Structure of the crust beneath Cameroon, West Africa, from the joint inversion of Rayleigh wave group velocities and receiver functions

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SUMMARY
The Cameroon Volcanic Line (CVL) consists of a linear chain of Tertiary to Recent, generally alkaline, volcanoes that do not exhibit an age progression. Here we study crustal structure beneath the CVL and adjacent regions in Cameroon using 1-D shear wave velocity models obtained from the joint inversion of Rayleigh wave group velocities and P-receiver functions for 32 broad-band seismic stations deployed between 2005 January and 2007 February. We find that (1) crustal thickness (35–39 km) and velocity structure is similar beneath the CVL and the Pan African Oubanguides Belt to the south of the CVL, (2) crust is thicker (43–48 km) under the northern margin of the Congo Craton and is characterized by shear wave velocities ≥ 4.0 km s\(^{-1}\) in its lower part and (3) crust is thinner (26–31 km) under the Garoua rift and the coastal plain. In addition, a fast velocity layer (\(V_s\) of 3.6–3.8 km s\(^{-1}\)) in the upper crust is found beneath many of the seismic stations. Crustal structure beneath the CVL and the Oubanguides Belt is very similar to Pan African crustal structure in the Mozambique Belt, and therefore it appears not to have been modified significantly by the magmatic activity associated with the CVL. The crust beneath the coastal plain was probably thinned during the opening of the southern Atlantic Ocean, while the crust beneath the Garoua rift was likely thinned during the formation of the Benue Trough in the early Cretaceous. We suggest that the thickened crust and the thick mafic lower crustal layer beneath the northern margin of the Congo Craton may be relic features from a continent–continent collision along this margin during the formation of Gondwana.

Key words: Inverse theory; Body waves; Surface waves and free oscillations; Cratons; Crustal structure.

1 INTRODUCTION
The Cameroon Volcanic Line (CVL) is a major geological feature that cuts across Cameroon from the southwest to the northeast. It is a unique volcanic lineament which has both an oceanic and a continental sector and consists of a chain of Tertiary to Recent, generally alkaline, volcanoes stretching from the Atlantic island of Pagalu to the interior of the African continent (Fig. 1, Fitton 1987; Lee et al. 1994). The oceanic sector includes the islands of Bioko (formerly Fernando Po) and São Tomé and Príncipe while the continental sector includes the Etinde, Cameroon, Manengouba, Bamboutos, Oku and Mandara mountains, as well as the Adamawa and Biu Plateaus.

In addition to the CVL, three other major tectonic features characterize the region: the Benue Trough located northwest of the CVL, the Central African Shear Zone (CASZ), trending N70°E, roughly parallel to the CVL and the Congo Craton in southern Cameroon (Fig. 1). The origin of the CVL is still the subject of considerable debate, with both plume and non-plume models invoked by many authors (e.g. Dorbath et al. 1986; Fairhead & Binks 1991; Lee et al. 1994; Ebinger & Sleep 1998; King & Ritsema 2000; Burke 2001; Ritsema & Allen 2003; Ngako et al. 2006; Déruelle et al. 2007). An example of a plume model is the flow of plume material from the Afar guided to the CVL by thinned lithosphere beneath central Africa rift system (Ebinger & Sleep 1988). Non-plume models include decompression melting beneath reactivated shear zones (Fairhead & Binks 1991) and small-scale mantle convection resulting from edge flow along the northern boundary of the Congo Craton (King & Ritsema 2000).

Crustal and upper mantle structure beneath Cameroon has been investigated previously using active (Stuart et al. 1985) and passive (Dorbath et al. 1986; Tabod 1991; Tabod et al. 1992; Plomerova et al. 1993) source seismic data, revealing a crust about 33 km thick at the south-western end of the continental portion of the CVL.
Figure 1. Simplified geological map of the study area. Volcanic areas that are part of the CVL are shown in red. The approximate northern boundary of the Congo Craton is shown with a green dashed line, and the small solid circles show approximate location of xenolith occurrences. FSZ, Foumban shear zone; CASZ, Central African shear zone. The approximate locations of the Etinde, Cameroon, Manengouba, Bamboutos, Oku and Mandara mountains are shown by numbers 1–6, respectively, and the Biu Plateau is shown with the number 7.

In this study, we investigate crustal structure more broadly beneath the CVL and the adjacent regions in Cameroon using 1-D shear wave velocity models obtained from the joint inversion of Rayleigh wave group velocities and P-receiver functions for 32 broad-band seismic stations (Fig. 2). From the 1-D shear wave velocity models, we obtain new insights into the composition and structure of the crust and upper mantle across Cameroon.

2 TECTONIC SETTING

2.1 The CVL

The continental sector of the CVL can be divided into a southern part, which extends from the coast to the southern edge of the Adamawa Plateau, and a northern part, which consists of two branches; one passing through the Adamawa Plateau and the other stretching towards the Biu Plateau in north-eastern Nigeria (Fig. 1). Both parts of the CVL are underlain by Pan African basement rocks consisting mainly of schists and gneisses intruded by granites and diorites (Fitton 1987; Déruelle et al. 2007, 1991). Cretaceous sediments, mostly sandstones and small amounts of limestone and shales, are found in the coastal plain.

Volcanic rocks that comprise the CVL range in composition from basalts to trachytes. For example, Mt. Manengouba consists of basalt, trachyte and rhyolite lavas, Mt. Cameroon, the largest of the continental volcanoes, consists mainly of alkaline basalts (Hedberg 1968) and Mt. Etinde, one of the older volcanoes, is made of nephelinitic lavas (Hedberg 1968; Nkoumbou et al. 1995). Other examples include Mt. Bamboutos, which is made of alkali basalts and trachytes, and Mt. Oku, which consists of transitional basalt, quartz trachyte and rhyolite flows (Fitton & Dunlop 1985). The Mandara Mountains, along the northern Cameroon–Nigeria border where some of the oldest volcanic rocks are found (c. 34 Ma), consist of trachyte and rhyolite plugs and alkali basalt flows (Fitton &
Dunlop 1985). The geochemical and isotopic similarities between the CVL oceanic and continental basalts attest that the continental crust did not play any role in the magma genesis and that the source is not of lithospheric origin (Déruelle et al. 2007). In spite of the similarities in geochemical composition of volcanic lavas, no evidence has been found for a consistent migration of volcanic activity with time beneath the CVL.

Along the CVL, mantle-derived (ultramafic) xenoliths have been found in several locations in basaltic lavas (Fig. 1, Déruelle et al. 1991; Princivalle et al. 2000; Déruelle et al. 2007). The xenoliths provide evidence for metasomatism within the upper mantle beneath the CVL (Déruelle et al. 2007).

2.2 The Benue trough

The Benue trough is a NE–SW trending basin that extends from the Niger delta basin (Gulf of Guinea) to Lake Chad (Fig. 1). Its origin is linked to the opening of the South Atlantic Ocean in the Cretaceous (Guiraud & Maurin 1992). The Yola trough, also known as the Garoua rift, and the Mamfe basin are eastward extensions of the Benue trough into Cameroon (Figs 1 and 2). The similar Y-shape of the Benue trough and CVL, together with the similarity in the composition of the alkali basalt in both the Benue trough and CVL (Coulon et al. 1996), suggest common geodynamic controls on their formation (Fitton 1987). For example, Guiraud & Maurin (1992) have argued that the orientation of the trough and CVL may be controlled by northeast-trending Pan African dextral shear zones.

2.3 The Oubanguides belt and the central African shear zone

Most of the Precambrian basement in Cameroon north of the Congo Craton belongs to the Pan African Oubanguides or North Equatorial fold belt (Poidevin 1983; Nzenti et al. 1988). The Oubanguides Belt is a branch of the much larger Neoproterozoic Pan African—Brazilian Belt, which resulted from the collision between the São Francisco Craton, the Congo Craton and the West African Craton.
during the formation of Gondwana (Castaing et al. 1994; Toteu et al. 2004). The Oubanguides Belt includes several mylonitic shear zones, most notably the Sanaga Fault and the Central African Shear Zone (CASZ) (Toteu et al. 2004, Fig. 1).

The CASZ is a major tectonic feature extending from the Darfur region in Sudan across central Africa to the Adamawa Plateau (Fairhead & Okereke 1987). From the Adamawa Plateau, the CASZ continues to the southwest, where it is known as the Foumban Shear Zone, before disappearing beneath the Tertiary to Recent volcanic cover in southwestern Cameroon. The shear zone can also be traced into Brazil, where it is called the Pernambuco lineament (Burke et al. 1971; Browne & Fairhead 1983).

### 2.4. The Congo Craton

The Congo Craton occupies a large part of Central Africa and its northern edge in southern Cameroon is referred to as the Ntem Complex (Vicat et al. 1996, Fig. 1). The Ntem Complex consists predominantly of Archaean rocks with some reworking in the Palaeoproterozoic (Tchameni et al. 2001). The Archaean rocks are preserved in greenstone belts surrounded by tonalite-trondhjemit-granodiorite (TTG) suites (Tchameni et al. 2000; Tchameni et al. 2001; Nsifa 2006). The palaeoproterozoic reworking of the Ntem Complex was contemporaneous with the Eburnean orogenic cycle. It is characterized by the intrusion of mafic doleritic dykes and a thermal or hydrothermal event c. 1800 Ma (Tchameni et al. 2001; Nsifa 2006).

The boundary between the Oubanguides Belt and the Congo Craton in southern Cameroon is poorly known and extends to the northeast towards the Central African Belt (Boukeke 1994; Toteu et al. 2004). Along this boundary, Pan African rocks overthrust the Ntem Complex, forming an intracrustal discontinuity (Boukeke 1994).

### 3. DATA AND METHODOLOGY

The data used for this study were recorded between 2005 January and 2007 February by the Cameroon Broadband Seismic Experiment, which consisted of 32 portable broad-band seismometers installed across the country (Fig. 2). Each station was equipped with a broadband seismometer (Guralp CMG-3T or Streckeisen STS-2), a 24-bit Reftek digitizer and a GPS (Global Positioning System) clock. Data were recorded continuously at 40 samples per second. Eight stations were installed in 2005 January and operated for 2 yr; the remaining 24 stations were operated only for the second year of the experiment. The station spacing during the second year of operation was about 50 to 150 km.

Data from the Cameroon Broadband Seismic Experiment have been used to perform a joint inversion of P-wave receiver functions and Rayleigh wave group velocities. Receiver functions are time-series that represent the radial impulse response of the shallow structure of the Earth in the vicinity of the seismic station (Langston 1979). They can be used to image velocity contrasts across discontinuities, and when modelled jointly with Rayleigh wave group velocities, the non-uniqueness inherent in receiver function inversion can be reduced (Julia et al. 2000).

#### 3.1 Rayleigh wave group velocities

Fundamental mode Rayleigh wave group velocities have been measured on 101 events of magnitude five and above with epicentral distances of less than 40° and recorded by the stations in Cameroon. A single station method was used for measuring the group velocities based on the multiple filter method of Dziewonski et al. (1969). Prior to making the measurements, the quality of each seismogram was checked and the instrument effect was removed.

The group velocity measurements were added to the database of similar measurements for Africa from Pasyanos (2005), and a tomographic inversion based on ray approximation was performed that increased the resolution of group velocity estimates within Cameroon compared to the group velocity maps published by Pasyanos (2005) and Pasyanos & Nyblade (2007). The new measurements increase significantly the density of stations and ray paths across the West Africa region that contains Cameroon. Fig. 3 shows examples of the ray coverage and variations in group velocities for periods of 20, 40 and 60 s. Both the CVL and Benue Trough are associated with a slow velocity anomaly in the 60-s period map. These results are generally similar to preliminary Rayleigh wave phase velocity maps from Euler et al. (2008), who have analysed the Cameroon data using a two plane-wave decomposition (Forsyth & Li 2005). Resolution tests of the group velocity model indicate that the spatial resolution at periods most sensitive to crustal structure (~10–50 s) is 2°–3°, and thus the group velocity models have sufficient resolution to image differences in group velocities between regions that are ~200–300 km wide. A single dispersion curve for each station from 7 to 100 s was obtained from the group velocity maps. The curves were smoothed using a three-point running average before using them in the joint inversion.

#### 3.2 Receiver functions

Receiver functions were computed using data from 69 teleseismic events that occurred at epicentral distances between 30° and 95° and that had magnitudes ≥5.5 (Fig. 4). Two overlapping frequency bands corresponding to Gaussian widths of $a = 1.0$ and $a = 2.5$ (corner frequencies of 0.5 Hz and 1.2 Hz, respectively) were used to compute the receiver functions because they help discriminate gradational transitions from sharp discontinuities in the receiver structure under the station (Julia 2007).

To compute the receiver functions, the selected waveforms were decimated to 10 samples per second, windowed between 10 s before and 100 s after the leading $P$ arrival, de-trended, tapered and high-pass filtered above 50 s to remove low-frequency, instrumental noise. Radial and transverse receiver functions were then obtained from the filtered traces by rotating the original horizontal components around the corresponding vertical component into the great-circle path, and deconvolving the vertical component from the radial component through the iterative time domain deconvolution procedure of Ligorio & Ammon (1999), with 200 iterations.

The percentage of recovery of the original radial waveform was evaluated from the rms misfit between the original radial waveform and the convolution of the radial receiver function with the original vertical component, and the events that were recovered to less than 85 per cent were rejected. The remaining waveforms were visually inspected for coherence and stability, and were then stacked and clustered by ray parameter and backazimuth (Fig. 4). At least three waveforms were required to perform the stacks with the exception of station CM27, for which only two waveforms were successfully recovered.

The transverse receiver functions (see Tokam 2010) were computed to check the degree of lateral heterogeneity and isotropy of the propagating medium (Cassidy 1992). Small amplitudes...
Figure 3. Rayleigh wave group velocity maps from an updated version of the maps presented in Pasyanos (2005). Top left-hand side: ray coverage for 20 s Rayleigh waves including the additional data from the Cameroon network. Other panels show group velocities for the periods indicated above each panel.

on the transverse component indicate laterally homogeneous and isotropic media beneath a station. The lateral extent of the area sampled by a receiver function roughly equals the depth to the deepest reflecting interface, and may vary with azimuth depending on the approaching direction of the incoming $P$ wave (Julià et al. 2000). In general, compared to radial receiver functions, the transverse waveforms for selected azimuths were found to have small amplitudes, apart from the waveforms for stations CM09 and CM15, indicating for both of these stations a high degree of lateral heterogeneity.
The joint inversion was performed using the method developed by Julià et al. (2000, 2003). The method is based on a linearized inversion procedure that minimizes a weighted combination of least-squares norms for each data set, a model roughness norm and a vector-difference norm between inverted and pre-set model parameters. The velocity models obtained are consequently a compromise between fitting the observations, model simplicity and a priori constraints. To make the contribution of each data set to the joint least-squares misfit comparable, a normalization of the data set is necessary, and this is done using the number of data points and variance for each of the data sets. An influence factor is used to control the trade-off between fitting the receiver functions and the group velocity curves. During the inversions, equal contribution of both dispersion data and receiver functions was assumed (i.e. an influence factor value of 0.5 was used).

The starting model used in the joint inversions consisted of an isotropic medium with a 37.5-km-thick crust and a linear shear wave velocity increase in the crust from 3.4 to 4.0 km s\(^{-1}\) overlying a flattened PREM (Preliminary Reference Earth Model) model (Dziewonski & Anderson 1981) for the mantle. Poisson’s ratio was set to 0.25 for the crust, and crustal densities were deduced from P-wave velocities through the empirical relationship of Berteussen (1977). The starting model consisted of constant velocity layers that increase in thickness with depth. The thickness of the first and second layers are, respectively, 1 and 2 km, while the thickness increases to 2.5 km between 3 and 60.5 km depth, 5 km between 60.5 and 260.5 km depth and 10 km below a depth of 260.5 km.

The models obtained from the inversion show a good fit to the group velocity curves and receiver functions, with the exception of stations CM03, CM05, CM09, CM15 and CM23, where fits to some of the receiver functions are not as good. In general, the group velocities below 10 s period are sensitive to shallow crustal structures and sedimentary basins. For periods up to 40–50 s, they are sensitive to the whole crust and between periods of 50 and 100 s, they preferentially sample the uppermost mantle (Pasyanos 2005). Consequently, models were constrained to PREM velocities (Dziewonski & Anderson 1981) below 200 km depth in the inversion while at the same time inverting for velocity structure above 200 km depth. A Poisson’s ratio of 0.25 was fixed for the crust during the inversions and the Poisson’s ratio for the PREM model was used for the mantle.

### 3.4 Model uncertainties

The approach of Julià et al. (2008) was applied to check for laterally varying structures around each station. In this approach, receiver functions are stacked in groups by backazimuth and ray parameter, and because the receiver functions were computed for each event at overlapping frequencies \(a = 1.0\) and \(a = 2.5\), two receiver function stacks were obtained for each group. Each receiver function group at each station was then jointly inverted with the corresponding group velocity curve to obtain a shear velocity model. For comparison to the velocity models obtained for each group, an average S-velocity model was obtained by inverting a stack of all the receiver functions from that station with the same dispersion curve. The procedure was applied to stations that have at least four groups of receiver functions. The result for station CM07 is illustrated in Fig. 5 with event details provided in Table 1.

In general, there should not be any significant variations between the models for the average and single groups for stratified, isotropic media. Variations between the models should result from the interaction of the \(P\) waves with lateral heterogeneities in the crust and upper mantle, a dipping interface or strong azimuthal anisotropy. For all of the stations (except CM03, CM05, CM09, CM15 and CM23), there is good agreement between models, as illustrated for station CM07 in Fig. 5.

To estimate the uncertainties in the S-velocity models, we computed the standard deviation for the models obtained for each group...
Figure 5. (a) Joint inversion results for station CM07 for individual groups of receiver functions. The top, middle and bottom panels in each column display receiver functions, group velocities and shear wave velocity models, respectively. For the receiver functions, observations are shown in grey and predictions in black. Above each cluster of receiver functions is given in sequence the number of waveforms ($N$), average backazimuth (degrees) ± one standard deviation ($\text{baz} \pm \text{bazdev}$) and average ray parameter ($s \text{ km}^{-1}$) ± one standard deviation. For the group velocity plots, the observations are shown with triangles and the model results are shown with the solid line. For the velocity models, the dashed line is the starting model and the solid line is the model obtained from the inversion. (b) Shear wave velocity models obtained using the whole group of receiver functions. (c) Superposition of single (grey lines) and full (black line) joint inversion models.
of receiver function stacks. Estimates of uncertainties in shear velocity for stations with four or more groups are less than 0.2 km s$^{-1}$ for the crustal layers. In addition, previous studies using the joint inversion method have investigated the uncertainties introduced by fixing the crustal Poisson’s ratio to 0.25 and found that shear wave velocities change by no more than $\sim$0.1 km s$^{-1}$ for a reasonable range of values for Poisson’s ratio (Dugda et al. 2009; Kgaswane et al. 2009). Therefore, we place the overall uncertainty in shear velocity at 0.2 km s$^{-1}$ for any given crustal layer in the model. This results in an uncertainty of 2–3 km in Moho depth for most stations where a rapid increase of velocity can be observed at the crust–mantle boundary, and no more than 5 km where a smoothly varying shear velocity is found indicating a gradational Moho.

## 4 RESULTS

Data from two stations in the Cameroon network (CM08 and CM14) were not included in this study because due to equipment failures not enough high quality waveforms were available for computing receiver functions. Results from the joint inversion for the remaining stations are shown in Figs 6 and 7, summarized in Table 2 and illustrated further in the Supporting Information.

Fig. 6 shows the modelling results for selected stations, and in Fig. 7 the shear wave velocity models are shown for all of the stations clustered by tectonic terrain. Crustal thickness beneath each station was determined by placing the Moho at the depth where the shear wave velocity exceeds 4.3 km s$^{-1}$. Shear wave velocities for typical lower crust lithologies obtained by using experimentally determined $P$-wave velocities and $Vp/Vs$ ratios (e.g. Christensen & Mooney 1995; Christensen 1996) show that shear wave velocities in the lower crust cannot be higher than 4.3 km s$^{-1}$. Therefore, we take shear wave velocities above 4.3 km s$^{-1}$ to indicate the presence of lithologies with mantle compositions, and we place the Moho where the shear wave velocities are close to or exceed that value. For many stations (e.g. stations of the Adamawa Plateau except CM27 on Fig. 7), there is a significant increase in velocity at the depth at which the shear wave velocity exceeds 4.3 km s$^{-1}$, but for other stations (e.g. CM03, CM10 and CM17 of the Oubanguides Belt) the change in shear wave velocity is gradational from the lowermost crust into the upper mantle.

### 4.1 Cameroon Volcanic Line

As described in Section 2.1, we split the CVL into a southern part, comprising the region to the southwest of the Adamawa Plateau, and a northern part, which is comprised of the Adamawa Plateau. Beneath the highland regions of the southern CVL, crustal thickness estimates are on average about 36 km, compared to 28 and 31 km in the Kumba Graben and Mamfe Basin, respectively (Table 2). Beneath Mt. Cameroon (station CM09), the shear wave velocity is complicated, and at two depths the shear wave velocity reaches 4.3 km s$^{-1}$, making it difficult to determine crustal thickness (Fig. 7).

An average crustal shear wave velocity of 3.7 km s$^{-1}$ is found at most of the stations in the southern part of the CVL, and slightly slower velocities (3.6 km s$^{-1}$) for the Mamfe Basin and Kumba Graben (Table 2). The models in Figs 6 and 7 show within the top 15 km of the upper crust the presence of a fast layer, with velocities of 3.6–3.8 km s$^{-1}$, creating the appearance of a low velocity zone below it in the mid-crust. The crust below 20 km depth for the southern part of the CVL has an average shear wave velocity of about 3.9 km s$^{-1}$ and a thin, higher velocity layer ($V_s \geq 4.0$ km s$^{-1}$) at the base of the crust. The thickness of this high velocity layer is about 3 km at most stations, but is thicker than that at a few stations (CM19, CM23 and CM27). The average uppermost mantle velocity, determined by computing the average velocity between the Moho and 60 km depth, is 4.4–4.5 km s$^{-1}$ (Table 2).

The structure of the northern part of the CVL, on average, is very similar to the southern part. The Moho is on average at a depth of 36 km beneath the Adamawa uplift, the average crustal $V_s$ is 3.7 km s$^{-1}$, the crust below 20 km depth has an average $V_s$ of 3.9 km s$^{-1}$, and the high velocity layer at the base of the crust is either very thin or else absent (Fig. 7 and Table 2).

### 4.2 Oubanguides belt

Stations CM03, CM10, CM12 and CM17 are located in the part of the Oubanguides Belt away from the CVL. The crustal thickness varies from 36 to 43 km with an average crustal shear wave velocity of about 3.8 km s$^{-1}$ for the area (Fig. 7 and Table 2). For crust below 20 km depth, the shear velocity is 3.9–4.0 km s$^{-1}$, on average, and a high velocity layer is also observed at the base of the crust with a thickness of 5–8 km. The uppermost mantle has an average shear wave velocity of 4.5 km s$^{-1}$, slightly higher than the uppermost mantle velocity beneath the CVL.

### 4.3 Congo Craton

The Congo Craton is characterized by a crustal thickness of 43–48 km with an average crustal shear wave velocity of 3.9 km s$^{-1}$. A fast layer (3.6–3.8 km s$^{-1}$) in the upper crust is visible in the velocity models for all the stations (Fig. 7), and below this fast layer there is a gradual increase of velocity with depth to the Moho. The shear wave velocity below 20 km depth is 4.0–4.2 km s$^{-1}$, which is substantially faster than shear wave velocities in that depth range beneath the CVL and Oubanguides Belt. This high velocity layer is the prominent feature of the lower crust across the northern part of the Congo Craton.
Figure 6. Sample of joint inversion results for selected stations computed with at least four clusters of receiver functions. Each column shows results for a single station. The top, middle and bottom panels in each column display receiver functions, group velocities and shear wave velocity models, respectively. Labelling is similar to Fig. 5.
Figure 7. Shear wave velocity profiles grouped by tectonic terrain. Lines on each profile are shown for reference. The solid line is at 4.0 km s$^{-1}$ and the dashed line at 4.3 km s$^{-1}$. 
Joint inversion in Cameroon

4.4 Coastal plain

The Moho at stations CM01 and CM05 is found at a depth of 28 km (Table 2). The fast layer in the upper crust can be seen in velocity models for both stations, as well as an average lower crustal shear wave velocity of 4.1 km s\(^{-1}\). The terrain is underlain by uppermost mantle with a shear wave velocity of 4.4 km s\(^{-1}\).

4.5 Garoua rift

The crustal thickness is variable across this part of Cameroon. The Moho is found at a depth of 26 km beneath station CM29, and 31–33 km beneath stations CM28, CM31 and CM31 to the north and south of the rift. The average crustal shear wave velocity is about 3.4–3.5 km s\(^{-1}\) across the region. The upper crustal structure is characterized by a shallow, thin layer with a low shear wave velocity (\(V_s < 3\) km s\(^{-1}\)). The shear wave velocity below 20 km is variable with an average of 3.7–3.9 km s\(^{-1}\) for the region. The average uppermost mantle shear wave velocity is 4.3–4.4 km s\(^{-1}\) across the region, which is somewhat slower compared to the uppermost mantle velocity for the other parts of Cameroon.

5 DISCUSSION

To summarize, there are four main findings that come from the joint inversion of the receiver functions and Rayleigh wave group velocity curves. (1) Crustal structure is similar beneath the CVL and the Oubanguides Belt to the south of the CVL. (2) The crust is thicker under the Congo Craton than the Oubanguides Belt and is characterized by shear wave velocities \(\geq 4.0\) km s\(^{-1}\) in the lower part of the crust. (3) Thinner crust is found under the Garoua rift and the coastal plain. (4) A fast velocity layer (\(V_s\) of 3.6–3.8 km s\(^{-1}\)) in the upper crust is found beneath many of the seismic stations, and is not confined to just one region. In this section, we first briefly compare our results to previously published results on crustal structure in Cameroon and then we examine to what extent, if any, these findings, as well as the average structure of the crust of each region, are anomalous with respect to the structure of similar aged crust in other parts of Africa.

5.1 Comparison with previous estimates of crustal structure

Our estimates of crustal thickness are in good agreement with previous estimates based on both gravity and seismic data for a number of regions (Table 3). Many of these estimates come from interpreting gravity data, and in Fig. 8 we show that there is a strong correlation between crustal thickness and Bouguer gravity anomalies along the length of the CVL. Small departures between the Bouguer gravity anomaly and crustal thickness can be attributed to near-surface structure. For example, the small decrease in the Bouguer anomaly around station CM29 is likely the effect of 3–4 km of sediments in the Garoua rift (Stuart et al. 1985), while the small increase at station CM25 correlates with the high velocity layer in the upper crust beneath this station.

The correlation between the Bouguer gravity anomaly and crustal thickness also corroborates our choice of Moho depth on the velocity profile for station CM09. On that profile, there are two depths...
Table 2. Summary of crustal structure by geological terrains shown in Fig. 7.

<table>
<thead>
<tr>
<th>Terrain</th>
<th>Stations</th>
<th>Average crustal thickness (km)</th>
<th>Average crustal thickness ± standard deviation (km)</th>
<th>Average Uppermost mantle Vs (km s⁻¹)</th>
<th>Average crustal Vs below 20 km depth (km)</th>
<th>Average thickness of crustal layers with Vs ≥ 4.0 km s⁻¹ (km)</th>
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<tr>
<td>Coastal plain</td>
<td>CM01</td>
<td>28</td>
<td>28</td>
<td>4.4</td>
<td>3.7</td>
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<td></td>
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<td></td>
<td>3.7</td>
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<td>4.1</td>
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<td>44.7 ± 1.5</td>
<td>4.6</td>
<td>3.8</td>
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<td>4.1</td>
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<td>Pan African</td>
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<td>3.8</td>
<td>4.0</td>
<td>7.5</td>
</tr>
<tr>
<td></td>
<td>CM12</td>
<td>38</td>
<td></td>
<td>3.8</td>
<td>3.9</td>
<td>7.5</td>
</tr>
<tr>
<td></td>
<td>CM17</td>
<td>35.5</td>
<td></td>
<td>3.7</td>
<td>3.9</td>
<td>5</td>
</tr>
<tr>
<td>Southern CVL</td>
<td>Mt. Cameroon High lands</td>
<td>CM09</td>
<td>25.5 ± 2.5</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td></td>
<td>CM15</td>
<td>33</td>
<td>35.5 ± 3.1</td>
<td>4.5</td>
<td>3.6</td>
<td>3.9</td>
</tr>
<tr>
<td></td>
<td>CM16</td>
<td>35.5</td>
<td></td>
<td>3.7</td>
<td>3.9</td>
<td>5</td>
</tr>
<tr>
<td></td>
<td>CM19</td>
<td>35.5</td>
<td></td>
<td>3.7</td>
<td>3.9</td>
<td>5</td>
</tr>
<tr>
<td></td>
<td>CM20</td>
<td>33</td>
<td></td>
<td>3.7</td>
<td>3.9</td>
<td>2.5</td>
</tr>
<tr>
<td></td>
<td>CM23</td>
<td>40.5</td>
<td></td>
<td>3.7</td>
<td>4.0</td>
<td>10</td>
</tr>
<tr>
<td></td>
<td>Kumba graben Mamfe basin</td>
<td>CM11</td>
<td>28</td>
<td>29.2 ± 1.8</td>
<td>4.4</td>
<td>3.6</td>
</tr>
<tr>
<td></td>
<td>CM18</td>
<td>30.5</td>
<td></td>
<td>3.6</td>
<td>3.9</td>
<td>2.5</td>
</tr>
<tr>
<td>Adamawa Plateau</td>
<td>(northeastern CVL)</td>
<td>CM21</td>
<td>35.5</td>
<td>35.5 ± 1.6</td>
<td>4.4</td>
<td>3.7</td>
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<td></td>
<td>CM22</td>
<td>35.5</td>
<td></td>
<td>3.7</td>
<td>3.9</td>
<td>2.5</td>
</tr>
<tr>
<td></td>
<td>CM24</td>
<td>35.5</td>
<td></td>
<td>3.7</td>
<td>3.8</td>
<td>-</td>
</tr>
<tr>
<td></td>
<td>CM25</td>
<td>38</td>
<td></td>
<td>3.7</td>
<td>3.9</td>
<td>2.5</td>
</tr>
<tr>
<td></td>
<td>CM26</td>
<td>33</td>
<td></td>
<td>3.6</td>
<td>3.8</td>
<td>-</td>
</tr>
<tr>
<td></td>
<td>CM27</td>
<td>35.5</td>
<td></td>
<td>3.7</td>
<td>3.9</td>
<td>5.0</td>
</tr>
<tr>
<td>Garoua rift</td>
<td>South of Garoua rift</td>
<td>CM28</td>
<td>30.5</td>
<td>30.5 ± 3.5</td>
<td>4.3</td>
<td>3.5</td>
</tr>
<tr>
<td></td>
<td>Garoua rift</td>
<td>CM29</td>
<td>25.5</td>
<td>25.5 ± 3.5</td>
<td>4.3</td>
<td>3.4</td>
</tr>
<tr>
<td></td>
<td>North of Garoua rift</td>
<td>CM30</td>
<td>28</td>
<td>30.5 ± 2.5</td>
<td>4.4</td>
<td>3.4</td>
</tr>
<tr>
<td></td>
<td>CM31</td>
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<td>3.5</td>
<td>3.9</td>
<td>2.5</td>
</tr>
<tr>
<td></td>
<td>CM32</td>
<td>33</td>
<td></td>
<td>3.5</td>
<td>3.9</td>
<td>2.5</td>
</tr>
</tbody>
</table>

Table 3. Comparison of crustal thickness estimates from this study with crustal thickness estimates from previous studies.

<table>
<thead>
<tr>
<th>Average crustal thickness (km)</th>
<th>This study</th>
<th>Other studies</th>
<th>Type of data used</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>Garoua rift</td>
<td>25.5</td>
<td>23</td>
<td>Seismic</td>
<td>Stuart et al. (1985)</td>
</tr>
<tr>
<td></td>
<td>24</td>
<td></td>
<td>Gravity</td>
<td>Kamguia et al. (2005)</td>
</tr>
<tr>
<td>CVL in general</td>
<td>-</td>
<td>30-34</td>
<td>Gravity</td>
<td>Fairhead &amp; Okereke (1987)</td>
</tr>
<tr>
<td>CVL southern part</td>
<td>35.5</td>
<td>~33</td>
<td>Seismic</td>
<td>Tabod (1991)</td>
</tr>
<tr>
<td>Adamawa Plateau</td>
<td>35.5</td>
<td>33</td>
<td>Seismic</td>
<td>Stuart et al. (1985)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Gravity</td>
<td>Nnange et al. (2000)</td>
</tr>
<tr>
<td>Congo craton</td>
<td>45.0</td>
<td>50 ± 10</td>
<td>Gravity</td>
<td>Poudjom et al. (1995)</td>
</tr>
</tbody>
</table>

at which the shear wave velocity increases to >4.3 km s⁻¹, one at 25.5 km and the other at 40.5 km. Our placement of the Moho at the shallower discontinuity is consistent with the Bouguer gravity anomaly, which indicates a thin crust beneath CM09 (Fig. 8).

5.2 Comparison with crustal structure in other parts of Africa

For the comparison of crustal structure in Cameroon to elsewhere in Africa, we use the structure of crust in eastern and southern Africa, where the structure of Archaean and Proterozoic crust has been well imaged using the same methods as used in this study. The relevant details of crustal structure are summarized in Table 4.

5.2.1 Pan African

The Mozambique Belt in eastern Africa developed during the Pan African orogenic event along the eastern side of the Archaean Tanzania Craton. The structure of the Mozambique Belt crust has been well studied using the joint inversion of receiver functions and
Figure 8. Top panel: Crustal thickness along the profile shown in Fig. 1. Bottom panel: point Bouguer anomalies along the same profile, extracted from the simple Bouguer gravity anomaly map of Kamguia et al. (2008). A strong correlation is observed between the Bouguer gravity anomalies and crustal thickness estimates along the profile. The decrease observed around CM29 is likely the effect of the sediments in the Garoua rift.

Table 4. Comparison of crustal structure in Cameroon to similar aged terrains in eastern and southern Africa.

<table>
<thead>
<tr>
<th>Terrain</th>
<th>Average Moho depth (km)</th>
<th>Average Mafic layer thickness (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tanzania craton</td>
<td>40 ± 3</td>
<td>&lt;10</td>
</tr>
<tr>
<td>Mozambique belt</td>
<td>38 ± 3</td>
<td>&lt;10</td>
</tr>
<tr>
<td>Zimbabwe craton</td>
<td>36 ± 1</td>
<td>10 ± 4</td>
</tr>
<tr>
<td>Kaapval craton</td>
<td>38 ± 3</td>
<td>9 ± 5</td>
</tr>
<tr>
<td>Bushveld complex</td>
<td>41 ± 3</td>
<td>10</td>
</tr>
<tr>
<td>Limpopo belt</td>
<td>41 ± 3</td>
<td>14 ± 2</td>
</tr>
<tr>
<td>Congo craton</td>
<td>45 ± 2</td>
<td>23 ± 4</td>
</tr>
<tr>
<td>Oubanguides belt</td>
<td>39 ± 3</td>
<td>6 ± 2</td>
</tr>
<tr>
<td>CVL</td>
<td>36 ± 3</td>
<td>5 ± 3</td>
</tr>
<tr>
<td>Garoua rift</td>
<td>26 ± 3</td>
<td>2 ± 1</td>
</tr>
</tbody>
</table>

Julià et al. (2005) Kgaswane et al. (2009) This study

Rayleigh wave group and phase velocities (Julià et al. 2005; Dugda et al. 2009), and also by seismic refraction profiling (Fuchs et al. 1997 and references therein).

The crustal thickness beneath the Mozambique Belt varies between 36 and 42 km with an average of 38 km. The bottom of the Mozambique Belt crust is characterized by a 5–10 km thick layer of high velocity ($V_s \geq 3.9 \text{ km s}^{-1}$) rock, possibly indicating a mafic lithology. Julià et al. (2005) suggest that the origin of the mafic rock could be from underplating during magmatic events. Crustal structure beneath the Pan African Oubanguides Belt, including the southern CVL and Adamawa Plateau, is very similar to the Mozambique Belt, with crustal thickness ranging between 36 and 43 km and a high velocity layer at the base of the crust with a thickness of 5 to 8 km. Therefore, the CVL appears to have not affected significantly the bulk crustal structure of the Pan African crust in Cameroon. Certainly the crust under some of the stations along the CVL has been perturbed by the magmatism, but overall there is no discernable modification to average crustal structure, at least not that can be easily resolved with the joint inversion method used in this study.
In contrast to that conclusion, the Pan African crust under the Garoua rift and coastal plain has been thinned by 10 or more kilometres. However, this thinning has probably not resulted from the CVL but rather from other tectonic events. The crust beneath the coastal plain was likely thinned during the opening of the southern Atlantic Ocean, while the crust beneath the Garoua rift was likely thinned during the formation of the Benue Trough in the early Cretaceous (Stuart et al. 1985; Baudin 1991; Kamguia et al. 2005).

5.2.2 Congo Craton

Using the joint inversion of receiver functions and Rayleigh wave group and phase velocities, crustal structure of the Tanzanian Craton has been imaged by Julià et al. (2005) and crustal structure of the Zimbabwe and Kaapvaal Cratons has been imaged by Kgaswane et al. (2009). The Tanzania Craton is underlain by crust that is between 38 and 42 km thick with and average of 40 km, while crustal thickness beneath the Kaapvaal and Zimbabwe Cratons is between 36 and 40 km, with averages of 38 and 36 km, respectively. The average thickness of the high velocity (mafic) lower crust is 10 km or less for all three cratons.

In comparison, the crust beneath the Congo Craton, which ranges from 43 to 48 km, is significantly thicker, and the high velocity layer in the lowermost crust is also much thicker. The average thickness of the high velocity layer is 23 km, more than 10 km thicker than that found beneath the other cratons (Table 4). Interestingly, the thickness of the high velocity layer is as thick as or thicker than the high velocity layer found beneath the Bushveld Complex in the northern part of the Kaapvaal Craton and parts of the Limpopo Belt that form a suture zone between the Zimbabwe and Kaapvaal Cratons (Kgaswane et al. 2009).

Examining magmatic events that post-date the formation of the Ntem Complex (Congo Craton as defined in Section 2.4) does not provide much help in explaining the very thick mafic lower crust. Mafic dykes older than 2100 Ma associated with the rifting of the Ntem Complex crust, as well as the intrusion of gabbros and the development of greenstone belts, provide evidence for magmatic events that could have added mafic rock to the lower crust (Maurizot et al. 1986; Vicat et al. 1996; Tchameni et al. 2000; Tchameni et al. 2001; Shang et al. 2004; Nsifa 2006), but similar tectono-magmatic events have affected the Tanzania, Zimbabwe and Kaapvaal Cratons and therefore are not unique to the Ntem Complex. In addition, there is no indication that a magmatic event equivalent in scale to the Bushveld has affected the Ntem Complex, and so such an event is also not a plausible explanation for the thick mafic lower crustal layer of the Ntem Complex.

Another possibility is that both the thickened crust and the thick mafic lower crustal layer are relict features from a continent–continent collision during the formation of Gondwana along the northern boundary of the Congo Craton. A suture zone between the Congo Craton and the Oubanguides Belt has been suggested in several studies (e.g. Penaye et al. 1993; Totea et al. 2004; Tadjou et al. 2009), although the exact location of the suture is still a matter of some debate.

In Precambrian sutures elsewhere that involve an Archaean craton, such as the Limpopo Belt (Kgaswane et al. 2009), the Superior Province (Gibb et al. 1983), the Tanzania Craton (Nyblade & Pollack 1992), the Yilgarn Craton (Mathur 1974; Wellmann 1978), the Indian shield (Subrahmanyan 1978) and the Mann shield (Blot et al. 1962; Louis 1978; Black et al. 1979), 5–10 km of crustal thickening is observed along with the presence of mafic units in a crust commonly affected by granulite facies metamorphism and extraction of a felsic partial melt component. Both the thicker crust and the large thickness of lower crust with high shear wave velocities in the Ntem Complex is consistent with typical ‘suture’ thickened crust found in other Precambrian terrains, and thus we suggest this is a viable explanation for the nature of crustal structure beneath the Ntem Complex.

5.3 Fast upper crustal layer

At many stations within all of the regions in Cameroon, a fast (Vs of 3.6–3.8 km s\(^{-1}\)) layer in the upper crust can be seen (Fig. 7), suggesting a fair amount of heterogeneity within the upper crust both within and between regions. We interpret these layers generically as mafic bodies intruded into the upper crust during magmatic events of various ages. The occurrence of ultramafic xenoliths along the CVL (Nkoumbou et al. 1995; Düruele et al. 2007) supports this interpretation, at least for the regions of Cenozoic volcanism.

6 SUMMARY AND CONCLUSIONS

A joint inversion of Rayleigh wave group velocities and receiver functions has been used to investigate the structure of the crust beneath Cameroon. The main findings are that (1) crustal structure is similar beneath the CVL and the Oubanguides Belt to the south of the CVL, (2) the crust is thicker under the Congo Craton than the Oubanguides Belt and is characterized by shear wave velocities ≥4.0 km s\(^{-1}\) in the lower part of the crust, (3) the crust is thinner beneath the Garoua rift and the coastal plain and (4) there is a fast velocity layer (Vs of 3.6–3.8 km s\(^{-1}\)) in the upper crust beneath many of the seismic stations. Our estimates of crustal thickness are in good agreement with previous estimates based on both gravity and seismic data for a number of areas within Cameroon.

We have compared crustal structure in Cameroon to the structure of crust in eastern and southern Africa, where the structure of Archaean and Proterozoic crust has been well imaged using the same methods as in this study. Crustal structure beneath all of the Pan African Oubanguides Belt, including the southern CVL and Adamawa Plateau, is very similar to the Mozambique Belt, and therefore the CVL appears to have not affected the bulk crustal structure of the Pan African crust in Cameroon. However, the Pan African crust under the Garoua rift and coastal plain has been thinned by 10 or more kilometres. The crust beneath the coastal plain was probably thinned during the opening of the southern Atlantic Ocean, while the crust beneath the Garoua rift was likely thinned during the formation of the Benue Trough in the early Cretaceous.

In comparison, the crust beneath the Congo Craton is significantly thicker than beneath other parts of Cameroon and is, on average, also significantly thicker than the crust beneath the Tanzania, Kaapvaal and Zimbabwe cratons. In addition, the Congo Craton crust has a high velocity layer in the lowermost crust more than 10 km thicker than that found beneath the other cratons. We suggest that the northern boundary of the Congo Craton could be a suture zone that formed during the Pan African orogeny, and that both the thickened crust and the thick mafic lower crustal layer are relict features from a continent–continent collision during the formation of Gondwana.

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REFERENCES


**SUPPORTING INFORMATION**

Additional Supporting Information may be found in the online version of this article:

**Figure S1.** Joint inversion results for all stations used in this study. Each column shows results for a single station. The top, middle and bottom panels in each column display receiver functions, group velocities and shear wave velocity models, respectively. The top, middle and bottom panels in each column display receiver functions, group velocities and shear wave velocity models, respectively. Observations are shown in black and predictions in red. Above each cluster of receiver functions $|N|ba z \pm s b a z r a y \pm s r a y$, where $N$ is the number of waveforms, baz and sbaz are the average and one-standard deviation backazimuths, and ray and sray are the average and one-standard deviation ray parameters in s km$^{-1}$.

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