Authors' reply to RC1

Specific comments RC1

Part 1:

RC1-1: The introduction is fine. However, the paragraphs (line 17, page 2 to line 14 page 3) describing the interactions between mantle dynamics and lithosphere is not clear. What kind of interactions are you talking about? (is it rifting? volcanism? doming?). It is not clear how your model can help to improve the understanding of these interactions. Please be more accurate.
We have rephrased and complemented the last part of the introduction of the revised manuscript to clarify this point:
"The models allow a straightforward spatial correlation between modelled strength heterogeneities and regional deformation structures of the rift, seismicity patterns and major locations of volcanic activity, which we briefly discuss in terms of causal relationships. To improve upon these aspects, future studies are planned, which will integrate the presented 3D models into numerical forward geodynamics experiments in order to test hypotheses on the entire Cenozoic deformation history of the study area. In any case, the model as it now stands already shows how far a compositionally heterogeneous crust has controlled lithospheric deformation and thus rift localisation and propagation processes."

RC1-2: You consistently refer to western and eastern Kenya throughout the paper. What do they represent? Is western Kenya located west of the rift and eastern Kenya east of the rift? Please specify it at the beginning of the paper (in the introduction for example).
We have provided a short paragraph at the end of the new chapter "2 Geological Setting" (see below) to describe *a priori* differences between western and eastern Kenya with respect to the topography, basement depth and the distribution of KRISP refraction seismic profiles. With respect to this point, we would like to add that, given the complexity of the rift system we could not draw any clear demarcation line to separate western from eastern Kenya in a way that might be consistent with all of the observations presented in the manuscript. In the discussion of the revised manuscript, we also refer to this distinction, while again its meaning depends on the property discussed and is best derived from the figures referred to.

RC1-3: A section "Geological setting and/or history" is missing between the introduction and part 2. Such a section may be useful to readers who are not familiar with the geology of the Kenya rift area and the main geodynamic events (amalgamation, rifting episodes, plume emplacement…).
We have added the section "2 Geological Setting" to the manuscript. Thereby, all information given in this section has been derived from existing sections, i.e. the sections "1 Introduction", "2.1 Constraints on the density configuration of the sedimentary and volcanic rocks" and "2.2 Constraints on the density configuration of the crystalline crust" of the original manuscript. This has helped not to unnecessarily increase the length of the manuscript. Please note the required adjustments also in these sections of the revised manuscript.
The new section "2 Geological Setting" of the revised manuscript reads as follows:
"The formation of the continental crust in East Africa dates back to the Neoproterozoic when the East African Orogeny (at ≈650-620 Ma) led to the amalgamation of numerous terranes to form central Gondwana (e.g., Fritz et al., 2013). This orogeny resulted from collisions of the Arabian Nubian Shield and its southward continuation, the Mozambique Belt (Holmes, 1951), with the Tanzania (Nyanzian) Craton to the west and the Azania microcontinent to the east. According to Fritz et al. (2013), five major tectonothermal domains of different Precambrian ages and lithologies are juxtaposed against each other in the study area (Fig. 1b). From W to E, these are

(i) the Nyanzian System (Clifford, 1970) of the Archean Tanzania Craton; (ii) the Western Granulites representing reworked pre-Neoproterozoic crust of the Mozambique Belt (Maboko, 1995; Möller et al., 1998); (iii) the Eastern Granulites representing Neoproterozoic juvenile crust of the Mozambique Belt (Möller et al., 1998; Maboko and Nakamura, 2002; Tenczer et al., 2006); (iv) the Neoproterozoic Arabian Nubian Shield; and (v) the microcontinent Azania representing reworked pre-Neoproterozoic crust (Fritz et al., 2013). Near-surface observations indicate that rocks of the Mozambique Belt and the Arabian Nubian Shield structurally overlie both the craton in the west and the microcontinent in the east (Fritz et al., 2013).

The top of the crystalline basement in the study area (Fig. 2a) is overlain by sedimentary and volcanic rocks of Permo-Carboniferous to Holocene ages (Beicip, 1987), while its geometry reflects different phases of sedimentary basin formation and localised subsidence. For instance, the Mandera and Lamu Basins in eastern Kenya (Fig. 2a) are regarded as the north-easternmost extension of the Karoo rift system, the initiation of which was related to the Late Carboniferous-Early Permian assembly and subsequent breakup of Pangea (Catuneanu et al., 2005). After an early phase of eastward rifting of Madagascar (and India) away from the conjugate block of Kenya and northern Tanzania (e.g. Reeves et al., 2002), from ≈185 180 Ma Madagascar moved southwards (e.g. Cox, 1992), leading to the formation of oceanic crust in the Indian Ocean (at <166 152 Ma; Seton et al., 2012) and transforming the Lamu Basin area into a passive margin setting.

Farther west, the oldest structural elements of the NW-SE oriented Anza Basin (Fig. 2a; Bosworth and Morley, 1994) and the N S oriented Lotikipi Plain, Turkana, Lokichar, and North Kerio basins (e.g., Morley, 1999) began forming during the Cretaceous and continued subsiding into the Cenozoic (Foster and Gleadow, 1996; Morley, 1999; Tiercelin et al., 2012). The Anza Basin has been regarded as part of the E-W striking Central African Rift System (Guiraud et al., 2005; Heine et al., 2013), which formed under the influence of (i) the northeastward movement of the Arabian-Nubian block, (ii) ongoing seafloor spreading between Madagascar and East Africa (which ceased at around 120 Ma; Seton et al., 2012), and (iii) the opening of the South Atlantic (since 132 Ma).

During the past 35-45 Ma, East Africa was moving northward relative to the East African plume (e.g. Ebinger and Sleep, 1998; Wichura et al., 2015), which resulted in regional doming, extensional tectonics and volcanism from the Turkana Divergence in northern Kenya to the North Tanzania Divergence (Fig. 2a; e.g., Morley et al., 1999). In the northern Kenya Rift, earliest extension began during the Paleocene-Eocene (Morley et al., 1992; Ebinger and Scholz, 2012). New thermo-chronological data from the Elgeyo Escarpment in Kenya’s central rift segment also reveal Paleocene-Eocene rift initiation, subsequent subsidence and heating, followed by renewed cooling and formation of major rift-bounding faults after 15 Ma (Torres Acosta et al., 2015). Along the Nguruman Escarpment of the southern Kenya rift, extensional faulting is shown to have started at approximately 7 Ma (Crossley, 1979). In contrast, farther south within the Tanzania Divergence, thermo-chronological data suggest that extensional faulting and cooling began during the Cretaceous and continued into the Paleocene-Eocene (Noble et al., 1997; Mbede, 2001).

Rifting is generally thought to have followed shortly after volcanism started in the different rift segments (e.g. Morley et al., 1992). The oldest volcanics in northernmost Kenya are as old as ≈39 45 Ma (e.g. Ebinger et al., 2000), while volcanism reached the intersection between the northern/central Kenya Rift and the Nyanza Rift (Fig. 2a) at ≈20 Ma (Pickford, 1982; Fitch et al., 1985) and the oldest volcanics in the southern Kenya Rift are between 20 and 16 Ma (Baker et al., 1972; Chapman et al., 1978; Smith, 1994; Hay et al., 1995). A recent synopsis on the onset of volcanic activity in East Africa by Michon (2015) suggests that earlier interpretations of a N S migration of volcanism and tectonic activity (Nyblade and Brazier, 2002) may not apply to the

Kenya Rift and that these processes were rather highly disparate in space and time in the EARS, as also suggested previously by Zeyen et al. (1997).

To summarise, western Kenya has strongly been affected by Cenozoic mantle dynamics as becomes evident from the high topography (as a result of doming; Fig. 1a) and the narrow basement lows (graben structures; Fig. 2a). In contrast, eastern Kenya shows low topographies (Fig. 1a) and considerably broader and deeper basins (Fig. 2a) that largely trace back to Mesozoic times. For western Kenya, the KRISP seismic experiments provide distributed information on deep crustal structures, while for eastern Kenya such information is confined to the south-easternmost parts of the proposed microcontinent Azania (Fig. 1b)."

Part 2:

RC1-4: At the beginning of part 2 some important information are missing such as the dimensions of the model and an accurate identification of the different density layers you are considering (for example, you should indicate that your modeled mantle has two layers, the first between Moho and 100 km deep and the second between 100 km deep and 200 km deep…).

As the model dimensions and resolution have been chosen according to the results of the analysis of diverse datasets as described in the revised chapters 3.1-3.3, we provide more details on the model setup in chapter "3.4 3D gravity modelling" of the revised manuscript.

"We have used the above constraints on the structure and density of the sedimentary and volcanic cover (Fig. 2, Table 1, Table 2), the crystalline crust (Fig. 3, 4), and the mantle (Fig. 5) to set up a starting 3D density model. This model spans 850 km in E W direction and 1100 km in N S direction (black rectangle in Fig. 1a). To model discrete density bodies, the corresponding scattered information on delineating structural interfaces has been interpolated to regular grids of 50×50 km horizontal resolution. For example, the initial depth to the top of the Basal Crustal Layer has been obtained through interpolation (and extrapolation) of the corresponding KRISP refraction seismic information (Fig. 4a; Appendix B) to cover the entire continental crustal domain of the study area. In the same way, we have generated regular 50×50 km-grids for all first-order model layers, i.e. gridded tops for all sedimentary and volcanic units (Tab. 1, 2), the Upper Crustal Layer (Fig. 2a), and the mantle (Moho; Fig. 3a). Accordingly, the vertical resolution of the crustal parts of the generated 3D density model is variable as it is determined by the thicknesses of the different units. This applies also to the upper mantle domain between the Moho and 100 km, modelled by six units, each showing a constant density as derived from P-wave velocities (Fig. 5a). The S-wave derived density configuration of the lower mantle domain reaching from 100 to 200 km depth, on the other hand, is represented by point-wise density information, i.e. by the generated voxel grid with a regular spacing of 50 km horizontally and 20 km vertically.

The constant densities assigned to the modelled sedimentary and volcanic units are presented in Table 2, those of the shallowest mantle in Figure 5a. For the starting density model we have further chosen ρ=2750 kg m 3 for the Upper Crustal Layer (cf. Fig. 4d) and ρ=3000 kg m 3 for the Basal Crustal Layer (cf. Fig. 4e), while the oceanic crust has been assigned a value of ρ=2900 kg m ^{3}. […]"

RC1-5: Line 21 (page 4) to line 17 (page 5): this part describing the basin formation would fit better in a "Geological setting" section.

We moved these paragraphs to the new chapter "2 Geological setting".

RC1-6: For the upper mantle density distribution (between Moho and 100 km deep): which data do you favor: KRISP or Achauer and Masson (2002)?

To clarify this, we have added the following sentence at the end of section 3.3.1 of the revised manuscript:

"Hence, we have chosen the overall larger density contrasts as indicated by the KRISP velocity profiles for the starting density model to be tested against the gravity field."

RC1-7: Line 11 (page 9): why "starting" model? Do you test different density models for the mantle?
Yes, we have tested different density configurations for the shallowest mantle (between the Moho and 100 km depth). An example is given in the second paragraph of section "5.5.1 Model sensitivity and robustness" of the original manuscript ("6.5.1" of the revised manuscript). There, we provide information on the calculated changes in the modelled gravity (50 mGal) corresponding to a decrease of the across-rift density contrast (from 25 kg m^{-3} to 10 kg m^{-3}). The starting density model is consistent with the KRISP-velocity-derived densities. We have decided to present this mantle configuration also as the final model, since the larger KRISP mantle density contrasts (compared to the smaller contrasts of the tomographic model of Achauer and Masson, 2002) produce a gravity anomaly large enough to keep the crustal densities very close to those directly derived from KRISP velocities using Eq. (1) (Fig. 4d, e).

RC1-8: Part 2.3.2: I understand the density is computed for depth between 100 and 200km. Are those density depth-averaged? This is not clear...
In chapter 2.3.2 of the original manuscript, we state that "we have complemented the model of Adams et al. (2012) towards the N and E by the model of Fishwick (2010) and 3D interpolated the scattered point information to obtain a voxel grid of regular spacing of 50 km horizontally and 20 km vertically." For the 3D gravity modelling, we have made use of this voxel (3D) grid. To clarify this point, we have added a sentence at the end of the first paragraph of section "3.4 3D gravity modelling" of the revised manuscript that reads:
"The S-wave derived density configuration of the lower mantle domain reaching from 100 to 200 km depth, on the other hand, is represented by point-wise density information, i.e. by the generated voxel grid with a regular spacing of 50 km horizontally and 20 km vertically."

Part 3.4:
RC1-9: I understand that the density and thickness of the upper crustal layer blocks and basal crustal layer blocks are constrained in order to best-fit the observed gravity. However, the way it is done is not clear. Is it done manually? Have you used a specific method? Are the blocks consistent with the five tectono-thermal crustal domains? What is the uncertainty on the crustal structure? Please be more accurate. The crustal structure is important for your later discussion.
We agree with the reviewer in that the crustal configuration as derived from the gravity-modelling step is of primary relevance for the whole discussion on the lithospheric strength, its zonation and inferred ideas on the rifting process. Therefore, we decided to include a more detailed description explaining the 3D gravity modelling procedure, by revising the last three paragraphs of section "3.4 3D gravity modelling" of the revised manuscript as summarized in what follows.
"It is important to note that the main focus of our study is to assess the density configuration of the continental crystalline crust across the whole study area. Therefore, we have only modified the starting 3D density model by varying this particular structural domain. Indeed, we have found that a reasonable fit between calculated and observed gravity can be obtained when keeping the density configurations of the sedimentary and volcanic cover as well as the mantle domains fixed (Fig. 6; Section 4).
In order to reproduce the observed long-wavelength variations in the gravity field, we have systematically modified the crustal 3D density configuration in our model. For this purpose, we have followed a "step-wise approach" relying on the IGMAS+ software capabilities. First, we have modified the topology of the top Basal Crustal Layer at locations not constrained by the

KRISP refraction lines in an attempt to arrive at a better agreement between calculated and observed gravity anomalies. We have followed a procedure in which we have varied (i.e. increased or decreased) the thickness of the Basal Crustal Layer along the selected 2D working sections while keeping track of the calculated gravity response of the model. It is worth mentioning that with these first imposed changes to the starting density model, we did not alter the thickness of the whole crustal layer; instead, any imposed variation in the Basal Layer thickness was complemented by respective variations in the thickness of the Upper Crustal Layer.

In a second stage, we have checked and confirmed (see Section 4) that a further improvement of the model fit on first-order gravity anomalies can be obtained through the implementation of the trends observed in the P-wave velocity configurations of the Upper and Basal Crustal Layers (Fig. 4d, e). This integration of lateral variations of density within both crustal layers was systematically done while interactively quantifying the gravity response of the whole model to each modification step. In a final step, the Moho topology has been adjusted in order to improve the fit between modelled and observed gravity anomalies, though limited to an area of small lateral extent where no gravity-independent constraints were available (hatched area; Fig. 3a)."

As we agree with all reviewers that the manuscript requires a more detailed discussion of the uncertainties of the modelling results, we have added the following five paragraphs to the revised section "6.1.1 Model sensitivity and robustness":

"In our gravity modelling approach we consider one single degree of freedom, which is the density configuration of the crystalline crust. However, given the relationship between two differently dense crustal layers and the resulting gravity response, the solution to our problem requires to take into account an additional free parameter, which is the depth of the top of the Basal Crustal Layer outlining the thickness variations of the two layers. For this purpose, we present the map of the obtained average crustal density (Fig. 7c) together with the thicknesses and densities of the two crustal bodies (Fig. 7a, b). While the average crystalline crustal density (as derived from the density and thickness configurations of the two crustal units) may be regarded as the more appropriate interpretation of the observed gravity anomalies across wide parts of the study area, it under-interprets the structural constraints provided by the KRISP profiles in western Kenya.

In the final 3D model, as constrained via the conversion of P-wave velocities and by gravity modelling, lateral variations in the density configuration are more reliable than absolute density values. This is because of uncertainties inherent in the density structure considered as the starting model. The most important determined trend, however, in terms of density gradients between western and eastern Kenya (Fig. 7c), is consistently mapped by both an eastward increase in the thickness of the relatively denser Basal Crustal Layer and by the lateral density variations of the two crustal units.

The quality of the final modelling results rely on the quality of the input data used to build up the starting 3D density model. Uncertainties associated with each dataset are, however, partly unknown (such as for the basement depth; Beicip, 1987), different in type (similar to the data), and are also transferred in a different manner to the 3D model (via interpolation, velocity-density conversion etc.). All of this hampers a quantification of uncertainties. It is also worth noting that any gravity-guided manual adjustment to the density configuration is subject to the modeller's decision. Therefore, there is an inevitable degree of non-uniqueness in the way density variations are partitioned. Although we have carried out all modifications in a systematic way, the modelling approach does not permit any straightforward quantitative assessment of related uncertainties with respect to the final 3D density configuration.

The five tectono-thermal domains that are proposed to represent surface expressions of a complex juxtaposition of interlocked crustal units (Fritz et al., 2013) have not been used as input

for the 3D modelling. Since most of the study area is covered by Mesozoic-Cenozoic sediments and volcanics, the spatial distributions of these five domains (Fig. 1b) and their geometrical continuation towards greater crustal depths have only been interpreted from scattered outcrop observations (including fault geometries; Fritz et al., 2013). Our seismic velocity- and gravity-guided 3D density model for the first time provides the basis for a joint interpretation of deep geophysics and surface geological observations concerning the configuration of the crust across the entire study area (see section 6.1.2).
To summarise, we present a 3D density model that is not only consistent with the observed gravity field, but also cross-checked with a wide spectrum of gravity-independent criteria and observations. The strength of our modelling approach thus stems from an efficient integration and usage of a large variety of different datasets. Furthermore, as already discussed above, the obtained trends in crustal density heterogeneities would have remained of the same order even if the density configurations of the sediments and mantle would have been implemented differently from what was done in this study, though still within the respective data constraints."

RC1-10: What's the reference density column for the gravity computation?
The reference (background) density applied for the calculations of the gravity response of the 3D model is 3250 kg m^{-3}. IGMAS+ calculates gravity anomalies by considering densities of the 3D model as density anomalies with respect to this overall reference density. Hence, we have chosen the value for the reference density to represent an overall average density of the entire modelled volume.
In the revised manuscript, we have added this information to section 3.4 (third paragraph).

Part 3:
RC1-11: The sentence: "the modeled thickness… … and the Moho geometry" (line 31-33 page 12) is not clear. Again, how are the different crustal blocks delineated? I understand those are constrained from KRISP data but how are they constrained away from the KRISP profiles?
We have rephrased and complemented this sentence for the revised manuscript:
"The modelled thickness anomalies of the Basal Crustal Layer differ significantly in wavelength (<150 km) and spatial distribution from both its internal segmentation into four regional density domains (Fig. 7b) and the Moho geometry (Fig. 3a). This demonstrates that it was possible to differentiate between thickness and density characteristics of this layer since they conform to different components of the observed gravity field."
As we detail in the final paragraph of chapter 3.4 (revised manuscript), we have used the gravity signal to model crustal densities for the regions away from the KRISP seismic profiles. Nevertheless, the two different crustal blocks of the Upper Crustal Layer are not only consistent with the gravity anomaly pattern that indicates larger masses in the east (Fig. 6a), but also with an abrupt eastward increase of velocities along KRISP lines D, E, and F (Fig. 4d). After having split the Upper Crustal Layer, we have obtained an intermediate status of the residual gravity that indicated (i) mass excess along and around the northern parts of the rift, consistent with relatively lower velocities in the Basal Crustal Layer as revealed by KRISP (Fig. 4e) and (ii) mass deficits in the southeastern parts of the study area, consistent with higher velocities in the KRISP Basal Crustal Layer there (Fig. 4e). Based on this intermediate result, we have split the Basal Crustal Layer into four parts, two of them showing different densities than the one characterising the Basal Crustal Layer in the starting 3D density model (3000 kg m^{-3}).
We have refrained from describing all the intermediate modelling steps and results in the manuscript, as they would not improve the reliability of the final model, at least not more than the demonstrated consistency between the final model and the different observables, such as gravity (Fig. 6), seismic velocity (Fig. 4d, e) and geology (Azania's western margin; Fig. 7a).

Part 4.2:

RC1-12: What is the error between the modeled and observed heat flow?

The first-order result of calculating the 3D conductive thermal field is a temperature configuration. The first-order measurement in heat-flow assessments is temperature, too (Nyblade et al., 1990; Wheildon et al., 1994). For this reason, we compare model results and observations in terms of temperatures (geothermal gradient differences; Fig. 8c). Comparing modelled and "measured" heat flow would mean to depart from the original modelling and measurement results, since in both cases heat flow is calculated based on Fourier's law, i.e. assuming a certain thermal conductivity as the coefficient to be multiplied with the (modelled or measured) thermal gradient.

Part 5.1.1:

RC1-13: The sentence "hence these local… … thickness maxima" (line 32 page 16 –line 2 page 17) is not clear.

Since the Moho depth presented by Woldetinsae (2005) is largely based on gravity modelling and due to its finer spatial resolution deviating locally from the global Moho model of Pasyanos et al. (2014), there is some uncertainty in fixing this important density discontinuity. The largest differences in Moho depth between the two models correspond with differences in gravity response that do not exceed 30 mGal. We have rephrased the following sentence for the revised manuscript:

"Hence, no matter which of these two models we had integrated, one of the main findings of this study would remain, namely that northeastern Kenya is regionally underlain by a lower crust of high density (ρ=3000 kg m^{-3}) with NW-SE oriented thickness maxima (Fig. 7b)."

Part 5.1.2:

RC1-14: Conclusions are drawn from the upper and basal crustal layer density distribution. However, the data constrains are poor. So a discussion on the uncertainty of the crustal density distribution would be interesting.

Here, we would like to refer to our comment to **RC1-11** about data constraints. Again, although the overall distribution of differently dense crustal domains is mainly gravity constrained, the existence of major contrasts in seismic velocity confirms these modelled density contrasts. Concerning the uncertainties of the modelling results (crustal density configuration), we would like to refer to our comment to **RC1-9**.

RC1-15: Line 5 (page 18): you interpret mafic rocks below the Kenya rift though it is low Vp and low density. Could it be something else? What are the reasons to interpret this as mafic rock despite low Vp and density?

With a density of 2920 kg m^{-3} below the northern Kenya Rift, the Basal Crustal Layer does indicate mafic rocks there (given that gabbroic rocks typically show densities of 2900 kg m^{-3}, for instance). In the first paragraph of the sub-section "Basal Crustal Layer" in chapter 5.1.2 we refer to a number of other studies that have previously interpreted the seismically constrained Basal Crustal Layer as representing mafic intrusions related to the rifting process. At the end of this paragraph, we provide a possible explanation for the density/velocity to be low with respect to the values in the remaining study area (3000-3050 kg m^{-3}): they might be low "due to elevated mantle and crustal temperatures (cf. Fig. 5c, 8a) and related thermal expansion of the rocks (see also e.g. Maguire et al., 1994)".

RC1-16: Again, it is not clear how the thickness of the basal crustal layer is constrained away from the KRISP profiles though your interpretation is based a lot on this result.

Again, we would like to refer to our comment to **RC1-11** about data constraints.

Part 5.2.1:
RC1-17: What are the depths of observed seismic peaks at the various points? This could be useful to include those peaks on the YSE profiles of fig. 9.
For points E-H (Fig. 10a), Table 5 relates the “Modelled depth of the top of the brittle-ductile transition” (for the crust and mantle, respectively) to the observed “Depths of peak seismicity”. For the locations of the YSE profiles, we do not have corresponding observations on seismicity. The locations of the YSE profiles have been chosen to span a structural profile across the Kenya Rift as illustrated by Figure 9.

RC1-18: Line 11-21 (page 23). The discussion on the plume-lithosphere interaction is not clear. That would be useful to indicate the location of plume impingement on a figure (figure 10 a for example).The link between the plume and strain localization is not clear. What is the link between the plume emplaced beneath a compositionally heterogeneous crust and the focusing of crustal thinning within the Southward tapering Arabian Nubian Shield?
The plume discussed here refers to a mantle thermal anomaly observed at the present day that reaches far beyond the limits of the modelled area. It is also known as the East African Superplume and was used by Koptev et al. (2015) to test the hypothesis that a mantle plume starting to rise beneath the Tanzania Craton may have been responsible for extensional tectonics and volcanism along the eastern branch of the East African Rift. The observed smaller-scale thermal anomaly underneath the Kenya Rift might be a derivative of this superplume and is exemplarily shown by the depth of the 1350°C-isotherm as derived from S-wave velocities which will give some indication on the plume impingement domain in the study area (Fig. 8a).

To clarify our discussion about plume-lithosphere interactions, we have rephrased and complemented the third paragraph of section 6.2.2 of the revised manuscript:
“The presented 3D model is the first to jointly integrate the present-day mantle thermal anomaly, crustal composition and related strength variations within the crust and lithospheric mantle. This opens the possibility for new hypotheses on plume-lithosphere interactions, i.e. on how dynamic mantle buoyancy forces contributed to tensional stresses in the lithosphere and how the latter responded. According to the 3D model, the plume-related lithospheric thinning would have been taking place beneath a compositionally and rheologically heterogeneous crust (Fig. 7; Table 4) – even though its structural configuration and, above all, its thermal state have certainly not been the same in the past. Crustal thinning obviously focussed within the southward tapering Arabian Nubian Shield (Fig. 1b) as the easternmost part of the rheologically weaker domain of western Kenya (Fig. 10a, b). At the same time, the configuration of eastern Kenya comprising Azania upper crust and remarkably thick, dense, and stiff lower crustal rocks (Fig. 7) might have formed a strong barrier against crustal deformation. Hence, strain localisation (induced by mantle dynamics and related tensional stresses) would have been facilitated by pre-existing contrasts in rheological properties between western and eastern Kenya.”

RC1-19: Line 9 page 24: it looks volcanism is offset towards western (and not eastern) boundary on the fig. 10a in the northern Kenya rift.
This is true. We have rephrased the sentence to be more precise about the area we want to discuss:
“In the central Kenya Rift (i.e. just north of the Kenya Rift-Nyanza Trough junction; cf. Fig. 2a), this narrow zone of strongest crustal thinning and volcanism is locally offset from the rift centre towards the eastern boundary of the surface expression of the rift (Fig. 10b).”

Technical corrections
RC1-20: Line 4 (page 5): "earliest extension" is written twice.
Corrected.

RC1-21: Line 4 (page 5): "Paleo-Eocene" instead of Paloecene.
We have changed "Paleo-Eocene" to "Paleocene-Eocene".

RC1-22: Figures: please add a title for each figure for clarity.
We have rephrased the legend descriptions so that they function as titles.

RC1-23: Please add the names of the main structural features when it is relevant (Anza basin, Turkana Trough, etc…).
Done.

RC1-24: Figure 5a: what is the grey square?
The grey square indicated the XY-dimensions of the tomographic P-wave velocity model of Achauer and Masson (2002), which is indicated in the figure caption.

RC1-25: Indicate on fig 10a the location of plume impingement.
We have plotted the 125 km-contour of the 1350°C-isotherm in Fig. 10a to illustrate where the "plume" reaches shallow depth levels.

Some comments on the references:
RC1-26: ref. Allen and Allen 2009 or 2013?
In the text and references we refer to the latest version (Allen and Allen, 2013).

RC1-27: ref. Baker, Mitchell and Williams 1988 is missing in the text.
Deleted from the references list.

RC1-28: ref. Cacace et al. (2016): isn't it Cacace and Scheck-Wenderoth (2016)?
Corrected.

RC1-29: ref. Melnick et al. (2015). in the reference list it is 2012.
Corrected.

RC1-30: the ref. Morley et al. (1999) (line 27 page 2) is missing in the reference list.
In the text it should read as "Morley (1999)" as in the reference list. Corrected.

RC1-31: ref. Strecker et al. (1990) is missing.
Deleted from the reference list.

RC1-32: ref. Turcotte and Scubert (2014) is missing.
Added to the reference list.

RC1-33: ref. Goetze (1978) is missing.
Corrected to Goetze and Poirier (1978).

RC1-34: ref Onuonga et al. (1997) is missing in the reference list.
Added.

RC1-35: ref. Halls et al. (1987) is missing in the reference list.
Added.

RC1-36: ref. Burov (2011) is missing in the reference list.
Added.

RC1-37: ref. Catuneanu et al. (2005) is missing in the reference list.
Added.

RC1-38: ref. Seton et al. (2012) is missing ion the reference list.
Added.

RC1-39: ref. Fuchs et al. (2013) is missing in the reference list.
Added.