Crustal structure of the Khartoum Basin, Sudan

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A B S T R A C T

The crustal structure of the northern part of the Khartoum Basin has been investigated using data from 3 permanent seismic stations within 40 km of Khartoum and two modeling methods, H–k stacking of receiver functions and a joint inversion of receiver functions and Rayleigh wave group velocities. The Khartoum Basin is one of several Mesozoic rift basins in Sudan associated with the Central African Rift System. Results from the H–k-stacking indicate that crustal thickness beneath the Khartoum Basin ranges between 33 and 37 km, with an average of 35 km, and that the crustal Vp/Vs ratio ranges from 1.74 to 1.81, with an average of 1.78. From the joint inversion of receiver functions and Rayleigh wave group velocities, we obtained similar results for Moho depth, as well as an average shear wave velocity of 3.7 km/s for the crust. These results provide the first seismic estimates of Moho depth for a basin in Sudan. When compared to average crustal thickness for unfractured Proterozoic crust in eastern Africa, our results indicate that at most only a few km of crustal thinning may have occurred beneath the Khartoum Basin. This finding is consistent with estimates of effective elastic plate thickness, which indicate little modification of the Proterozoic lithosphere beneath the basin, and suggests that there may be insufficient topography on the lithosphere–asthenosphere boundary beneath the Sudanese basins to channel plume material westward from Ethiopia.

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1. Introduction

The Khartoum Basin is one of several Mesozoic rift basins in Sudan associated with the Central African Rift System (CARS) (Fig. 1). The initial development of the Sudanese basins has been linked to extension throughout the Afro-Arabian region associated with the separation of east and west Gondwana and sea-floor spreading of the proto-Indian Ocean in the Jurassic (Reynolds, 1993). A second phase of rift development occurred in the Cretaceous during the opening of the South Atlantic (Browne et al., 1985; Fairhead and Green, 1989) and faulting along the Central African Shear Zone (Fig. 1) (Bosworth, 1989; Browne and Fairhead, 1983; Browne et al., 1985; Fairhead, 1986).

The structure of the Sudan rift basins, which align in a northwest to southeast direction (Fig. 1), has been investigated previously using potential field data, seismic reflection data and well logs (e.g., Hussein, 1992; Mann, 1989; Schull, 1988). The basins are bounded by listric normal faults (Mann, 1989; Reynolds, 1993), and the amount of extension is variable among them. In the Muglad Rift, for example, about 45 km of crustal extension is documented (Browne and Fairhead, 1983; Jorgensen and Bosworth, 1989), while in the White Nile rift basin about 10 km of extension has occurred (Brown et al., 1980). Similarly, sediment thickness also varies between the basins, with up to 15 km of fill in the Muglad basin (Schull, 1988) and possibly as little as 1 km in the Blue Nile rift (Jorgensen and Bosworth, 1989).

Little is known about the deep structure under the basins, in particular the depth of the Moho, limiting our understanding of their tectonic development. The only information on deep crustal structure comes from regional gravity studies (Bermingham et al., 1983; Browne and Fairhead, 1983; Browne et al., 1985; Jorgensen and Bosworth, 1989) and continental scale tomography models (e.g., Pasyanos and Nyblade, 2007), but these studies crustal thickness is not well constrained. In this paper, we present the first estimates of Moho depth for a Mesozoic rift basin in Sudan determined using receiver functions and Rayleigh wave group velocities, along with estimates of crustal Poisson’s ratio and the shear wave velocity structure of the crust. We then comment on the tectonic implications of our findings for lithospheric extension within the CARS, as well as the development of Cenozoic volcanic fields in central Africa vis-à-vis the channeling of plume material from Ethiopia.

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2. Background

2.1. Geology and tectonic setting of the Khartoum Basin

The Khartoum Basin is part of the Blue Nile rift basin system, which lies in the north between the White Nile and the Blue Nile, and stretches to the southeast across the Blue Nile (Fig. 1) (Awad, 1994). The basin may be a ‘sag’ basin because its edges do not appear to be controlled by major bounding faults, and it may be quite shallow, with a maximum sediment thickness of not more than a few kilometers (Mohammed, 1997).

The basin sediments lie uncomformably on Pan-African (Neo-proterozoic) metamorphic and igneous rocks of the Mozambique Belt, which are exposed north of Khartoum in the Sabaloka area (Fig. 2). Kröner et al. (1987) suggested that the gneisses found there formed along an ancient continental margin onto which juvenile arc assemblages were accreted. The Albian age Omdurman formation crops out across much of the study area (Awad, 1994; Schrank and Awad, 1990) (Fig. 2), and consists of the Umm Badda and Merkhiyat members (Awad, 1994). Within these members, the dominant lithology is the Nubian sandstone, which is composed of sandstones, pebbly sandstones, and conglomerates, often capped by ferruginous beds (Whiteman, 1971). The late Tertiary and Quaternary Gezeria formation crops out in the area between the White Nile and the Blue Nile and on the east bank of the Blue Nile (Awad, 1994) (Fig. 2). The formation consists of sandstones, silts and clays (Adamson and Williams, 1980; Andrew, 1948). Quaternary deposits in the area include the Nile Silt and Wadi deposits (Fig. 2).

2.2. Seismic stations

Data used for this study come from the Sudan Seismic Network (SSN), which consists of 3 seismic stations, MRKH, SLAT and JAWL, installed in 2001 within 40 km of Khartoum (Al Hassan et al., 2007) (Fig. 2). The stations are equipped with intermediate period LE-3D/20 sensors, which have a flat frequency response from 0.05 to 20 Hz, Mars-88 data loggers with 24-bit digitizers, GPS clocks for timing and bidirectional RF telemetry. Until recently, data were recorded in trigger mode. The MRKH station is located on an outcrop of Nubian sandstone at Jebel Merkhiyat. The JAWL station is located on an outcrop of Nubian sandstone at Jebel Aulia, and the SLAT station is located on a ridge of Nubian sandstone to the north-east of Khartoum (Fig. 2).

3. Receiver functions

Receiver functions are time series obtained from teleseismic P-waveforms recorded at single seismic stations after deconvolving the vertical component of ground motion from the corresponding horizontal components. The deconvolution isolates the seismic response of the local structure beneath the recording station and removes the signature of the instrument response and source time function (Langston, 1979). Modeling receiver functions has become a standard procedure for investigating Moho depth and the shear wave velocity structure of the crust (e.g. Ammon et al., 1990; Julià et al., 2000; Owens and Zandt, 1997).

For this study, we have used data from seismic sources located at epicentral distances between 30° and 90° and with moment magnitude greater than 5.5 (see Table 1 and Fig. 3) recorded between 2003 and 2010. Before computing the receiver functions, the data were windowed 10 s before and 40 s after the P wave. The data were then de-trended, tapered and band pass filtered between 0.05 and 8 Hz to remove low-frequency noise and avoid aliasing before being decimated to 10 samples per second. The horizontal seismograms were rotated into the great circle path to obtain the corresponding radial and transverse components, and the iterative time-domain deconvolution method of...
Ligorria and Ammon (1999) was employed to compute the receiver functions using 500 iterations. For each teleseismic event, receiver functions were computed at two overlapping frequency bands, one with a Gaussian filter width of 1.0 (f < 0.5 Hz) and the other with a filter width of 2.5 (f < 1.2 Hz).

To evaluate the quality of the deconvolved traces, a least-squares-misfit criterion was utilized, in which the radial waveform for each receiver function is reconstructed by convolving the receiver function back with the corresponding vertical waveform and comparing it with the original radial waveform. Only receiver functions that were recovered at an 85% level or higher were further considered for analysis (Ligorria and Ammon, 1999).

The degree of lateral heterogeneity and isotropy of the propagation medium can be assessed by examining transverse receiver function amplitudes (Cassidy, 1992). Small amplitudes on the transverse component indicate a homogeneous and isotropic medium beneath the station, which is assumed in both the H–k stacking and joint inversion modeling procedures. Therefore, in cases where we obtained transverse receiver functions with amplitudes similar to the radial receiver function, the data were rejected.

3.1. H–k stacking technique

Receiver functions have been modeled using the H–k stacking technique of Zhu and Kanamori (2000) to estimate crustal thickness (H) and crustal Vp/Vs ratio (k). The stacking procedure transforms the receiver function waveforms from the time-amplitude domain into the H–k parameter space through a weighted sum along theoretical phase moveout curves obtained for given values of H and k and an assumed value of P-wave velocity. The weighted sum is given by

\[ s(H,k) = \sum_{j=1}^{N} w_1 r_j(t_1) + w_2 r_j(t_2) - w_3 r_j(t_3) \]

where \( w_1, w_2 \) and \( w_3 \) are the a priori weights for the Ps, PpPs, and PpSs phases, respectively, with summation equal to 1. \( r_j(t_i), i = 1, 2, 3 \) are

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**Fig. 2.** Geological and tectonic map of the Khartoum area with seismic stations shown by black triangles. Map redrawn from Awad (1994).
The basic idea behind the H–k algorithm is that the stacked amplitudes will attain its maximum value when there is coherent stacking along the phase moveout curves resulting from H and k values consistent with the location of a subsurface discontinuity (Zhu and Kanamori, 2000). Moreover, if lateral variations in Earth structure are moderate, by stacking receiver functions from different backazimuths and different distances, the effects of lateral structural variation are reduced and an average crustal thickness can be obtained.

In applying the H–k technique, it is necessary to select weights w1, w2, and w3 (Eq. (1)). Although the choice for weights is somewhat subjective, we give more weight to the phases that are most clearly observed and less weight to phases that are not clearly identified or inconsistent among receiver functions. For stations SLAT and MRKH, for instance, all three phases are clearly seen in the receiver function waveforms, so we give similar weight to each of them by using w1 = 0.4, w2 = 0.3 and w3 = 0.3. On the other hand, for JAWL the PsPs + PpSs multiple is not as clear as the other two phases, and so the weights are re-adjusted to w1 = 0.5, w2 = 0.1 and w3 = 0.4.

In the stacking procedure, a mean Vp for the crust must be assumed. An average crustal P-wave velocity of 6.5 km/s was used, which is representative of Precambrian crust worldwide (e.g., Christensen and Mooney, 1995). However, as explained below, other values of Vp were investigated in order to determine uncertainties in H and k due to this assumption.

Following the approach of Julià and Mejía (2004), we estimate formal uncertainties in H and k by bootstrapping the receiver function dataset for each station using 200 replications (Efron and Tibshirani, 1991). The uncertainties for each station using a mean crustal Vp of 6.5 km/s for the H–k stacking are given in Table 2 and shown in Fig. 4. For evaluating the uncertainty in H and k arising from the choice of mean crustal Vp, the H–k stacks were recomputed using a range of Vp between 6.3 and 6.8 km/s (Table 2; and also shown in the Supplementary material). To obtain an overall uncertainty in H and k, the formal uncertainty obtained from the bootstrap method was combined with the range of H and k values obtained when using the range of mean crustal Vp values. The overall uncertainties are ±3–4 km in Moho depth and ±0.05 in Vp/Vs for each station.

Fig. 4 shows the H–k results for each station and Table 2 provides a summary of the results. Crustal thicknesses are 35, 37 and 33 km for stations SLAT, MRKH and JAWL respectively, and the average is ±2 km. The overall uncertainties in H and k are ±3–4 km in Moho depth and ±0.05 in Vp/Vs for each station.

Table 1
Events with magnitude Mb > 5.5 recorded on the Sudan Seismic Network (SSN) used in this study.

<table>
<thead>
<tr>
<th>Event date (yyyy/mm/dd)</th>
<th>Event time (h.min.s)</th>
<th>Latitude (deg)</th>
<th>Longitude (deg.)</th>
<th>Magnitude (Mw)</th>
<th>Ray parameter (s km(^{-1}))</th>
<th>Backazimuth (deg)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2003/12/10</td>
<td>04:38:11.6</td>
<td>23.03</td>
<td>121.362</td>
<td>6.8</td>
<td>0.046</td>
<td>68.08</td>
</tr>
<tr>
<td>2004/10/8</td>
<td>14:36:6.10</td>
<td>13.925</td>
<td>120.534</td>
<td>6.4</td>
<td>0.045</td>
<td>77.11</td>
</tr>
<tr>
<td>2004/11/17</td>
<td>20:58:22.3</td>
<td>39.189</td>
<td>71.857</td>
<td>5.7</td>
<td>0.073</td>
<td>47.92</td>
</tr>
<tr>
<td>2004/12/26</td>
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<td>3.295</td>
<td>95.982</td>
<td>9.0</td>
<td>0.059</td>
<td>94.11</td>
</tr>
<tr>
<td>2005/5/14</td>
<td>05:05:18.5</td>
<td>0.587</td>
<td>98.459</td>
<td>6.7</td>
<td>0.057</td>
<td>96.03</td>
</tr>
<tr>
<td>2006/1/2</td>
<td>06:10:54.8</td>
<td>−60.934</td>
<td>−21.575</td>
<td>7.4</td>
<td>0.043</td>
<td>203.27</td>
</tr>
<tr>
<td>2007/7/25</td>
<td>23:37:31.5</td>
<td>7.157</td>
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<td>5.5</td>
<td>0.062</td>
<td>90.66</td>
</tr>
<tr>
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<td>24.340</td>
<td>122.219</td>
<td>6.1</td>
<td>0.046</td>
<td>66.69</td>
</tr>
<tr>
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<td>−56.02</td>
<td>−28.03</td>
<td>6.0</td>
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<td>−34.90</td>
<td>6.5</td>
<td>0.057</td>
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</tr>
<tr>
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<td>349.26</td>
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<td>7.5</td>
<td>0.056</td>
<td>97.09</td>
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<tr>
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<td>−0.52</td>
<td>101.50</td>
<td>6.6</td>
<td>0.054</td>
<td>98.39</td>
</tr>
<tr>
<td>2009/10/3</td>
<td>17:36:06.1</td>
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<td>0.046</td>
<td>67.49</td>
</tr>
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<td>70.72</td>
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<td>51.46</td>
</tr>
<tr>
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<td>68.17</td>
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<tr>
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<td>91.90</td>
<td>6.0</td>
<td>0.062</td>
<td>90.03</td>
</tr>
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<td>0.046</td>
<td>203.00</td>
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<td>130.70</td>
<td>6.9</td>
<td>0.043</td>
<td>47.13</td>
</tr>
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<td>122.40</td>
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</tr>
<tr>
<td>2010/3/3</td>
<td>00:18:51.2</td>
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<td>120.79</td>
<td>6.3</td>
<td>0.047</td>
<td>68.42</td>
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<td>0.050</td>
<td>99.96</td>
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<td>−2.75</td>
<td>83.68</td>
<td>5.8</td>
<td>0.065</td>
<td>105.58</td>
</tr>
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<td>16:54:46.7</td>
<td>13.67</td>
<td>92.83</td>
<td>6.7</td>
<td>0.062</td>
<td>83.43</td>
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<td>2.38</td>
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<td>6.6</td>
<td>0.054</td>
<td>100.05</td>
</tr>
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<td>096.03</td>
<td>7.2</td>
<td>0.059</td>
<td>93.61</td>
</tr>
</tbody>
</table>
Table 2
Crustal thickness and Vp/Vs ratios for the Khartoum Basin.

<table>
<thead>
<tr>
<th>Station</th>
<th>Weight</th>
<th>H (1) (km)</th>
<th>k (1)</th>
<th>H (2) (km)</th>
<th>k (2)</th>
<th>H (3) (km)</th>
<th>k (3)</th>
<th>HJ (km)</th>
<th>VsJ (km/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>SLAT</td>
<td>w₁ = 0.4, w₂ = 0.3, w₃ = 0.3</td>
<td>345 ± 1.1</td>
<td>1.81 ± 0.05</td>
<td>36.4</td>
<td>1.80</td>
<td>33.2</td>
<td>1.82</td>
<td>38 ± 2.5</td>
<td>3.7</td>
</tr>
<tr>
<td>MRKH</td>
<td>w₁ = 0.4, w₂ = 0.3, w₃ = 0.3</td>
<td>36.9 ± 1.2</td>
<td>1.74 ± 0.05</td>
<td>38.7</td>
<td>1.74</td>
<td>35.6</td>
<td>1.75</td>
<td>38 ± 2.5</td>
<td>3.7</td>
</tr>
<tr>
<td>JAWL</td>
<td>w₁ = 0.5, w₂ = 0.1, w₃ = 0.4</td>
<td>32.7 ± 1.4</td>
<td>1.81 ± 0.07</td>
<td>37.4</td>
<td>1.81</td>
<td>31.6</td>
<td>1.82</td>
<td>38 ± 2.5</td>
<td>3.6</td>
</tr>
<tr>
<td>Average</td>
<td></td>
<td>35</td>
<td>1.78</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>38</td>
<td>3.7</td>
</tr>
</tbody>
</table>

H₁, H₂, H₃, k₁, k₂, k₃: average of Vp used for H–k stack method. (1) Vp = 6.5 km/s, (2) Vp = 6.8 km/s and (3) Vp = 6.3 km/s.
HJ = depth from joint inversion method.
VsJ = shear velocity from joint inversion method.

35 km. Vp/Vs ratios are 1.81, 1.74 and 1.81 for stations SLAT, JAWL and MRKH respectively, with an average of 1.78.

3.2. Joint inversion technique

To investigate the shear wave velocity structure of the crust, we have used the method of Julià et al. (2000, 2003) to jointly invert the receiver functions with Rayleigh-wave group velocities. The combination of surface-wave dispersion curves and receiver functions provides better constraints on the shear velocity of the crust compared to modeling either data set independently (Dugda et al., 2007; Julià et al., 2000, 2003, 2005; Ozalaybey et al., 1997). This method makes use of a linearized inversion procedure that minimizes a weighted combination of least squares norms for each data set (i.e., receiver functions and group velocities). To make the contribution of each data set to the joint least squares misfit comparable, a normalization of the data sets is necessary, and this is done using the number of data points and variance for each of the data sets. An influence factor is then used to control the trade-off between fitting the receiver functions and the group velocity dispersion curves. Because the velocity models are parameterized as a stack of thin layers, a smoothness constraint is necessary to stabilize the inversion process. The velocity models obtained are, therefore, a compromise between fitting the observations, model simplicity and a priori constraints (Julià et al., 2003).

Dispersion curves for each station were obtained from a revised version of the Rayleigh wave group velocity model published by Pasyanos and Nyblade (2007). Rayleigh wave group velocities between 20 and 75 s were used. As the original dispersion curve was rough due to limited path coverage across Sudan, a 3-point running average was applied to the dispersion curve to smooth it out before using it in the inversion. Dispersion velocities in the 20–50 s period range constrain crustal and uppermost mantle velocity structure; however, dispersion velocities at periods up to 75 s were included to model velocity structure down to ~100 km depth during the inversion.

The starting model used in the joint inversion consisted of a 37.5 km thick crust with a linear shear wave velocity increase from 3.4 at the surface to 4.0 km/s at the Moho, and a flattened PREM (Preliminary Reference Earth Model; Dziewonski and Anderson, 1981) model for the mantle. The model parameterization consists of a stack of constant velocity layers that increase in thickness with depth. Thicknesses at the top of the model were 1 and 2 km, respectively, 2.5 km for layers between 3 and 60.5 km depth, and 5 km between 60.5 and 260.5 km depth. Poisson’s ratio was fixed to 0.25 down to 38 km depth during the inversions and to PREM values for deeper structure. Densities were obtained from P wave velocities using the empirical relation of Berteussen (1977). Velocities below a depth of 100 km were fixed to PREM values to account for the partial sensitivity of the longer periods to deeper structure.

In order to investigate lateral variations in crustal structure around the stations, receiver functions were first sorted by backazimuth and ray parameter and then averaged within groups. Grouping by backazimuth was done by overlapping individual receiver function waveforms and visually correlating waveform shapes, while groupings by ray parameter were done by limiting ray parameter values to a maximum variation of 0.01 s km⁻¹ within the backazimuth groups. Four or five groups of receiver functions were obtained for each station. For each group, receiver functions with both lower (f < 0.5 Hz) and higher (f < 1.25 Hz) frequencies were computed and stacked with a minimum of two receiver functions per stack. Each receiver function group for each station was then jointly inverted with the corresponding group velocity curve to obtain a shear wave velocity model. The results (Fig. 5) show that comparable models with similarly good fits to the data are obtained for each group and for each station, indicating little variability in crustal structure around the stations. Therefore, a best average velocity model for each station was obtained by jointly inverting all receiver function averages for a given station with the group velocity curve.

Following the approach of Julià et al. (2005), uncertainties in the shear wave models have been assessed by repeatedly performing the inversions with a range of inversion parameters, such as weighting parameters and Poisson’s ratios. We obtain an uncertainty of approximately 0.1–0.2 km/s for the velocity in each layer, which translates into an uncertainty of ~2–3 km in the depth of any crustal boundary observed in the model, including the Moho.

4. Discussion and conclusions

Following the approach of Kgaswane et al. (2009), Tokam et al. (2010), and Tugume et al. (2012), crustal thickness is determined from the shear wave velocity models in Fig. 5 by placing the Moho at the depth where the shear wave velocity exceeds 4.3 km/s. Shear wave velocities for typical lower crust lithologies are not higher than 4.3 km/s (Christensen, 1996; Christensen and Mooney, 1995). Hence, we take shear wave velocities above 4.3 km/s to indicate the presence of lithologies with mantle composition. At station SLAT, a marked velocity increase occurs at 38 km depth, where velocities increase to values > 4.3 km/s (Fig. 5). However, at stations MRKH and JAWL, the velocity increase with depth is gradational and a sharp velocity discontinuity is not imaged (Fig. 5).

The crustal thickness beneath all three stations determined from the station average models is 38 km (Fig. 5 and Table 2), and the average shear wave velocity for stations SLAT, MRKH and JAWL is 3.6, 3.7 and 3.7 km/s, respectively. The Moho depths determined from the joint inversion models are uniformly deeper than those obtained from the H–k stacking procedure (Table 2), but still comparable within the reported uncertainties (i.e., 3 to 4 km for the H–k stacking estimates and ±2.5 km for the joint inversion estimates).

To summarize, our results indicate that crustal thickness beneath the Khartoum Basin is between 33 and 38 km thick, with an average of 35 km, the crustal Vp/Vs ratio is on average 1.78 (Poisson’s ratio of 0.27), and the average Vs of the crust is 3.7 km/s. In the remainder of this section, we compare these results to other estimates of crustal structure in Sudan and then investigate their implications for lithospheric thinning and the channeling of plume material.

A number of gravity studies have reported estimates of Moho depth in Sudan that are in agreement with our results. Browne et al. (1985) and Mohamed et al. (2001) estimated a crustal thickness of about 35 km beneath much of Sudan, Jorgensen and Bosworth (1989) reported Moho depths of 32 to 37 km for central Sudan, and...
Fig. 4. H–k stack of receiver functions for stations (A) SLAT, (B) MRKH and (C) JAWL respectively, for a mean crustal Vp of 6.5 km/s. To the left of each receiver function, the top number gives the event azimuth and the bottom number gives the event distance in degrees. Contours map out percentage values of the objective function given in the text.

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a crustal thickness map in Tedla et al. (2011) shows Moho depths of 33 to 36 km for central Sudan.

As discussed previously, the Pan-African basement beneath the Khartoum Basin belongs to the Mozambique Belt. To estimate the amount of crustal thinning beneath the Khartoum Basin, therefore, we compare our results to estimates of Moho depth in areas of east Africa where the Proterozoic Mozambique Belt has not been rifted. The crustal structure of Mozambique Belt in Kenya has been studied by the Kenya Rift International Seismic project (KRISP) using seismic refraction profiling (Fuchs et al., 1997; Prodehl et al., 1994 and references therein), and by Dugda et al. (2005) using receiver functions. Results from these studies show that crustal thickness is 40 to 42 km away from the Kenya rift. The Mozambique Belt in Tanzania has been also investigated using receiver functions and Rayleigh wave group and phase velocities (Dugda et al., 2005; Julià et al., 2005; Last et al., 1997). Moho depths were found to range from 37 to 42 km. If these results are averaged, then a mean crustal thickness of 38 ± 2 km is obtained (Tugume et al., 2012). If we compare our estimated average crustal thickness of 35 km for the Khartoum Basin with the average crustal thickness from these studies, then at most there has been only a few km of crustal thinning. This small amount of crustal thinning is similar to the amount of thinning in the southern part of the Kenya rift (Birt et al., 1997), but is significantly less than that in northern Kenya where the Cenozoic rift and the Mesozoic Anza graben coincide (Fig. 1). In this region, the 20–25 km thick crust (Benoit et al., 2006) is probably the result of two episodes of thinning, one during Mesozoic rifting and then again in the Cenozoic.

Given the similarity between our crustal thickness estimates for the Khartoum Basin and estimates of crustal thickness elsewhere in central and southern Sudan from gravity models, it follows that only small amounts of crustal thinning have occurred throughout the CARS in Sudan. And similarly, it can be inferred that the lithosphere has also undergone only minor amounts of extension, if any. This conclusion is supported by estimates of effective elastic plate thickness (Te). Pérez-Gussinye et al. (2009) obtained Te estimates of 80 to 100 km for the region beneath the Khartoum Basin, as well as for several of the other basins in central and southern Sudan, including the Muglad basin, the largest one. A Te of 80 to 100 km is similar to or greater than estimates for the Mozambique Belt in parts of eastern Africa where the lithosphere has not been affected by the Cenozoic East African rift system. Pérez-Gussinye et al. (2009) show a narrow region within central Sudan where Te is between 50 and 60 km, but this region does not encompass any of the larger rift basins. Thus, the Te estimates for the basins are consistent with our conclusion that there has been little thinning of the crust or lithosphere beneath them.

**Fig. 5.** Joint inversion results for stations (A) SLAT, (B) MRKH and (C) JAWL. The top, middle and bottom panels in each column display receiver functions, Rayleigh wave group velocities and shear wave velocity models, respectively. For the receiver functions, observations are shown in black and predictions in red. Above each group of receiver functions, the number of waveforms (N), average backazimuth (degrees) ± one standard deviation (baz ± sbaz) and average ray parameter (s km$^{-1}$) ± one standard deviation is given in sequence. For the group velocity plots, the observations are shown with triangles and the model results are shown with the solid red lines. For the velocity models, the grey line is the starting model and the red line is the model obtained from the inversion. The panel to the right shows the shear wave velocity model obtained by jointly inverting all groups of receiver functions with the group velocity curve. The dashed line shows the Moho depth. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

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The conclusion that the crust and lithosphere under the Sudan basins have not been thinned to any significant extent has important implications for understanding Cenozoic volcanism in central Africa. Ebinger and Sleep (1998) and P rez-Gussinye et al. (2009) have proposed that thinned lithosphere beneath the Sudan basins created a channel in the lithosphere–asthenosphere boundary for magma from the Afar/Ethiopia region to migrate to the west, reaching as far as the Cameroon Volcanic Line. Our results showing little crustal thinning combined with the Te estimates from P rez-Gussinye et al. (2009) indicate that there may be insufficient topography on the lithosphere–asthenosphere boundary, at least beneath some of the basins, to create a channel for the migration of plume material. This finding raises the possibility that the Cenozoic volcanism in regions like the Darfur and the Cameroon Volcanic Line might not have resulted from plume material migrating from the Ethiopia/Afar region to the west beneath thinned lithosphere.

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