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# Some remarks on the structure and geodynamics of the Kenya Rift

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ABSTRACT

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Since 1985 the KRISP working group has recorded a number of seismic profiles and carried out array experiments in the Kenya Rift valley. Long range profiling has shown that across the Rift at latitude 0.5°, the crust thins from 40 km beneath the western flank to 30 km beneath the graben proper, and thickens again to 35 km beneath the eastern flank. The crustal thickness beneath the graben decreases northwards from 35 km beneath the culmination of the Kenya dome to 20 km under Lake Turkana, suggesting a change to a highly extended terrain in the northern Kenya Rift. Furthermore, the anomalous low mantle velocities of 7.5–7.6 km/s suggest that the upper mantle immediately beneath the Rift may contain reservoirs of magmas, which were generated at greater depth. The anomalous mantle under the graben proper, as seen by the tomographic imaging, is characterized by a large, steep walled low velocity zone with a velocity decrease of at least 10% in the central part between 65 and 100 km depth which reaches down to more than 150 km. A comparison of the shape and size of the observed Bouguer anomaly with the LVZ suggests that the upper mantle beneath the Rift probably includes some 5% partial melt, concentrated in the areas with lowest velocities. These results provide new constraints for geodynamic models of the evolution of the Kenya Rift.

## 1. Introduction

The Kenya Rift (Fig. 1a) is perhaps the most striking of the Cenozoic Rifts. It forms part of the East African Rift System which comprises a series of rift zones which extend over a distance of approximately 3200 km from the Afar triple junction at the Red Sea–Gulf of Aden intersection to the Zambesi River in southern Africa. This rift system bifurcates around the Archean Nyanza craton which coincides with the uplifted East African Plateau. The Kenya Rift transects the Kenya dome, which itself is superimposed on the

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eastern margin of the East African Plateau. Between latitudes 2°N and 2°S the Rift, whose floor reaches an elevation of up to 1800 m, is characterized by well defined 50–70 km wide half grabens; this breakaway fault scarps have a topographic expression of up to 1600 m from the floor to the shoulder of the Rift. At its northern and southern extremities, the elevation of the Rift floor falls below 600 m and the fault structures splay out over widths of 200 km or more, whereby the Rift loses its graben like appearance. It is associated with some 144,000 km$^3$ of volcanics erupted since the early Miocene (Williams, 1972). At its northern extent, in Lake Turkana, the Rift intersects the Jurassic Anza graben, and it is located close to the margin of the Archean Nyanza craton and the Pan-African Mozambique shear belt along its western flank. Its precise location

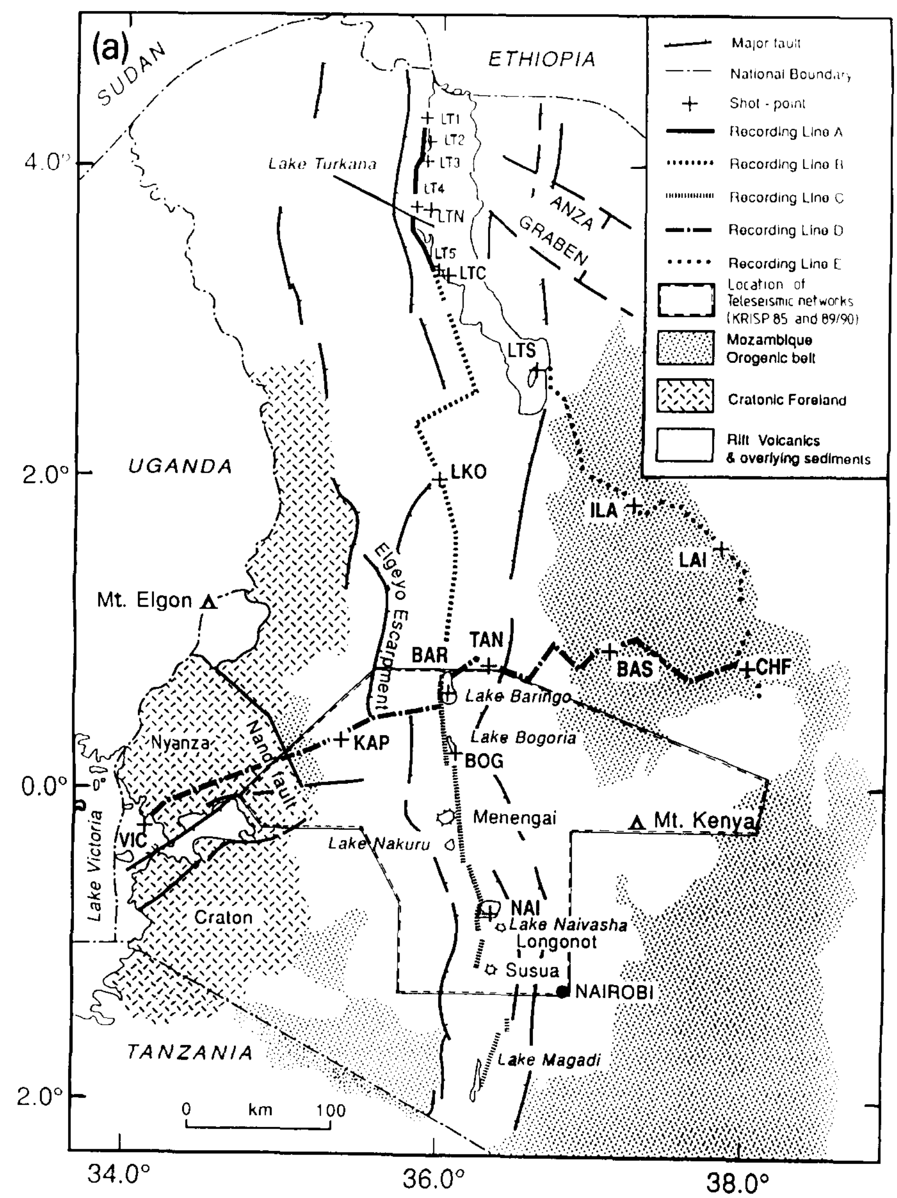


Fig. 1. (a) Tectonic map of Kenya including the location of the refraction profiles as well as the teleseismic arrays of KRISP85 and KRISP89/90. (b) Map showing the station locations of the 1985 and 1990 KRISP teleseismic programs including the proposed accommodation zones (Bosworth et al., 1986).

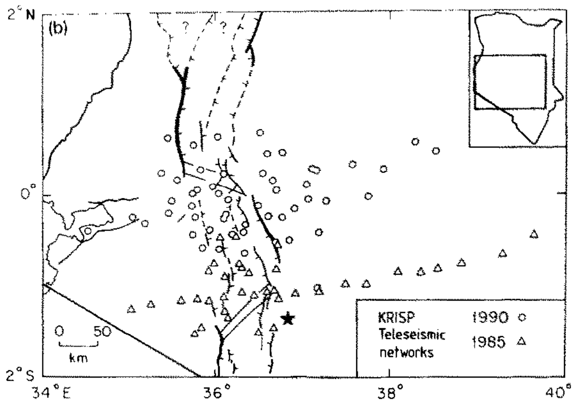

Fig. 1 (continued).

may have been controlled by pre-existing Precambrian structures of the crystalline basement (Fairhead and Stuart, 1982).

The results of gravity and teleseismic research indicate that the Rift is underlain by low density mantle material, suggesting an upward perturbation of the lithosphere–asthenosphere boundary. Beneath the Kenya Rift, partially molten material may have ascended, close to the base of the crust and even may have penetrated the crust. Studies of the crust and underlying upper mantle, using gravity, electromagnetic, earthquake and a small number of limited explosion seismic experiments, have resulted in a wide range of structural models. A number of these suggest the presence of a massive, rift wide, mantle derived intrusion penetrating close to the surface along the axis of the Rift (Baker and Wohlenberg, 1971; Mohr, 1987). More recently, results from long range explosion seismic experiments (Henry et al., 1990; KRISP Working Group, 1991a) suggest the presence of limited lower crustal intrusions, with dykes and magma chambers penetrating the upper crust along the Rift.

## 2. Crustal structure and kinematics

To develop an understanding of the geodynamic processes affecting the Kenya Rift, two different lines of approach can be followed. The first examines the response of the brittle upper crust via structural and seismic studies and attempts to relate its response to the regional stress system. The second approach seeks to define the large scale structure and physical state of the lithosphere–asthenosphere system in the area of the Rift axis and to infer the geodynamic processes which caused the evolution of this rift system.

The variation in upper crustal structure, both across and along the Rift, provides information on the response of the brittle upper crust and the ductile lower crustal layer to extensional forces. Structural complexities may result from a non-uniform stress field and the heterogeneity of the crust in Kenya. The observed variations in structure include the presence of half grabens of varying sizes and polarities, associated accommodation zones (King, 1978; Bosworth et al., 1986), and the differing displacements and dips on the major faults within the Rift. The presence of intrusions, involving variation in the crustal thermal regime, will also have a significant effect on the response of the Kenya crust to extension. Detailed examination of Rift structures via surface mapping, reflection seismic surveys and the use of remote sensing data permits to draw conclusions on the crustal extensional process. For example, based on fault slip inversion, Strecker and Bosworth (1991) have recently suggested that the minimum compressive stress in East Africa was reoriented from E–W to NW–SE since the Mid-Pleistocene and that it is now obliquely oriented to the strike of the Rift, whereby its bounding faults have become progressively inactive in terms of extension.

Seismic activity will also reflect the response of the brittle upper crust to the present day stress field. There has been a small number of local seismicity studies within the Rift. The most detailed (Young et al., 1991) indicates that within the Rift, in the vicinity of Lake Bogoria, the brittle–ductile transition zone is located at a depth of approximately 12 km. This study also suggested that the bulk of the seismicity appears to be associated with the older major graben bounding faults of the Rift rather than with the young near surface "grid" faulting. A study of the 1928 Subukia magnitude 6 earthquake in the same region (Ambraseys, 1991) suggests that the majority of the strain occurring within the Rift is taken up along faults associated with such large magnitude events. An event of this magnitude probably occurred on a fault reaching down to the brittle–ductile transition. Such work provides necessary information to resolve the kinematics of crustal blocks in the study area. In order to obtain a better understanding of the kinematics of the entire Kenya Rift it will be necessary to analyze local seismicity and in particular the larger magnitude events throughout the length and breadth of the Rift zone. Such data is not yet available.

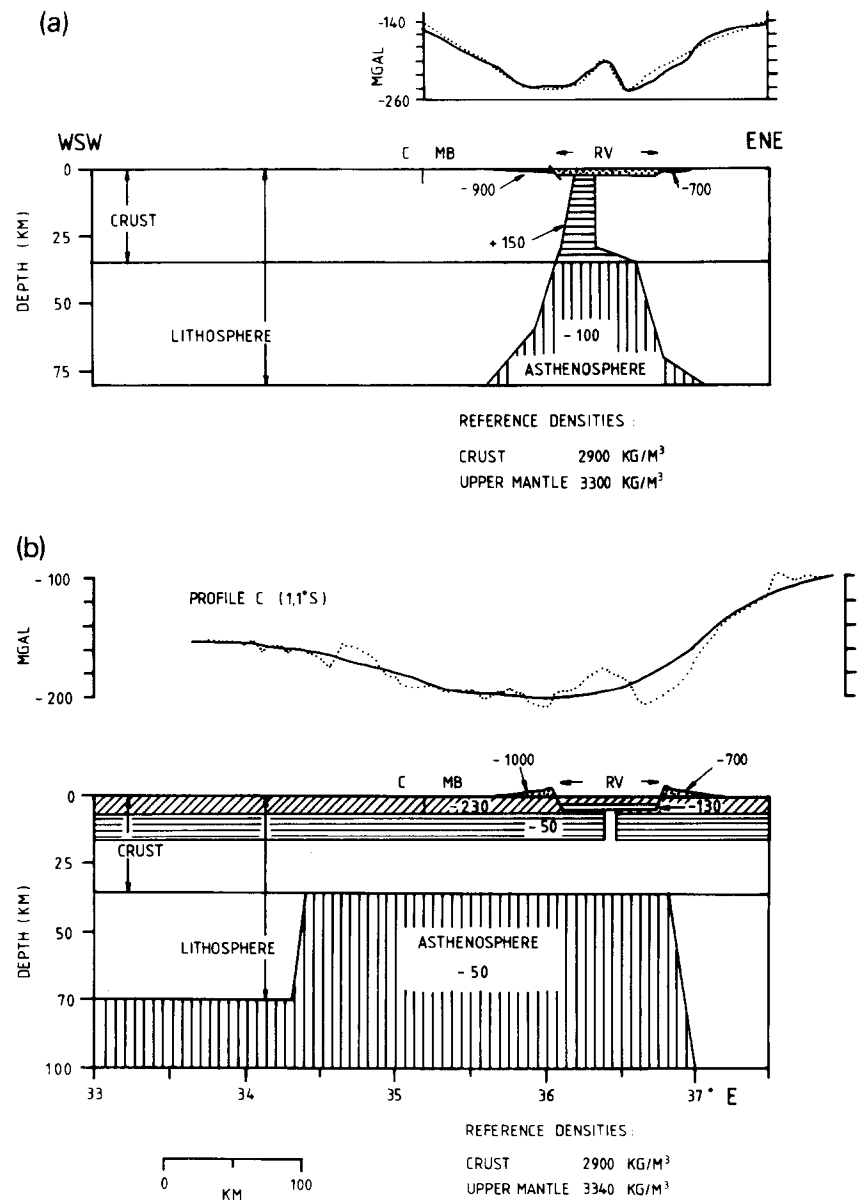


Fig. 2. Published gravity models for the Gregory Rift Valley. (a) After Baker and Wohlenberg (1972) at about the equator, showing a huge intrusion of mantle material close to the surface, a low density sheet at the surface and lithospheric thinning localized beneath the rift. *RV* = rift valley; *MB* = Mozambique belt; *C* = Nyanza craton. Numbers give density contrast in $kg/m^3$. (b) After Fairhead (1976) at latitude 1.1°S showing crustal dyke intrusion and lithospheric thinning beneath the rift valley (*RV*), and beneath the Mozambique belt (*MB*) and Nyanza craton (*C*) to the west of the rift.

## 3. Lithospheric structure

The second approach towards a further understanding of the geodynamic processes governing the evolution of the Kenya Rift involves the analysis of the structure and physical state of the lithosphere–asthenosphere system. Recent deep seismic studies, involving both explosion and earthquake seismology, permit the development of a crustal and upper mantle model of the Rift and focus attention on the degree of perturbation of the lithosphere-asthenosphere boundary, the nature of the sub-crustal material, and the amount of extension across the Rift. The new model incorporates results of previous seismic studies as well as gravity data.

### *3.1. Gravity studies*

Since the classic work of Bullard (1936), who used gravity data to investigate isostasy in East Africa, several authors (e.g., Sowerbutts, 1969; Searle, 1970; Baker and Wohlenberg, 1971; Darracott, 1972; Fairhead, 1976; Banks and Swain, 1978; Swain et al., 1981) have investigated the available gravity data (Khan and Swain, 1978). A long wavelength negative Bouguer anomaly with a width of 350 km and an amplitude of 170–180 mGal is centred over the rift valley; superimposed on it is a local positive, 40–60 km wide, 30–50 mGal amplitude anomaly, which coincides with the axis of the Rift. Several interpretations of the observed anomalies have been made in terms of low density material in the upper mantle (e.g., Searle, 1970; Baker and Wohlenberg, 1971; Fairhead, 1976; Banks and Swain, 1978) and intrusion of mantle-derived material into the crust (e.g., Searle, 1970; Darracott et al., 1972; Fairhead, 1976; Swain et al., 1981), suggesting widely varying dynamic models. Although it is implicit in the proposed density models that the anomalous mantle material is involved in supporting the topography of the Rift-dome centred on the Rift valley, only some of the authors have used as an explicit constraint the assumption that the density distribution satisfies some mechanism of isostatic compensation (Banks and Swain, 1978; Bechtel et al., 1987; Ebinger et al., 1989).

#### *3.1.1. Crustal structure from gravity*

Previous interpretations of the local gravity maximum, which centred on the rift axis, involve some 16–28 km wide dense intrusion of mantle-derived material (Fig. 2a; Searle, 1970; Baker and Wohlenberg, 1971) or a narrower about 10 km wide single dyke (Fig. 2b; Fairhead, 1976). Fairhead stated that also a massive sill-like intrusion in the rift crust, penetrating the crust to just beneath the surface, could equally well satisfy the gravity constraints. More recent work by Swain et al. (1981) and Swain (1992) suggests that only a modest increase in basement density, provided it is sufficiently extensive at depth, is able to satisfy the observed gravity data. Swain (1992) concludes that the densities of 2750–2760 kg/m$^3$ implied for the basement are consistent with the range of seismic velocities of 5.95–6.1 km/s measured along the rift axis during the KRISP85 experiment. Moreover, the results of the KRISP85 survey did not provide support for massive intrusion of mantle derived melts into the crust, as proposed by Searle (1970) and others.

#### *3.1.2. Mantle structure from gravity*

All authors explain the observed long wavelength Bouguer minimum, which is associated with the rift axis, by a low-density body located in the upper mantle, though with different shape, size and density contrast compared with the flanks. Baker and Wohlenberg (1971; Fig. 2a) assumed that the low density region of lithospheric thinning is confined to a narrow zone beneath the Rift and a near surface continuous low density sheet to account for the observed broad negative anomaly. In contrast, Fairhead (1976; Fig. 2b) proposes that a broad low density structure extends more than 100 km to the west of the Rift and replaces a normal lithospheric mantle. In Baker and Wohlenberg's model, mantle derived material intrudes the crust and reaches almost to the surface in the form of a dyke, in order to account for the observed local gravity maximum along the rift axis.

Recently, Nyblade and Pollack (1990) proposed a very different model for the observed anomaly. They argue that previous interpretations of the gravity field in this region did not take into consideration that the Rift is located in the area of a Pan-African suture zone and that the observed anomaly is possibly composed of two components, namely a Rift- and a suture-related one:

The 'rift' related component is the effect of a shallow, sediment filled rift basin, a lower crustal intrusion and a narrow low density zone in the upper mantle lithosphere beneath the rift axis.

The 'suture' component arises from a crustal root along the boundary between the Pan-African Mozambique belt to the east and the Precambrian Nyanza craton to the west, whereby the former is characterized by a higher density crust above part of the crustal root.

Nyblade and Pollack (1990) claim that this kind of paired gravity anomaly is commonly associated with continent-continent collision zones (e.g. Gibbs and Thomas, 1976) and conclude that, if a strong 'suture' signature contributes to the broad negative anomaly over this region, existing models may overestimate the amount of lithospheric thinning caused either by thermal processes or ductile lateral mass transfer during the Tertiary rifting episode.

### 3.2. Recent explosion seismic studies

Crustal and upper mantle P-wave velocity models for the Kenya Rift are based on a limited number of explosion seismic studies. The KRISP90 experiment (KRISP Working Group, 1991a) was carried out in 1990 as a follow up to earlier studies (Griffiths et al., 1971; Swain et al., 1981; Henry et al., 1990). Three seismic profiles were recorded, one along the axis of the Rift from Lake Turkana in the north to Lake Magadi in the south, the second across the Rift in the vicinity of Lake Baringo just to the north of the apex of the Kenya dome, and the third on the flanks of the Rift to the SE of Lake Turkana (Fig. 1a).

During this experiment, which was undertaken by an international group of seismologists from Kenya, Europe and the USA, over 7000 seismograms were obtained from 34 shots in both lakes and boreholes. Results of this survey indicate that:

(1) Beneath the sedimentary and volcanic layers, which reach a maximum thickness of about 5.5 km within the Rift, the crust consists of a layer of velocity 5.9–6.2 km/s above a layer of velocity 6.4–6.6 km/s which in turn overlies a higher velocity zone (of about 6.8 km/s) immediately above the Moho. To the west of the Nandi fault zone, which marks the boundary between the Archean Nyanza craton and the Pan-African Mozambique orogenic belt, the upper crustal velocity (5.9 km/s) is marginally lower than that (6.0 km/s) of the Mozambique belt (Fig. 3a). Beneath the axis of the Rift, the upper crustal layer has a slightly higher velocity (6.1 km/s) than beneath the rift margins. The extensive Cenozoic volcanism associated with the Rift suggests this increase could be caused by the presence of basic dykes. The velocity increase at about 10–15 km beneath each line almost certainly marks a change to a more basic average crustal composition. A velocity of about 6.8 km/s has been identified at the base of the crust on all three lines, except in the northern part of the axial profile. However, for this region the model resolution does not preclude the presence of an up to 2 km thick layer of 6.8 km/s within the lowermost crust. There is a marked difference in the reflectivity of this boundary beneath the

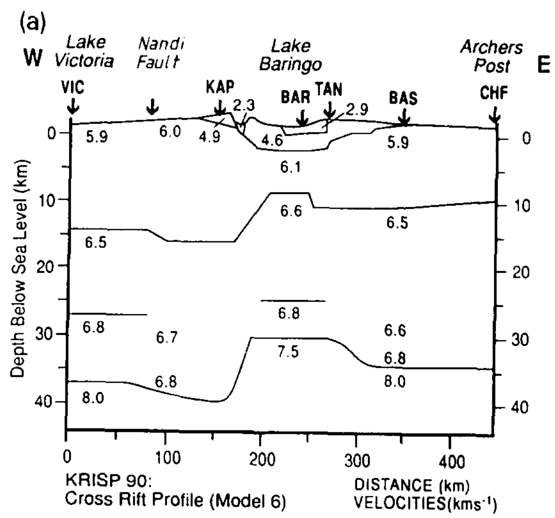


Fig. 3. Crustal and uppermost mantle velocity and interface model for (a) the cross profile, line D, and (b) the axial profile as obtained from raytracing (taken from KRISP Working Group, 1991a).

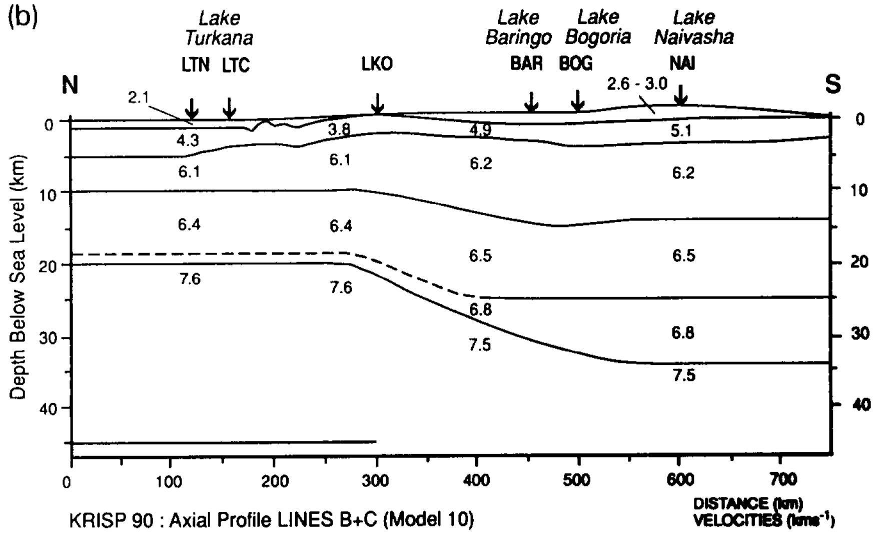


Fig. 3 (continued).

Nyanza craton and the Mozambique belt. This suggests that the latter is characterised by a gradual increase in mafic material towards the base of the crust whereas the Nyanza craton is characterised by a well defined boundary separating a lower crustal mafic layer from the mid-crustal layer. Beneath the southern part of the Rift, the 6.8 km/s layer again has a well defined upper boundary. Assuming that the material is the same as that observed elsewhere within the Mozambique belt, but is intruded by Cenozoic mafic material, and allowing for an increase in temperature of 300–500°C across the Rift margins at this level, the equivalent velocity, assuming a standard continental geotherm, would be about 7.0–7.1 km/s (Christensen, 1979). Comparison of this value with those observed beneath older Phanerozoic continental rifts (Mooney et al., 1983) and passive margins (White et al., 1987) suggests the volume of intruded material is smaller beneath the Kenya Rift.

(2) The Rift is associated with a marked thinning of the crust. In the vicinity of Lake Baringo the crustal configuration of the Rift is asymmetric in cross-section, whereby the crust is thicker beneath its western flank than beneath the eastern flank. Furthermore, the crust immediately to the west of the Rift margin is thickened as compared to the crust of the Nyanza craton. This is consistent with the Moho depths predicted by the recent re-evaluation of the available gravity data (Nyblade and Pollack, 1990).

(3) Beneath the Rift, the subcrustal material has an anomalous velocity of 7.5–7.7 km/s compared with the normal sub-Moho material beneath the Rift flanks which has a velocity of 8.0–8.1 km/s. From heat flow values (Morgan, 1982) and without wholesale melting of the lower crust the temperature at the base of the crust, cannot be more than about 1000°C. Under these conditions and at a depth of 35 km, the velocity of peridotite would be about 7.7 km/s. The sub-Moho velocities of about 7.5 km/s beneath the Rift suggests this zone contains pyroxene enriched, olivine depleted material and/or molten magma generated at greater depths.

(4) The crust thins substantially to the north from 35 km beneath the apex of the Kenya dome in the vicinity of Lake Naivasha to 20 km beneath Lake Turkana (Fig. 2b). This variation in crustal thickness correlates with a northward widening of the Rift and with decreasing topographic elevation.

### 3.3. Teleseismic studies

The analysis of teleseismic earthquake waves propagating through the upper mantle provides the best constraints on the deep structure of the lithosphere–asthenosphere system beneath the

Kenya Rift and its vicinity. Earlier studies of surface wave dispersion (e.g. Gumper and Pomeroy, 1970; Knopoff and Schlue, 1972), surface wave tomography (Nolet and Mueller, 1982) and array seismic studies (e.g. Long and Backhouse, 1976; Maguire and Long, 1976) indicated that the upper mantle beneath the Rift is characterized by anomalously low P- and S-velocities. However, these studies lacked either resolution or defined only very local velocity structures. Re-

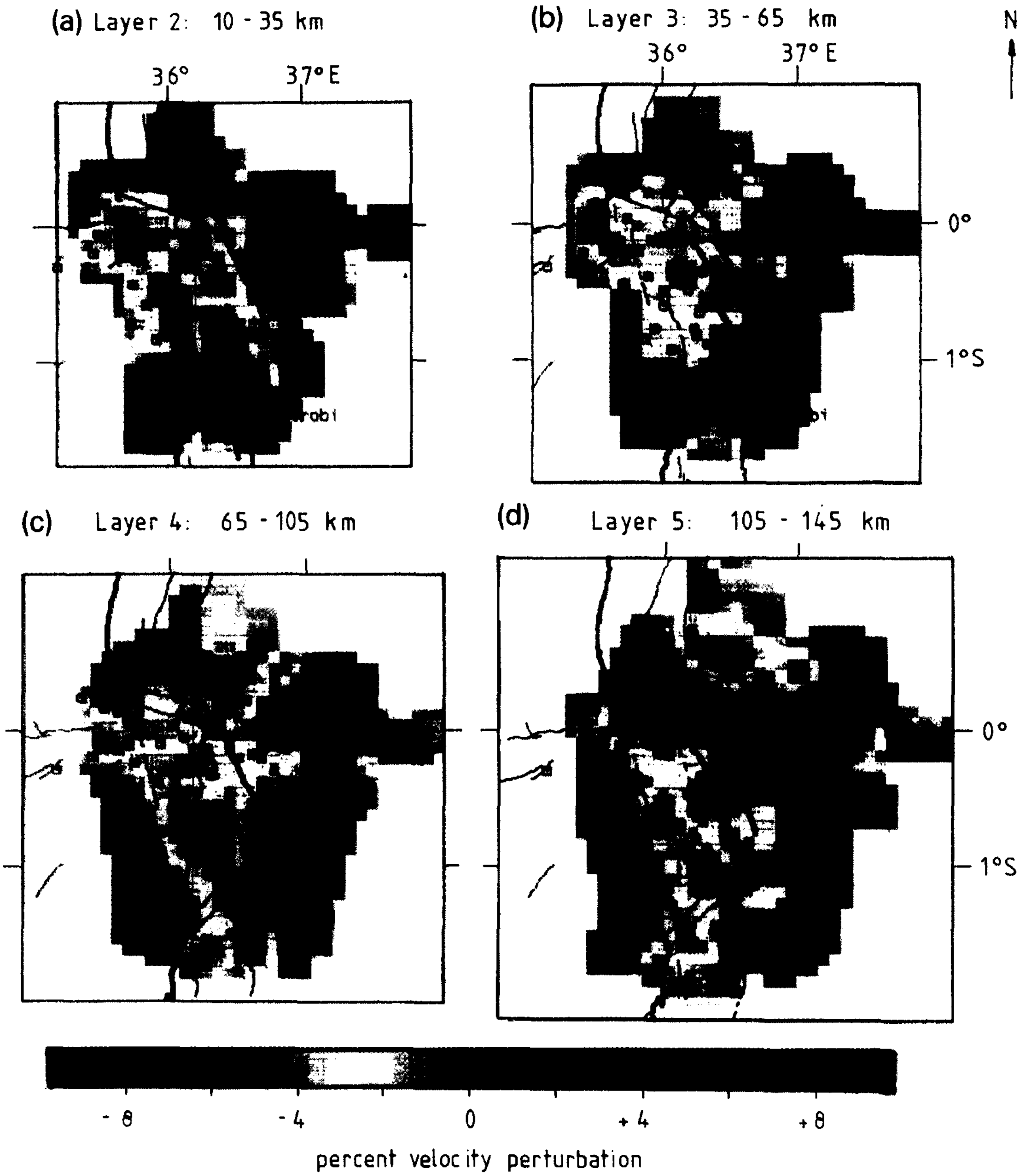


Fig. 4. Velocity distribution for the crust and upper mantle in the Kenya Rift valley from teleseismic delay time tomography (taken from KRISP Working Group, 1992). Colors indicate percent velocity perturbations, i.e. differences from average layer velocities. Red, orange and yellow are slower than average, green is near average and blue, indigo and violet are faster. The box outlines the layer, including unmodeled blocks, which are colorless. The star indicates Nairobi and small squares are station locations. (a) lower crust 10 to 35 km; (b) uppermost mantle 35 to 65 km; (c) second mantle layer 65 to 105 km; (d) third mantle layer 105 to 145 km.

cently, two teleseismic profiles across the rift (Savage and Long, 1985; Dahlheim et al., 1989) defined in more detail the low-velocity zone (LVZ) in the upper mantle. Both studies found that the LVZ extends to a depth of at least 200 km and that in addition a secondary low-velocity feature exists some 150 km east of the Rift axis. Dahlheim et al. (1989) suggested that the anomalous zone under the Rift proper contains about 5% partial melt and extends possibly up to the Moho. It is likely that the lithosphere–asthenosphere system is characterized by much more complex structures than previously anticipated.

Two seismic tomography studies (Green et al., 1991; KRISP Working Group, 1992), which were carried out in conjunction with the KRISP refraction profiles, provide a more detailed view of the upper mantle structure. For the first time the 3-D velocity structure of the Rift and its shoulders could be imaged on a regional scale down to about 150 km depth. Using a modified inversion algorithm (Achauer et al., 1986), based on the pioneering work of Aki et al. (1977), velocity information was obtained from the traveltime residuals observed by the seismographs of the teleseismic arrays. Two crustal and three mantle layers could be modelled of which layers 2 to 5 are shown in Fig. 4 . For the two crustal layers and the uppermost mantle layer the starting velocities were taken from the 1985 KRISP refraction study (Henry et al., 1990), while for the deepest two layers an average upper mantle velocity was used. Results of these studies are summarized below:

Layer 1 (not shown): The upper crustal layer (surface–10 km) cannot be discussed quantitatively because only the velocity structure directly beneath the recording sites is sampled. Nevertheless it exhibits strong lateral variations probably due to the presence of sediments and volcanics close to the surface in the Rift basins.

Layer 2: The lower crustal layer (10–35 km; Fig. 4a) shows a 12% velocity range (approximately 0.8 km/s) with high velocities beneath the rift axis relative to low velocities beneath the flanks. Along the rift axis significant velocity variations are observed whereby the zone of low velocities is limited to the north and south by accommodation zones, evident in the fault pattern of the Rift (Fig. 1b; King, 1978; Bosworth et al., 1986).

Layer 3: The uppermost mantle layer (35–65 km; Fig. 4b) shows low velocities beneath the Rift, having a range of 10%. To the north the low velocity area appears to terminate at the northern accommodation zone. However, it is evident that it extends in the direction of the young Nyanza Trough.

Layer 4: The intermediate mantle layer (65–105 km; Fig. 4c) shows that a pronounced relative low velocity body occurs beneath the rift axis along the entire Rift. The velocities have a 12.5% range (approximately 1.0 km/s) in the south and about 8% in the north. Only small variations can be seen from north to south, excluding the northernmost end of the modelled Rift, where only slightly reduced velocities are evident. The low-velocity zone correlates remarkably with the bounding faults of the Rift; only a slight deflection under the eastern rift-shoulder is evident north of the northern accommodation zone.

Layer 5: The 'deepest' mantle layer (105–145 km; Fig. 4d) is similar to layer 4 but has a broader anomalous zone and a reduced velocity range of about 9% (approximately 0.7 km/s). The deflection of the low-velocity zone under the eastern shoulder is more pronounced.

How can the 12% velocity range in the lower crust be explained? According to the results of the KRISP90 profile following the rift axis (Fig. 3b), high velocity material of 6.8–6.9 km/s is present at the base of the crust beneath the rift axis. Its layer thins to below the thickness of resolution between Lokori (LKO) and Lake Baringo (BAR) in the north but thickens to the south to a maximum of 10 km. This suggests the possibility that mantle derived material intruded along a narrow zone into the lower crust; this material may represent the residue of differentiated magmas which were extruded at the surface. However, there is no apparent correlation between the distribution of the central volcanoes and the observed lower crust velocity anomaly.

What could be the cause of the 10–12% velocity range in the upper mantle? Differences in lithologies are unlikely to account for the observed velocity changes. An increase in temperature would decrease the velocity but also the density. The Bouguer gravity anomaly derived from a velocity model based on the UCLA–Karlsruhe teleseismic data (Dahlheim et al., 1989) and the Nafe–Drake P-velocity–density relation, is twice as large as the observed Bouguer anomaly (Achauer, 1990). Partial melt, however, can have large effects on seismic velocities but only small effects on the density (Mavko, 1980). Laboratory measurements on dry peridotite at pressures of up to 1 GPa and temperatures of 1300°C show that 6–8% partial melt can cause the observed 12% velocity decrease (Sato et al., 1988). Green et al. (1991) therefore suggest that there are regions of up to 5–6% partial melt in the upper mantle under the Rift proper where extensive low velocities are observed. Since the resolution of the data is only a cuboid with dimensions of $20 \times 20 \times 30$ km, there is a trade-off between thickness and percentage-amount of the partial melt zones. This renders it difficult to estimate the depth to the top of the partial melt zone. However, the change in the velocity perturbation in layer 3 suggests that the top of this zone varies in depth along the rift axis. Also the correlation between the rift bounding faults and the observed low-velocity zone in layer 4 (Fig. 4c) suggests that the greatest degree of lithospheric thinning is directly beneath the Rift and that this zone of lithospheric thinning broadens only at depths greater than 100 km (Fig. 4d).

## 4. Implications concerning the geodynamics of the Kenya Rift

Recent geophysical studies have yielded important results about the variation in crustal thickness along the axis of the Kenya Rift, about magmatism and extension and the subcrustal velocity distribution.

### *4.1. Variation in crustal thickness*

Results of the explosion seismic project KRISP90 indicate that the crust has been extended by a factor of about 1.2 across the apex of the Kenya dome and about 1.8 at the northern end of the Rift beneath Lake Turkana. Due to the lack of knowledge of the age of the lower part of the thick sequence of sediments beneath Lake Turkana (which could be Jurassic and associated with the intersecting Anza graben), the estimate for the northern part is likely to be a maximum value. If the material of velocity 6.8 km/s observed at the base of the crust beneath the southern part of the Rift involves some mafic intrusion into the lower crust, the extension factor estimate for the southern part of the Rift is likely to be a minimum value. Due to the location of the flank line, which lies between the Jurassic Anza graben and the Tertiary Kenya Rift to the south of Lokori (LKO in Fig. 1a), it can safely be said that crustal thinning along the Kenya Rift with respect to that beneath the flank line must have been caused by Tertiary rifting.

The northwards crustal thinning along the Rift, together with its apparent splaying to the north of Lake Baringo, appears to be related to a change to an increasingly extended terrain in the northern Kenya Rift which may ultimately grade into the 150 km wide 'Gofa basin-and-range province' of Western Ethiopia (Moore and Davidson, 1978).

### *4.2. Magmatism and extension*

White and McKenzie (1989) relate the quantity of magmatic material generated during the rifting process to the amount of stretching and the temperature of the mantle beneath the extended zone. Such material may either be extruded at the surface or emplaced within or at the base of the crust as for example beneath the Hatton Bank (White et al., 1987).

In Kenya, the amounts of basaltic volcanic material extruded at the surface are small and decrease southwards. The maximum amount of uplift occurs above the deep anomalous mantle zone defined from teleseismic studies. It is likely that beneath this region maximum temperatures occur at the base of the crust. The lack of enormous volumes of surface volcanics, together with the predicted small amounts of intrusives into the lower crust (see item 1 in section 3.2), suggests that the amount of extension above the deep anomalous mantle zone has been very small. Further north where extension has been greater, as is evident from the abnormally thinned crust, volumes of extruded basaltic lava are still small (60,000 km$^3$), though they are greater than over the apex of the Kenya dome. In this region, it is therefore probable that the mantle temperatures, although elevated, are not raised as much above the potential temperature as beneath the Kenya dome. It is of interest to note that the velocity of the sub-crustal material increases marginally to the north, consistent with a temperature decrease in the same direction.

One particularly intriguing piece of data needs some comment. From a shot in Lake Turkana (LTN), the KRISP90 axial profile showed a reflector beneath the northern end of the line at a depth of 45 km (Fig. 3b). No velocity was obtainable for the material below this reflector. A model across the Rift at latitude 2°N (Khan and Swain, 1978), loosely constrained by the results of a combined seismic and gravity interpretation of a profile at 0.5°N, suggests the presence of an anomalous low density, sub-crustal zone below a depth of 20 km under the rift axis, which reverts to a normal mantle density at a depth of about 50 km; this is compatible with the presence of the 45 km deep reflector. It may be speculated, that this reflector delimits the underside of a mushroom-shaped mantle plume rising beneath the apex of the Kenya dome. It is possible that the plume head extends to the north due to the greater amount of extension occurring northwards along the Kenya Rift.

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