Crustal structure and high-resolution Moho topography across the Rwenzori region (Albertine rift) from P-receiver functions

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Abstract: The Rwenzori region, which is located between the Democratic Republic of Congo and Uganda, is part of the western branch of the East African Rift. With elevations of c. 5000 m a.s.l., the Rwenzori Mountains are situated between the Albert Rift and the Edward Rift segments and cover an area of approximately 120 km by 50 km. In this study we investigate the Moho topography beneath the Rwenzori region based on data from a network of 33 broadband seismic stations that were operated from September 2009 until August 2011. Variations of crustal thickness are obtained from the H-k stacking method applied to P-receiver functions. We discuss the effect of low velocity layers within the crust on the determined Moho depths, which range from 20 km up to 39 km. The lack of a crustal root beneath the Rwenzori Mountains and its location in an extensional setting are contrary to the orogenesis generated by collisions of tectonic units. Our results indicate crustal thinning and provide evidence for the alternative mechanism of crustal bending, triggered by the tensile stress and the elasticity of the crust.

Supplementary material: Examples and methods for identifying crustal structures and sediment layers are available at http://www.geolsoc.org.uk/SUP18801.

The Rwenzori region: tectonic setting and previous studies

The East African Rift System (EARS) was initiated c. 25–30 Ma ago (Nyblade & Brazier 2002; Rychert et al. 2012 and references therein) because of plume-related flood basalts magmatism in the Afar region. The assumption of an active rift related to an upwelling of mantle plumes below Ethiopia or Kenya (Nyblade & Brazier 2002; Montelli et al. 2006; Rychert et al. 2012 and references therein) is the favoured explanation for the processes of rifting and volcanism in East Africa. Three rift zones emanated from the Afar Triple Junction: the Red Sea Rift (RSR), the Gulf of Aden Rift (GOA) and the Main Ethiopian Rift (MER) (Rychert et al. 2012). As a result of southward-propagating rifting processes, tectonic movements were initiated in the eastern branch (Gregory Rift) of the EARS about 10–12 Ma ago (Nyblade & Brazier 2002). The transfer of extensional stresses across the strong lithosphere of the Tanzania craton caused the development of the western branch (Albertine Rift) by reactivating pre-existing shear faults (Klerkx et al. 1998; Nyblade & Brazier 2002) and rift-flank uplifts about 8 Ma ago (Ring et al. 1992; Ebinger et al. 1993; Ring & Betzler 1995). An unspecified tectonic impulse at about 2–3 Ma ago started an uplift of the Rwenzori Mountains. Apatite–helium dating confirms a fast uplift during the Pliocene to Pleistocene (Bauer et al. 2010 and references therein).

The Rwenzori Mountains are situated at the border between Uganda and the Democratic Republic of Congo (Fig. 1) and represent a horst structure consisting of metamorphic basement rocks of Archaean and Palaeoproterozoic ages (Link et al. 2010). They reach elevations of more than 5 km and cover an area of c. 120 km by 50 km orientated in a NNE–SSW direction. Mountain building is generally related to the collision of tectonic units, which leads to crustal thickening in consequence of folding and overthrusting. The location of the Rwenzori Mountains in an extensional setting and the absence of a crustal root beneath the Rwenzori range (Wölbern et al. 2010) contradict this type of mountain building. Other driving forces are necessary to explain the orogenesis of the Rwenzori Mountains. According to Sachau et al. (2013), several attempts have been made previously to explain the origin of the Albertine Rift (western branch of the EARS) and the uplift of the Rwenzori Mountains, including rift-induced delamination of the mantle lithosphere due to the upwelling of asthenospheric material (Wallner & Schmeling 2010). Accordingly, the delamination and the uplift are
realized by the two regional rift segments surrounding the Rwenzori Mountains – the northern Albert Rift segment and the southern Edward Rift segment (corresponding to Lake Albert and Lake Edward) – which lead to almost complete decoupling of the Rwenzori horst structure from the up to 3 km-high rift shoulders. Sachau et al. (2013) propose an increase in tensile stress between the layers close to the surface, which causes crustal bending and uplift of the Rwenzori Mountains, controlled by viscosity and elasticity of the crust.

The Albertine Rift (see Fig. 1) is a region characterized by seismic and geothermal activity, including the Buranga hot springs and the volcanic fields of Katwe-Kikorongo and Bunyaruguru (Bahati et al. 2010). Local and teleseismic travel-time tomography (Jakovlev et al. 2011, 2013) provide evidence for active magmatic processes in the crust and upper mantle beneath the Rwenzori region in relation to the rifting process.

While numerous receiver-function studies have focused on the eastern branch of the EARS (e.g. Hebert & Langston 1985; Last et al. 1997; Dugda et al. 2005; Hammond et al. 2011), only a few have covered the Albertine Rift (Wölbern et al. 2010; Tugume et al. 2012). In contrast to the study of Wölbern et al. (2010), which was restricted to the eastern part of the Rwenzori region (grey symbols in Fig. 1), the data of our study comprise the total width of the Albertine Rift system and can, thus, give a more complete picture of the structures within a developing continental rift.

We expect our study to provide new constraints on the rifting and uplift processes that are related
to the formation of the Albertine Rift and the Rwenzori Mountains, respectively. Our aim is to determine the thinning of the crust within the rift valleys compared to the rift shoulders in order to examine whether or not the Albertine Rift is generated by pure extensional processes. We assume that an additional driving force, such as the upwelling of partial melt from the asthenosphere, could coincide with the rifting and contribute to the process of orogenesis of the Rwenzori Mountains. On the basis of average $v_p/v_s$ ratios, we want to determine the crustal composition (e.g. Christensen 1996) and possible influences of partial melts. The crustal $v_p/v_s$ ratios and the value of crustal thinning in the rift are to be accommodated to recent geodynamic models (e.g. Wallner & Schmeling 2010; Sachau et al. 2013 and references therein).

Data and method

The data that we use for our study were recorded by a temporary network of 33 seismic stations equipped either with broadband sensors of type Trillium Compact and Taurus dataloggers or with Güralp CMG-3ESP sensors and EDL dataloggers. The network was installed and operated by the Rift-Link research group from September 2009 until August 2011 (http://www.riftlink.de). In contrast to a previous network, which was restricted to the Ugandan side of the Rwenzori region, our database covers a larger area and includes also the western shoulder of the Albertine Rift (Fig. 1). The data acquisition is made on continuous mode with a sampling rate of 100 Hz.

For the P-receiver function analysis we selected teleseismic events at epicentral distances between 30° and 95° and a minimum magnitude of M 5.0 (Fig. 2, red circles). Because of the poor ray coverage from western azimuths we include additional PP-phases from earthquakes at epicentral distances greater than 95° with minimum magnitudes of M 5.5 in order to enlarge the database (Fig. 2, green circles). After quality control and initial processing, the dataset included a total number of 8176 receiver functions.

The method of P-receiver functions is an established technique and based on recordings of P-to-S converted phases (denoted Ps) that originate at seismic discontinuities (e.g. Vinnik 1977; Langston 1979) in the crust and the lithospheric mantle beneath the receiving station. The observed travel time differences between the direct and converted wave can be analysed to estimate the depths of the discontinuities, provided that the seismic velocities are sufficiently well constrained. The method has been widely used to map the Moho and prominent upper mantle discontinuities by utilizing not only the direct converted phase Ps but also multiple phases such as PpPs and PpSs. In this article, we apply the processing steps as outlined by Yuan et al. (1997) and Wölbern et al. (2009). In order to separate the direct P and the converted S phases, the Z, N and E components of the recorded signal are rotated into the ray coordinate system with the components L, Q and T that are aligned parallel to the P, SV and SH polarizations, respectively. To compensate for different source–time functions and path effects, the waveform of the P-wave, as recorded on the L component, is deconvolved from the Q and T components. This will transfer the converted energy into (smoothed) spikes. Finally, the deconvolved traces are stacked at each station in order to further amplify the converted spikes. Before stacking, a moveout correction is applied to account for variations in epicentral distances, which affect the delay times of the converted phases. For this purpose the traces are compressed or stretched according to an average slowness of 6.4 s per degree, which corresponds to an epicentral distance of 67° (Kennett & Engdahl 1991).

The H-κ stacking method of Zhu & Kanamori (2000) is an effective algorithm to estimate Moho depths. It is based on the calculation of theoretical delay times of the converted phases Ps, PpPs and PpSs as a function of conversion depth $H$ and the $v_p/v_s$ ratio $κ$. 

![Fig. 2. Distribution of seismic events used for P-receiver analysis in this study. Concentric circles mark epicentral distances to the Rwenzori seismic network (central asterisk). For earthquakes in the distance range between 30° and 95° (red symbols) we analysed P arrivals, whereas PP arrivals were used at epicentral distances larger than 95° (green symbols).](image_url)
We use an input value \( v_P = 6.375 \text{ km s}^{-1} \), which is taken from a \( Pn \) study of Bram (1975) for the Albertine Rift. The horizontal slowness is calculated from the hypocentral coordinates of the teleseismic events used in this study on the basis of the IASP91 velocity model (Kennett & Engdahl 1991). A grid-search algorithm is used to stack the amplitudes of \( P \), \( PpP \) and \( PpS \) phases of all individual receiver functions at the theoretical arrival times. We apply an equal weight to all three converted phases. To account for the negative amplitude of the \( PpS \) multiple, the sign is reversed before stacking. It is assumed that the amplitude stack reaches its maximum if the calculated arrival times for a certain discontinuity (e.g. the Moho) match the real arrivals recorded in the data. The related pair of \( H \) and \( \kappa \) provides the best estimate of the actual Moho depth and \( v_P/v_S \) ratio.

To estimate the uncertainties of the \( H-\kappa \) stacking results, we apply a bootstrap test. For this purpose we removed 1/e (36.8%) of the traces and complemented the dataset by randomly duplicating a corresponding number of traces of the remaining receiver functions. To investigate the influence of the crustal properties, stations located in different terrains of the seismic network are chosen. It turns out that the errors are influenced by the different crustal structures and that stations on the rift shoulders show the smallest errors (see Table 1). The selected stations are located on the western rift shoulder (station C01), on the eastern rift shoulder (station U06), north of the Rwenzori Mountains (station U01), south of the Rwenzori Mountains (station U03) and on top of the Rwenzori Mountains (station U23). Generally, the uncertainties increase with increasing complexity of the crustal structure. The errors of the determined Moho depths range from \( \pm 0.56 \text{ km} \) to \( \pm 1.91 \text{ km} \). We estimate the maximal vertical resolution of our data from the upper cut-off frequency of bandpass filters (Sheriff & Geldart 1995; Helffrich 2000), which is generally utilized for our analyses of the receiver functions. The vertical resolution varies between 3.4 km and 4.4 km. The errors of the \( v_P/v_S \) ratios range from \( \pm 0.01 \) to \( \pm 0.05 \) and increase with increasing complexity of the crust.

We further investigate the influence of different bandpass filters with corner frequencies between 0.125 Hz and 1.5 Hz, as well as the effect of different weighting factors in the \( H-\kappa \) stacking. It turns out that these parameters have only minor influence on the derived Moho depths. Taking into account all of the analysed error sources, we expect a maximum uncertainty of \( \pm 2 \text{ km} \) for the calculated Moho depths and \( \pm 0.05 \) for \( v_P/v_S \) ratios.

In addition, the value of the average crustal P velocity \( v_P \), which is an input parameter of the Zhu & Kanamori (2000) method, has a strong impact on the \( H-\kappa \) stacking results. Based on an average \( v_P = 6.375 \text{ km s}^{-1} \) (Bram 1975) we calculate Moho depths by modifying this value. We find that a \( v_P \) variation of \( \pm 0.2 \text{ km s}^{-1} \) leads to a shift in Moho depth of \( \pm 1.49 \text{ km} \) at the rift shoulder (C01), \( \pm 1.14 \text{ km} \) at the rift valley (U01) and \( \pm 1.08 \text{ km} \) at the Rwenzori Mountains. The results of crustal \( v_P/v_S \) ratios are less influenced by the P-wave velocities and exhibit minor variations from 0.003 to 0.01.

### Results

Depending on the conditions at a particular station, we obtain between 50 (station U09) and almost 400 (station U20) individual receiver functions. The conversion points at the Moho are calculated for a depth of 30 km utilizing the UVI-N74 velocity model (Bram 1975). In general they show a good azimuthal coverage (Fig. 3). We first analysed receiver functions along the three profiles (I–III), with lengths of \( c. 160 \text{ km} \), covering the rift shoulders,

| Station | \( H^1 \) | \( v_P/v_S^1 \) | \( \Delta H^2 \) | \( \Delta v_P/v_S^2 \) |
|---------|-----------|----------------|-----------------|----------------|
| C01     | 37.08 ± 0.56 | 1.72 ± 0.02 | ±1.49 | ±0.01 |
| U01     | 33.59 ± 0.87 | 1.82 ± 0.04 | ±1.14 | ±0.003 |
| U03     | 31.05 ± 1.91 | 1.87 ± 0.02 | ±1.28 | ±0.01 |
| U06     | 34.67 ± 1.85 | 1.71 ± 0.01 | ±1.24 | ±0.01 |
| U23     | 28.71 ± 1.03 | 1.86 ± 0.05 | ±1.08 | ±0.01 |
| U03     | 15.25 ± 0.53 | 1.92 ± 0.07 | ±0.59 | ±0.01 |
| U06     | 13.20 ± 1.76 | 1.96 ± 0.17 | ±0.68 | ±0.05 |

Error estimation of \( H \) and \( v_P/v_S \) by different combinations of the datasets in the bootstrap procedure\(^1\) and by the variation of the average P-wave velocity\(^2\). The selected stations present the different terrains inside and outside the rift valley and the complex structure of the crust with the existence of the LVZ.
the rift valley and the Rwenzori Mountains, respectively (Fig. 3). The individual receiver functions were stacked at each station and the sum traces are shown in the vertical sections I, II and III (Fig. 4). Receiver functions related to black piercing points in Figure 3 are not included, because they are too far from the profiles.

Profile I extends east of the Rwenzori Mountains from the Edward Rift segment in the south to the Albert Rift segment in the north. Most remarkable is a strong negative phase at delay times between 2 and 3 s in the southern part of the profile, which is situated within the rift. The traces of stations U03, U02, U21, U08 and U09 are dominated by this strong signal, which is discussed further below. We first try to identify the converted phase of the Moho in the receiver functions that corresponds to a signal of positive amplitude. The largest positive amplitudes have delay times of less than 2 s; this agrees with a conversion depth of less than 16 km, assuming the UVI-N74 velocity model (Bram 1975). But the conversion depths are in contrast to the expected thickness of continental crust in rifting systems and to earlier studies in the East African Rift (e.g. Wölbern et al. 2010; Tugume et al. 2012). Further positive signals are determined at delay times between 3 and 5 s. The average observed delay time of approximately 4 s corresponds to a conversion depth of c. 33 km, which is close to the average thickness of the continental crust. In order to properly identify direct converted and multiple phases in the receiver–function traces, the moveout correction is applied separately for each of these cases. From this, in most cases, we are able to identify the direct converted phases from the Moho. The delay times for individual stations vary between 3.1 s below the Rwenzori Mountains (station U22) and 4.5 s in the eastern rift valley (station U01; marked in green bars). At some stations (e.g. U04, U17, U22) the signal from the
Moho is not as obvious. These cases will be investigated in more detail below.

To further estimate the robustness and significance of the converted signals in the receiver functions, we applied a bootstrap test to the data, as described above. As a result, in Figure 4, additional lines denote the ±2-sigma range of the original receiver–function stack. At stations U02, U17 and U16 the traces show no or only small amplitudes at delay times of c. 4 s, probably due to the interference of additional phases from crustal structures or due to noise. The further applications of the inversion algorithm of Kosarev et al. (1993) and the method of Zhu & Kanamori (2000) help to identify the additional marked phases (green bars).

Profile II exhibits a vertical section across the Albertine Rift, including the rift shoulders, the horst structure of the Rwenzori Mountains and the rift valleys of the two surrounding rift segments. The direct converted phases of the Moho show decreasing delay times beneath the horst structure and in the western rift valley, in comparison to the rift shoulders. Exceptions are the receiver functions of stations U21 and U09 in the eastern rift valley, which are also dominated by the strong negative phase, as described in profile I.

Profile III ranges from the western rift shoulder to the eastern rift shoulder. This section thereby crosses the rift valley and the northern foothills of the Rwenzori Mountains. The Moho signals exhibit relatively small delay times of about 3 s within the Albert Rift segment (stations C04, U14 and U15). Below the eastern and western rift shoulders the delay times increase to 4 and 4.6 s, respectively.

In the vertical sections of profiles I, II and III, the delay times of positive-amplitude signals in the range between 3 and 5 s are usually reduced for stations within the rift. However, at some stations (U02, U04, U14, U17, U19 and U22) we also observe strong positive amplitudes arriving after the interpreted Moho signals with delay times between about 5 and 6 s. At these stations a Moho signal...
cannot be identified unambiguously. Moveout analysis and initial modelling of the receiver–function traces shows that in some cases (station U14 and U19) this later arriving signal can be explained to result from intercrustal discontinuities. In other cases (stations U04, U17 and U22) this phase seems to require an additional discontinuity at a depth of about 38 km. Detailed modelling or inversion of receiver functions is not considered here due its non-uniqueness (see Wölbern et al. 2010). In the following, we therefore attempt to determine the crustal thickness beneath the stations by application of the method of Zhu & Kanamori (2000).

We now utilize the H-κ stacking method of Zhu & Kanamori (2000) to determine the Moho depth beneath each station. We apply a number of corrections to derive the Moho topography in the Rwenzori region. At first, the Moho depths are determined by stacking the receiver functions within the complete azimuthal range from 0° to 360° at each station (Fig. 5a). The results show a significant variation of the crustal thickness between the rift valley and the rift shoulders. Beneath the eastern and western rift shoulders the Moho depths vary between 31 and 39 km. Beneath the rift valley and the Rwenzori Mountains the crustal thickness is reduced to minimum values of 22 and 24 km, respectively.

In order to compensate for the pronounced surface topography of the Rwenzori region, we correct all Moho depths with respect to an average altitude level of 1.3 km (Fig. 5b). This correction affects mainly stations that are located within the rift valley (e.g. U19, U02) or at relatively high altitudes (U04, U23). The Moho depth of station U23, which is located on top of the Rwenzori Mountains, is significantly reduced from 29 to 26 km.

At some stations the receiver functions showed conspicuous azimuthal dependencies of the derived crustal thicknesses. This is probably caused by local small-scale variations of the crustal structure. In most of these cases it is necessary to select a limited azimuthal range in order to obtain a stable result for the crustal thickness. For this reason, the H-κ stacking is applied separately to selected azimuthal sections. Figure 5c shows Moho depths at all stations (9) that exhibit a significant azimuthal dependence. As indicated in Figure 5c, at two stations (U20, U21) we obtain two different values for the crustal thickness depending on the azimuthal range used in the analysis. The difference ranges between 2 and 3 km.

So far, we have used the relatively simple velocity model of Bram (1975) to determine the average crustal P-wave velocity (6.375 km s⁻¹) for the H-κ stacking method, disregarding the fact that the arrival times of the converted signals are affected by local variations in the velocity structure, especially by sediments. The thickness of the sediment layer in the area of Lake Albert has previously been derived from modelling of digital elevation model topography and free-air gravity, as well as from analysis of seismic reflection data (Upcott et al. 1996; Karner et al. 2000; Karp et al. 2012). These studies indicate a maximum basin depth of c. 4–6 km beneath Lake Albert.

Here, we estimate the sediment layer thicknesses in the Rwenzori area from the analysis of local earthquake records. At stations that are situated in the rift valley we identified S-to-P conversions (S_P) from the base of the sediments. Assuming sedimentary P and S velocities, the travel-time difference between S and S_P can be used to derive the thickness of the sediments (e.g. Mandal 2007, and references therein). So far there is no information about sedimentary velocities in the Rwenzori area. To test the influence of the assumed velocities on the calculated sediment thickness we applied two different sets of v_P and v_S values (see Table 2). The thicknesses (h_1) derived with velocities from the Kenya Rift are considerably larger than those derived with velocities from the Kachchh Rift basin (h_2) and correspond better to the results of Karner et al. (2000). For this reason we utilized these values to correct the initial Moho depths for the sediments.

The derived sediment thicknesses range from 1.6 to 6.6 km and exhibit, in general, lower values in the Edward Rift segment than in the Albert Rift segment, suggesting an earlier tectonic evolution of the latter. The largest values (between 4.5 and 6.6 km) are obtained at stations just south of Lake Albert (U01, U19). Accounting for the reduced P-wave velocity within the sediments leads to a modified value for the average crustal P-wave velocity of between 5.6 and 5.8 km s⁻¹. As a result, this leads to the decrease of the Moho depths by about 4 km at these stations. The effect is less pronounced at other stations within the rift and, on average, accounts for about 2 km in reduction of Moho depth (Fig. 5d). The Moho depths beneath the Edward Rift segment reach values up to 37 km, which is significantly larger than the values obtained for the Albert Rift segment (ranging between 22 and 30 km).

A further modification of the average crustal P-wave velocity arises from a low-velocity zone (LVZ) within the crust that is obvious at some stations of the Edward Rift segment. As shown in profile I of Figure 4, strong negative phases dominate the traces between 2 and 3 s, which can be related to the upper bound of an LVZ and is confirmed by calculating receiver functions for simple crustal models. Converted phases due to the LVZ can be identified at six stations covering almost the entire Edward Rift segment up to U06 at the...
The upper bound of the LVZ is located at depths between 12 km (station U06) and 21 km (station U21) with an average depth of 17 km, which is close to the transition between upper and lower crust from model UVI-N74 of Bram (1975). The resulting $v_P/v_S$ ratios of the LVZ are generally high (Table 3) compared to the values derived at stations outside of the Edward Rift segment (see Fig. 7b). Similar to Hammond et al. (2011), we may thus argue that the LVZ is related to partial melts. Taking into account the low velocity zone in the Edward Rift segment requires an additional correction to the Moho depth because the average P velocity in the crust has to be adjusted. We determined the thickness of the LVZ at individual stations by applying the method of Zhu & Kanamori (2000) (from the corresponding extreme in the H-$\kappa$ diagram; see Fig. 5).
Fig. 6). Additionally, lines of constant delay time, which correspond to the phases Ps (light green), PpPs (green) and PpSs (dark green), have been calculated and included in Figure 6. These model curves have characteristic slopes for each phase type. The curves are adjusted so that they intersect at the maximum of the amplitude stack. In the case of ambiguous maxima we verify that the calculated delay times of Ps, PpPs and PpSs correlate with the delay times of the corresponding phases in the receiver functions to avoid false interpretations.

The thickness of the LVZ ranges between 10 and 15 km. An estimate for the velocity reduction is 1.6 km s$^{-1}$, derived from results in the Kenya Rift (Henry et al. 1990). This leads to overall average crustal P-wave velocities of about 5.5 km s$^{-1}$.

A contribution to the crustal velocity is also provided by the sediment layer thicknesses (Table 2). Among the stations considered in this study, station U09 shows the thickest sedimentary cover (18 km), which can cause a change of 0.01 in the crustal velocity, as described above, not only affect the determination of the Moho depth but also the $v_P/v_S$ ratio. However, the effect on the $v_P/v_S$ ratio is relatively small; a change in velocity of 0.2 km s$^{-1}$ may cause a change of 0.01 in the $v_P/v_S$ ratio, which is well below its uncertainty (see Table 1). Therefore, it is not necessary to calculate the $v_P/v_S$ ratios with the correction factors of $v_P$. We obtain average crustal $v_P/v_S$ ratios from the H-$\kappa$ diagrams, which range between 1.65 and 1.94 (Fig. 7b). The highest $v_P/v_S$ ratios occur in the Edward Rift segment SE of the Rwenzori block with a maximum of 1.94 at station U09. We observe some systematic variations in the spatial distribution of $v_P/v_S$ ratios in the Rwenzori region: the results can be grouped according to the rift shoulders ($v_P/v_S \approx 1.74$), the Albert Rift segment ($v_P/v_S \approx 1.80$), the Edward Rift segment ($v_P/v_S \approx 1.87$), and the southern part of the Rwenzori Mountains ($v_P/v_S \approx 1.86$).

### Discussion

#### Significance of the crustal $v_P/v_S$ ratios

Our results show that the average $v_P/v_S$ ratio of the rift shoulders (c. 1.74) agrees with the value obtained previously for the Rwenzori Belt (Tugume et al. 2012) and also with the average value for the continental crust of c. 1.77 (Christensen 1996). Additionally, our average $v_P/v_S$ ratios correspond to the results from the western Kenyan Rift (Dugda et al. 2005), which exhibits averages of c. 1.73 at the rift shoulders and c. 1.84 in the rift valley.

The Rwenzori Mountains and the rift shoulders are composed of metamorphic basement rocks; this includes the main Gneissic Granulite complex (Archaean unit), as well as the interlayered Buganda–Toro belt (Proterozoic unit) of micaceous schists, amphibolites, quartzites and calcisilicates (Link et al. 2010). Our results show no differences in the $v_P/v_S$ ratios between the Archaean and Proterozoic units. Because of the generally similar lithology, we assume that another effect is responsible for

| Station | $t_S-t_{Sp}$ (s) | $h_1$ (km) | $h_2$ (km) |
|---------|------------------|------------|------------|
| C04     | 0.64 ± 0.05      | 1.93 ± 0.12| 0.78 ± 0.05|
| C09     | 0.77 ± 0.12      | 2.38 ± 0.49| 0.94 ± 0.18|
| U01     | 2.22 ± 0.13      | 6.59 ± 0.48| 2.64 ± 0.19|
| U02     | 1.10 ± 0.18      | 3.37 ± 0.53| 1.34 ± 0.22|
| U03     | 0.86 ± 0.07      | 2.50 ± 0.23| 1.00 ± 0.10|
| U08     | 0.97 ± 0.13      | 2.93 ± 0.41| 1.16 ± 0.16|
| U09     | 0.79 ± 0.04      | 2.53 ± 0.14| 0.99 ± 0.05|
| U14     | 1.51 ± 0.20      | 4.57 ± 0.72| 1.83 ± 0.29|
| U15     | 0.52 ± 0.09      | 1.62 ± 0.29| 0.64 ± 0.11|
| U19     | 1.44 ± 0.10      | 4.48 ± 0.39| 1.76 ± 0.15|

* $t_S-t_{Sp}$: observed travel-time differences of S and Sp phases.

* $h_1$: sediment thicknesses calculated with $v_P = 3.0$ km s$^{-1}$ and $v_S = 1.6$ km s$^{-1}$, derived from results in the Kenya Rift (Kosarev et al. 1993).

* $h_2$: sediment thicknesses calculated with $v_P = 2.9$ km s$^{-1}$ and $v_S = 0.9$ km s$^{-1}$, Kachchh Rift basin, India (Mandal 2007).

### Table 3. Conversion depths and $v_P/v_S$ ratios of the LVZ, which could be determined with the H-$\kappa$ stacking method

| Station | Depths of LVZ (km) | $v_P/v_S$ of LVZ |
|---------|-------------------|-----------------|
| U02     | 20                | 1.93            |
| U03     | 15                | 1.91            |
| U06     | 12                | 1.99            |
| U08     | 18                | 1.99            |
| U09     | 18                | 1.80            |
| U21     | 21                | 1.86            |

Utilize the modified P-wave velocities to obtain the corrected values of the Moho depths with the H-$\kappa$ stacking method, which range between 20 and 39 km (Fig. 7a). These values are in accordance with the Moho depths that are determined with the correction factors of $v_P$ (see Table 1). The Moho depths for the two rift segments exhibit a more uniform distribution in comparison to the initial values (Fig. 5a). We obtain average Moho depths of 26 km in the Albertine Rift and 34 km beneath the rift shoulders, which is in general agreement with previous studies (Wölbern et al. 2010; Tugume et al. 2012).
the larger $v_p/v_s$ ratios of c. 1.86 for the horst structure in comparison to the rift shoulders (c. 1.74).

The Albertine Rift is an active seismic and geothermal region including volcanic deposits, the Buranga hot springs and the volcanic fields of Katwe-Kikorongo and Bunyaruguru (Bahati et al. 2010). According to local travel-time tomographic studies (Jakovlev et al. 2011, 2013), there are indications for active magmatic processes below the Rwenzori Mountains and the Sempaya hot springs, as indicated by decreasing seismic velocities. In our study we inferred an LVZ in the crust, which is also an indication for magmatic activity within the Edward Rift segment. Therefore, we associate the observed significantly higher $v_p/v_s$ values with partial melts in the crust.

**Stretching factors as indicators for different styles of rifting**

To determine the actual state of rifting and to understand the rifting mechanism, we calculate the stretching factors as described by Royden & Keen (1980). Required input parameters to determine the stretching factors of the crust (denoted $\delta$) and the sub-crustal lithosphere (denoted $\beta$) are the initial thicknesses before rifting and the final thicknesses in the current condition. As a first approximation we may obtain the initial and final thicknesses from the average Moho depths beneath the rift shoulders and the Albertine Rift of 34 and 26 km, respectively. Values for the average thickness of the lithosphere beneath the Albertine Rift (c. 120 km) and the rift

**Fig. 6.** H-$k$ stacking results for stations (a) U03 and (b) U08. The derived values for Moho depth and $v_p/v_s$ ratio are marked by a green cross. The superimposed green lines represent delay times related to the Moho conversions $P_s$ (light green), $PpPs$ (green) and $PpSs$ (dark green). In addition to the Moho results, two additional extreme values of the H-$k$ stacking are also shown, corresponding to the top (marked in yellow) and the bottom (marked in blue) of a low velocity zone in the lower crust.
shoulders (c. 190 km) are taken from Wölbern et al. (2012). We obtain relatively similar values for the stretching factors, \( \delta = 0.8 \) and \( \beta = 1.8 \). This similarity of the stretching factors and the relatively late onset of volcanism (about 70 ka ago as described by Link et al. 2010) in comparison to the initial stages of rifting (at about 8 Ma as described by Ring et al. (1992), Ebinger et al. (1993) and Ring & Betzler (1995)) indicates that the Rwenzori region of the Albertine Rift can be described by a passive rift model (Huismans et al. 2001 and references therein).

We utilize the stretching factors to calculate the subsidence of the Albertine Rift, triggered by different extensional stresses (e.g. McKenzie 1978; Royden & Keen 1980) relative to the decoupling horizon of the crust. We assume that the depth of the decoupling horizon corresponds to the observed decrease of seismicity at the brittle–ductile transition zone. In the Rwenzori region this decrease occurs at a depth of about 22 km (Tugume & Nyblade 2009; Lindenfeld et al. 2012). We further assume: the density of the mantle \( \rho_m = 3.3 \text{ g cm}^{-3} \) the mean density of the crust \( \rho_c = 2.9 \text{ g cm}^{-3} \) and the thermal expansion coefficient \( \alpha = 3.2 \times 10^{-5} \text{ C}^{-1} \). For the fraction of dykes we take \( \gamma = 0.3 \) in the lithosphere (estimated after Wölbern et al. 2012) and for the temperature at the bottom of the lithosphere \( T_1 = 1420 \text{ C} \). The parameter \( T_1 \) is derived from the upper mantle temperature beneath the Kivu and Virguna Rift in the western branch of the EARS (Rooney et al. 2012). Application of the method of Royden & Keen (1980) leads to a subsidence value of about c. 1.3 km within the Albertine Rift.

The sediment layer thicknesses in the Albertine Rift have an average value of about 3.3 km (see Table 2), that is the calculated subsidence is insufficient to explain the current surface topography of the rift with a decrease in elevation of about 0.5 km in comparison to the rift shoulders. The elevation changes due to the rifting processes generally depend on two opposing effects, the uplift by lithospheric heating and the subsidence by crustal thinning (Steckler 1985). Our observations do not support a significant effect due to lithospheric heating. We therefore assume that the evolution of the Albertine Rift is based primarily on passive rifting processes. In response to extensional stresses, the development of the rift is controlled by the viscosity and the elasticity of the crust as described by Sachau et al. (2013). Their model also predicts subsidence values between 2.5 and 4 km, which are consistent with our results for the thickness of the sedimentary layer. However, their modelled elevations of the Rwenzori Mountains are not consistent with the observed topography. Additionally, the model of Sachau et al. (2013) requires a contrast of viscosity at the brittle–ductile transition with a low-viscous layer underneath. The reduced viscosity can be explained possibly by the effects of lithospheric delamination (Wallner & Schmeling 2010). In our view it seems likely that a combination of the extensional model of Sachau et al. (2013) and the rift-induced lithospheric delamination proposed by Wallner & Schmeling (2010)
Conclusions

Our conclusions can be summarized as follows:

(1) We have calculated receiver functions of \( P \) and \( PP \) phases to map the Moho topography in the Rwenzori regions with high resolution. Crustal thicknesses and \( v_P/v_S \) ratios have been derived using the \( H-K \) stacking method of Zhu & Kanamori (2000) covering the central rift area, the Rwenzori range, as well as the western and eastern rift shoulders.

(2) From S-to-P conversions in local earthquake records, we have derived the thickness of sedimentary layers beneath 10 seismic stations, which are located in the rift valley. Assuming realistic seismic velocities, we find sediment layers of up to 6.6 km thickness in the Albertine Rift segment (station U01) and up to 3.4 km in the Edward Rift segment (station U02). These results agree well with sediment thicknesses that were derived in the Lake Albert area (e.g. Karner et al. 2000).

(3) The \( H-K \) stacking algorithm yields stable Moho depth estimates with an estimated error of \( \pm 2.0 \) km. All Moho depths are calculated with respect to an average elevation level of 1.3 km in order to compensate for the surface topography and to reveal the true Moho topography. We further improved the results by removing the effect of the detected sediment layers from the calculated Moho depths.

(4) We determined distinctly different \( v_P/v_S \) ratios for the rift shoulders (\( v_P/v_S = 1.74 \)), the Albert Rift segment (\( v_P/v_S = 1.80 \)), the Edward Rift segment (\( v_P/v_S = 1.87 \)) and the Rwenzori Mountains (\( v_P/v_S = 1.86 \)). In view of previous studies, we associate the corresponding large values at the Rwenzori Mountains with decompressional melting (Schmeling & Wallner 2012) or ascending dykes of asthenospheric material. The high \( v_P/v_S \) ratio of the Edward Rift segment is probably related to a similar process.

(5) We mapped a pronounced LVZ in the middle crust, extending from Lake Edward to the eastern rift shoulder (at station U06). The receiver function analysis of phases related to this discontinuity yields significantly higher \( v_P/v_S \) ratios (up to 1.99) in comparison to the analysis of Moho-related phases. We assume that the low-velocity layer is due to partial melting within the crust. This anomaly was also discussed by Wölbern et al. (2010). However, in our study its approximate extent is better constrained than in the previous study.

(6) The receiver functions for stations on the eastern and western rift shoulders yield more stable results than for stations within the rift valley, probably because the crustal structure of the shoulders is less complex. Moho depths below the shoulders are significantly larger than in the rift valley. On the western rift shoulder Moho depths range from 29 to 39 km, and on the eastern shoulder the values range from 30 to 39 km. The average Moho depth of c. 34 km agrees with previous studies (Wölbern et al. 2010; Tugume et al. 2012) within \( \pm 2.0 \) km.

(7) Within the rift we observe a relatively uniform Moho depth. The derived depth varies between 25 and 31 km in the Edward Rift segment and from 22 to 30 km in the Albert Rift segment. The crustal thickness exhibits an average value of 26 km beneath the rift segment, which is similar to the situation beneath the southern Rwenzori Mountains. Notably, there is no indication for a crustal root beneath the Rwenzori Mountains. For stations U04 and U22 at the Rwenzori horst structure, we reveal even more reduced Moho depths compared to adjacent stations at the Rwenzori Mountains and in the surrounding rift segments. However, such variations also occur at individual stations that exhibit significant azimuthal variations indicative of small-scale crustal anomalies.

(8) Our observations of crustal thinning and the great depths of the rift basins, removing the sediment layer thicknesses, are in support of Sachau et al. (2013), who argue that the uplift of the Rwenzori Mountains is caused by the mechanism of crustal bending triggered by differences of lateral tensile stress in the crust. We further assume that the processes of passive rifting may be supported by the effects of lithospheric delamination (Wallner & Schmeling 2010) in order to explain the extreme uplift of the Rwenzori structure.

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