of the anomaly a gradient $d\rho/\Delta V_p$ of about −0.5 s deg$^{-1}$ per percent $\Delta V_p$ is found. Compared to the resolution of actual measurements (~0.2 s deg$^{-1}$) this means that an anomaly with $\Delta V_p > 2$ per cent should be identified as slowness anomaly. Especially if ray paths from different BAZ are analysed, then a characteristic pattern of slowness anomalies should be found at the subarrays SW, SG and SE.

In the same way as the $\rho$ deviations we show the BAZ deviations as functions of $\Delta V_p$ in Fig. 9. Again at subarrays SE and SW hardly any anomaly is found; the small deviations of ±1° from the theoretical value of 87° are due to some overlap of the anomaly with SE and SW, but they are below the resolution of real data. For subarray SG we find a small increase of BAZ for increasing $\Delta V_p$ amplitude with a gradient of about 0.6° per percent $\Delta V_p$. The BAZ deviations in Fig. 9 are relatively small and hard to recover in real data. However, if wave fronts from different BAZ are observed, then systematic pattern may be recovered at the three subarrays, if the anomaly has a significant $\Delta V_p$ of at least −5 per cent.

5. CRUSTAL CORRECTION

The crust of a continental rift is heterogeneous and causes significant traveltime residuals. Major delays (up to one second) can be caused by the sediment basins which are filled with seismic low velocity material and by Moho topography which laterally replaces mantle ($v_p \approx 8$ km s$^{-1}$) and crustal ($v_p \approx 6.5$ km s$^{-1}$) material. For subcrustal studies it must be avoided that such crustal residuals are mapped into the mantle. Therefore the crustal residuals should be determined *a priori* and separated (Lippisch et al. 2003; Martin et al. 2005). As reference we use the *iasp91* reference earth model (Kennett & Engdahl 1991). The *iasp91* model has a 20 km thick upper crust with a $v_p$ of 5.8 km s$^{-1}$ and a 15 km thick lower crust with a $v_p$ of 6.5 km s$^{-1}$. The *iasp91* Moho depth is at 35 km, about 5–8 km deeper than below the TIMO network. Our synthetic model (Fig. 6) has a similar crust as *iasp91* with a velocity gradient from 5.8 km s$^{-1}$ to 6.5 km s$^{-1}$. For comparison with measurements of the TIMO experiment we have to correct the real data set from the URG for all known crustal inhomogeneities. Our approach is described in Martin et al. (2005). For each station site we gathered the available information on station elevation, sediment thickness, sediment type and Moho depth (Table 1). The sediments were divided in two units: loose, mostly Cenozoic sediments ($v_{sed} \sim 2.7$ km s$^{-1}$) and more compact Mesozoic and Paleozoic sediments ($v_{sed} \sim 4–5$ km s$^{-1}$). Moho depth below the TIMO network is about 27 km at the centre with smaller values at the south and larger values towards west (Table 1). The Moho topography is displayed in Fig. 10 and it is quite similar to the European Moho map by Grad et al. (2008). For several station sites different values for the parameters in Table 1 were found in different references. For these ambiguous cases the average or a best guess parameter (e.g. based on methodology or up-to-datenness) was chosen.

The crust corrected traveltimes (eq. 4) are calculated based on the slowness of each associated teleseismic phase. In Fig. 10 the correction times are shown for a phase with 5 s deg$^{-1}$ slowness. The biggest delays (0.3–0.4 s) occur in the eastern part of the URG and are due to the thickest rift sediments (~3.7–4.8 km) there. In the western part the rift sediments are less thick due to the half-graben geometry. Outside the URG there are negative delays (advanced arrivals) due to the thinner crust (~26–28 km in the region) compared to the *iasp91* background model (35 km).
Figure 7. Synthetic relative traveltime residuals for model SM0, (a) complete TIMO network, (b) central URG region. The residuals are plotted at the station sites as function of slowness and backazimuth (see Fig. 5). The black square outlines the position of the synthetic velocity anomaly in the upper mantle. Cities: BAD, Baden-Baden; HD, Heidelberg; KA, Karlsruhe; KL, Kaiserslautern; LD, Landau; LUX, Luxembourg; M, Metz; MZ, Mainz; N, Nancy; S, Stuttgart; SB, Saarbrücken; SP, Speyer; STB, Strasbourg; TU, Tübingen; TR, Trier.

The only exception is the Saar-Nahe basin, west of Saarbrücken, where 3.0–3.3 km thick sediments cause a small delay of about 0.2 s.

The estimated maximum uncertainties are 1 km in sediment thickness and 2 km in Moho depth. Ray tracing tests indicate maximum errors of ±0.14 s in the crustal corrections with these uncertainties. However, much smaller uncertainties for the sediment thickness are more realistic at most sites and hence our correction times are probably more precise than the above given maximum values of ±0.14 s.
6. Application to TIMO Data Set

The measured crust corrected relative traveltime residuals, slowness and backazimuth values are shown in Figs 11–13. The measured residuals in Fig. 11 are plotted in the same way and with the same scale as the synthetic residuals in Fig. 7 for comparison. The map with the residuals in Fig. 11(a) displays an overview and contains two anomalies: at the western stations clearly delayed arrivals are observed reaching 0.6–0.7 s and inside the URG an anomaly with about 0.2–0.3 s delay are found. The other TIMO stations do not show any specific anomaly pattern (Fig. 11a). Most residuals are slightly negative (up to −0.1 s and −0.2 s) to compensate the larger positive anomalies (see also Section 4 on residual distributions). In Fig. 11(b) stations west of 7° E are excluded to achieve a higher resolution of the mapped residuals and allow us a more detailed discussion. In the URG the main but small traveltime contrast is found at stations in the western part in the area of Landau (stations TMO08–10, 19, 22, 24) with delays of about 0.2–0.3 s relative to their surrounding stations. Such a delay is relatively low compared to other continental rifts (see Section 7) and it is relatively locally confined.

At the stations in the western URG the delay seems to be independent from BAZ and p (and thus incidence angle). The spatial distribution of the residuals does not show any specific pattern for stations TMO19, 22 and 24. Such a uniform distribution pattern is an indication that the delay is caused by a shallow, upper crustal anomaly, which delays all teleseismic arrivals irrespective of their direction of approach (Fig. 15 and interpretation below). At stations TMO08–10 arrivals from southeast are not delayed, indicating that the upper crustal anomaly terminates in this direction. At station TMO07 in the west of Karlsruhe, all arrivals from sources in the east are not delayed, indicating the eastern rim of the shallow anomaly to be located at about the Rhine River. Towards south the termination of the anomaly cannot be determined due to missing stations. Towards west a comparison with neighbouring stations, located on the western rift boundary (TMO11 and 21), shows an abrupt change in the residual pattern (Figs 5 & 11): on the rift shoulder the amplitude of the residuals is much smaller (about ±0.15 s) also with no specific spatial pattern. Especially for teleseismic rays arriving from east, that travelled underneath the URG and arrived on the western shoulder, there is no delay. Inside the western URG the delay is about 0.2–0.3 s independent of the ray path direction. This abrupt contrast is a strong indication that the delays observed inside the western URG have a shallow, upper crustal origin. Along the eastern rift boundary, near the cities of Karlsruhe and Heidelberg, we do not find a similar difference between residuals determined inside the rift and residuals on the rift shoulder (Fig. 11).

The measured deviations of p (Δp) are shown in Fig. 12 for the subarrays SW, SE and SG as a function of the theoretical BAZ. Generally the Δp values are mostly below 1 s deg⁻¹. At subarrays SW and SE on the rift shoulders the variation of Δp is less compared to Δp at subarray SG inside the rift. This difference is partly due to the larger standard errors of the measurements at SG with fewer and noisier teleseismic arrivals compared to the measurements at the subarrays SW and SE on the rift shoulders (see Section 3.1). Compared to the synthetic modelling results in Fig. 8, the measured Δp exclude a mantle anomaly underneath the URG with a Δvp of more than 1–2 per cent positive or negative amplitude, because such a Δvp would cause a measurable Δp. As no direction (BAZ) dependent anomaly is found in our data set (Figs 11 and 12), even such a low amplitude Δvp anomaly is unlikely to exist in the upper mantle below the TIMO network.

The measured BAZ deviations (ΔBAZ) are also very small at the three subarrays (Fig. 13). As for Δp the standard errors are largest for subarray SG due to the noisier site conditions inside the URG. Most measured ΔBAZ are less than 5°, which is in the range...
of the resolution error. There is no systematic trend with BAZ in \( \Delta BAZ \), which could indicate an orderly deviation of the teleseismic wave fronts. The synthetic BAZ data in Fig. 9 show that systematic trends in \( \Delta BAZ \) should occur at the different subarrays in case a velocity anomaly \( \Delta v_p \) of more than 5 per cent is present in the lower lithosphere. In summary the measured traveltime, slowness and backazimuth perturbations across the URG do not indicate evidence for significant deep anomaly, instead only a minor shallow anomaly may be supported.

The absolutely largest residuals within the TIMO network are found at the westernmost stations WLF and TIMO18 (Fig. 11a), which are off the URG but close to the Eifel region where an upper mantle plume was found (Ritter et al. 2001). In Fig. 14 the residuals are displayed in detailed diagrams. The del ay of about 0.5 s or contrast of 0.6–0.7 s relative to the TIMO stations on the western shoulder of the URG are comparable to the residuals found by Ritter et al. (2000, 2001). The residuals do not have a significant direction depend pattern indicating that nearly all ray paths are affected by the velocity anomaly at depth. This is compatible with a wide low velocity anomaly in the upper mantle underneath the stations.

### 7. Interpretation

The amplitudes of the observed traveltime, slowness and backazimuth residuals across the URG are very small and therefore a tomographic inversion for 3-D velocity perturbations makes no
sense. In the central URG the only resolved anomaly of 0.2–0.3 s delay occurs in a confined area around Landau. At other continental rifts much larger residuals are determined between the rifts and their shoulders: Rio Grande rift: 1 s residual (Davis et al. 1993), Baikal rift: 1.6 s (Gao et al. 1994), Kenya rift: 2.5 s (Achauer 1992), Ethiopia rift: 2.5 s (Bastow et al. 2008). At these rifts low-velocity anomalies reaching $\Delta v_p$ of up to −20 per cent are assumed and interpreted as deep lithospheric and asthenospheric rift-related processes.

7.1 Mantle

Besides the shallow, crustal anomaly around Landau (see Section 7.2) and the anomaly at the western end in the Eifel, our residuals in Figs 11–13 across the central URG do not contain any systematic signal. Therefore we exclude any traveltime anomaly exceeding about 0.15 s, which corresponds to our resolution. This means that we can exclude the following anomalies at mantle depth: (i) A low- or high-velocity body exceeding 1–1.5 per cent $\Delta v_p$ and a vertical extension of about 65 km comparable to the anomaly in Figs 6 and 15(c). Due to the trade-off between vertical extension (corresponding to wavepath length) and velocity contrast, alternatively a 30 km thick anomaly could not have a $\Delta v_p$ of 2 per cent to 3 per cent without being detected. This excludes an upwelling hot (excess temperature more than 100 °C) asthenospheric body below the central URG. (ii) A significant Moho topography is unlikely. Our crustal corrections assume a rather flat Moho at about 27 km depth across the central URG (Table 1 and Fig. 10). A 3 km or larger additional Moho topography with a 6.5 km s$^{-1}$ to 8.0 km s$^{-1}$ seismic velocity contrast would cause a 0.1 s residual, which we can exclude from our analysis.

Our results are only partly comparable with earlier measurements in the southern URG, close to Strasbourg and further south. There Glahn et al. (1993) found crust corrected residuals of about 0.5 s. Their tomography model excludes a low-velocity anomaly below the southern URG and even contains a slight high-velocity (+2 per cent $\Delta v_p$) anomaly in the mantle below 50 km depth underneath the southern URG (Davis et al. 1993; Glahn et al. 1993). This model was a surprise since previous models predicted a low-velocity anomaly, sometimes called a mantle cushion, below the URG (Rhine Graben Research Group for Explosion Seismology 1974; Wenderoth 1978). Achauer & Masson (2002) inverted teleseismic residuals from different experiments in the region and also found no clear seismic velocity anomaly related with the URG. Our data set with a greater amount and more consistent measurements indicates that the central URG is not associated with a resolvable seismic low-velocity anomaly in the upper mantle. Clauser et al. (2002) analyse chemical tracers and helium in mineral waters and do not find a magmatic mantle component in the central URG. Their results are compatible with our finding that there is no hot mantle cushion below the central URG.

A recent study of SKS splitting with the TIMO data set (Wagner 2007) also did not resolve any mantle anomaly related with the URG. Thus we do not expect a major modification of the lower lithosphere or asthenosphere below the URG. As consequence we favour a rifting mechanism, which did not penetratively modify the upper mantle underneath the URG. This interpretation coincides with a passive rifting mechanism such as
Figure 11. Measured relative traveltime residuals of the TIMO project after a crustal correction (a) complete TIMO network, (b) central URG region. The residuals are plotted at the station sites as function of slowness and backazimuth (see Fig. 5). Cities: BAD, Baden-Baden; HD, Heidelberg; KA, Karlsruhe; KL, Kaiserslautern; LD, Landau; LUX, Luxembourg; M, Metz; MZ, Mainz; N, Nancy; S, Stuttgart; SB, Saarbrücken; SP, Speyer; STB, Strasbourg; TU, Tübingen; TR, Trier.

as shearing along a pre-existing and reactivated lithospheric detachment complex (Eisbacher & Fielitz 2008) in a tensile stress regime in the hinterland of the Alpine collision zone (Dezes et al. 2004).

7.2 Crust

A resolved anomaly is found in the western URG around Landau. The spatial pattern of the residuals, including their $p$ and $BAZ$ dependencies (Fig. 11), indicates a shallow origin, presumably in the
upper crust. This depth estimation can be also deduced from ray path diagrams. In Fig. 15 three velocity anomalies ($\Delta v_p = -3$ per cent) are placed in the upper crust, lower crust and uppermost mantle. Depending on the depth of the anomaly, traveltime residuals are observed at a wider station range (Fig. 15c) or a very narrow station range (Fig. 15a). The triangles in Fig. 15 correspond to the TIMO stations; white stations are not affected by traveltime anomalies, grey stations are affected only from one backazimuth (here either from east or west) and residuals of black stations are influenced from all backazimuths. Compared to the distribution of residuals at stations TMO11 and TMO10 in Fig. 5 the anomaly must be as shallow as the upper crust (Fig. 15a). For a deeper anomaly in the lower crust as in Fig. 15(b) the residuals from eastern sources

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**Figure 12.** Measured deviations of the slowness ($\Delta p$) at the three subarray (a) SW, (b) SG and (c) SE. The error bars correspond to the standard deviations.

**Figure 13.** Measured deviations of the backazimuth ($\Delta BAZ$) at the three subarray (a) SW, (b) SG and (c) SE. The error bars correspond to the standard deviations.
at station TMO11 must be delayed, however this is not observed (Fig. 5). In addition the synthetic residuals from western sources at station TMO10 are not affected by a lower crustal anomaly but delayed residuals are observed (Fig. 11b). Furthermore the concise distribution of positively delayed residuals in Fig. 11(b) indicates a shallow origin in the upper crust.

There are two possible explanations for this shallow anomaly with an average traveltime delay of 0.2–0.3 s: (i) inadequate crustal correction or (ii) a local seismic velocity anomaly in the crystalline basement. This region is tectonically quite complicated and characterized by a horst and graben structure inside the rift basin (Eisbacher & Fielitz 2010). A delay of 0.2–0.3 s can be explained with wave propagation along 350–550 m thicker sediments ($v_p \sim 4.5$ km s$^{-1}$) as given in Table 1 for the crustal correction replacing ray paths in the crystalline basement (granite with $\sim 6$ kms$^{-1}$). Such an error in our crustal model may be possible and cannot be completely excluded. However, deep boreholes for oil and geothermal energy provide good constraints on the sediment layers. Another option to explain the residuals may be highly modified granite in the uppermost crust, which was observed in some deep boreholes (B. Schmidt, personal communication, 2009). The location of the anomaly coincides with the heat flow anomaly (about $+50$ mW m$^{-2}$ relative to the surrounding URG), which is harvested as geothermal power. The estimated maximum temperature ($T$) anomaly is about $+40$ °C (Schulz & Schnellschmidt 1991) in the Landau area. Using a gradient $d v_p / d T$ of $-0.39 \times 10^{-3}$ km s$^{-1}$ °C$^{-1}$ for granite (Christensen & Mooney 1995), one can explain a delay of 0.004 s or 0.008 s for a 10 km or 20 km long propagation path, respectively. Thus a thermal effect alone cannot explain the travel time delay. However, our anomaly may be explained with a combined effect of sediment thickness variation together with heated and modified granite in the upper crust.

8. CONCLUSIONS

The TIMO experiment allows us the precise determination of teleseismic wave front anomalies in the central part of the URG. Synthetic examples show that traveltime, backazimuth and slowness deviations occur for velocity perturbations of about 3–5 per cent of \sim 50 km large upper mantle anomalies. These perturbations should also depend on the backazimuth of the incoming wave front and generate distinct deviation pattern. Wave fronts from 92 earthquakes were analysed at up to 39 stations. Our measurements do not reveal significant slowness or backazimuth deviations of the teleseismic wave fronts relative to the *iasp91* earth model (Kennett & Engdahl 1991). This finding excludes a seismic velocity anomaly ($\Delta v_p$) of more than $\pm 2$ per cent in the upper mantle and is consistent with the determined spatial pattern of traveltime residuals. We interpret these results as evidence for a minor modification of the lower lithosphere and/or asthenosphere below the central URG. Therefore, a passive rifting process is proposed, which left the deep structure more or less intact. Previous models with a mantle cushion or active magmatism below this part of the URG are not supported. The only minor traveltime anomaly (0.2–0.3 s delay) is found below the western part of the URG in the area of Landau. It may be related to insufficient crustal corrections, mainly inadequate sediment models, or a well-known local thermal anomaly, which modified the upper crustal basement. At the western end of our network we measured residuals reaching 0.7 s delay, which can be explained with the known mantle plume underneath the Eifel.

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Figure 15. Ray paths through synthetic seismic anomaly models (−3 per cent $\Delta v_p$) below the southern E–W station line of the TIMO network. Two representative wave fronts are plotted: source one in the west corresponds to a BAZ = 266° with $p = 4.7$ s deg$^{-1}$ (South America) and source two in the east with BAZ = 87° and $p = 5.4$ s deg$^{-1}$ (Sumatra). (a) Upper crustal anomaly at 5–20 km depth and 20 km lateral extension, (b) lower crustal anomaly at 20–35 km depth with 20 km lateral extension, (c) mantle anomaly at 35–100 km depth with 50 km lateral extension. Black stations (triangles) indicate that the anomaly causes a traveltime residual for both (east and west) backazimuths, grey stations are influenced only from one backazimuth region and white stations are not affected. SW, SG and SE indicate the position of the subarrays (see Fig. 2).

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