ELSEVIER

Tectonophysics 236 (1994) 465–483

TECTONOPHYSICS

The East African rift system in the light of KRISP 90

G.R. Keller [a], C. Prodehl [b], J. Mechie [b,1], K. Fuchs [b], M.A. Khan [c], P.K.H. Maguire [c], W.D. Mooney [d], U. Achauer [e], P.M. Davis [f], R.P. Meyer [g], L.W. Braile [h], I.O. Nyambok [i], G.A. Thompson [j]

[a] *Department of Geological Sciences, University of Texas at El Paso, El Paso, TX 79968-0555, USA*
[b] *Geophysikalisches Institut, Universität Karlsruhe, Hertzstrasse 16, D-76187 Karlsruhe, Germany*
[c] *Department of Geology, University of Leicester, University Road, Leicester LE1 7RH, UK*
[d] *U.S. Geological Survey, Office of Earthquake Research, 345 Middlefield Road, Menlo Park, CA 94025, USA*
[e] *Institut de Physique du Globe, Université de Strasbourg, 5 Rue René Descartes, F-67084 Strasbourg, France*
[f] *Department of Earth and Space Sciences, University of California at Los Angeles, Los Angeles, CA 90024, USA*
[g] *Department of Geology and Geophysics, University of Wisconsin at Madison, Madison, WI 53706, USA*
[h] *Department of Earth and Atmospheric Sciences, Purdue University, West Lafayette, IN 47907, USA*
[i] *Department of Geology, University of Nairobi, P.O. Box 14576, Nairobi, Kenya*
[j] *Department of Geophysics, Stanford University, Stanford, CA 94305, USA*

Received 21 September 1992; accepted 8 November 1993

Abstract

On the basis of a test experiment in 1985 (KRISP 85) an integrated seismic-refraction/teleseismic survey (KRISP 90) was undertaken to study the deep structure beneath the Kenya rift down to depths of 100–150 km. This paper summarizes the highlights of KRISP 90 as reported in this volume and discusses their broad implications as well as the structure of the Kenya rift in the general framework of other continental rifts. Major scientific goals of this phase of KRISP were to reveal the detailed crustal and upper mantle structure under the Kenya rift, to study the relationship between mantle updoming and the development of sedimentary basins and other shallow structures within the rift, to understand the role of the Kenya rift within the Afro-Arabian rift system and within a global perspective, and to elucidate fundamental questions such as the mode and mechanism of continental rifting.

The KRISP results clearly demonstrate that the Kenya rift is associated with sharply defined lithospheric thinning and very low upper mantle velocities down to depths of over 150 km. In the south-central portion of the rift, the lithospheric mantle has been thinned much more than the crust. To the north, high-velocity layers detected in the upper mantle appear to require the presence of anistropy in the form of the alignment of olivine crystals. Major axial variations in structure were also discovered, which correlate very well with variations in the amount of extension, the physiographic width of the rift valley, the regional topography, and the regional gravity anomalies. Similar relationships are particularly well documented in the Rio Grande rift.

To the extent that truly comparable data sets are available, the Kenya rift shares many features with other rift zones. For example, crustal structure under the Kenya, Rio Grande, and Baikal rifts and the Rhine Graben is generally symmetrically centered on the rift valleys. However, the Kenya rift is distinctive, but not unique, in terms of the amount of volcanism. This volcanic activity would suggest large-scale modification of the crust by magmatism.

[1] Present address: GeoForschungsZentrum, Telegrafenberg A3, D-14407 Potsdam, Germany.

SSDI 0040-1951(94)00046-C

Although there is evidence of underplating in the form of a relatively high-velocity lower crustal layer, there are no major seismic velocity anomalies in the middle and upper crust which would suggest pervasive magmatism. This apparent lack of major modification is an enigma which requires further study.

1. Introduction

As is described in detail by Swain et al. (1994), the Kenya Rift International Seismic Project (KRISP) started in 1968 as a largely British effort and has culminated to date with the large, integrated seismic investigations of KRISP 85 (KRISP Working Group, 1987; Khan et al., 1989; Henry et al., 1990; Achauer, 1990; Green et al., 1991) and KRISP 89–90 (KRISP Working Group, 1991; KRISP Working Party, 1991; Keller et al., 1992; Achauer et al., 1992) whose results have been described in detail in this volume. The purpose of this contribution is to summarize the results of the individual studies of the various seismic refraction/wide-angle reflection lines (Gajewski et al., 1994; Mechie et al., 1994a; Keller et al., 1994; Maguire et al., 1994; Braile et al., 1994; Prodehl et al., 1994) and teleseismic investigations (Achauer et al., 1994; Slack and Davis, 1994; Ritter and Achauer, 1994), analyze these results within the larger perspective of the East African rift system, and evaluate their implications for our understanding of continental rifting in general. We will also point out key questions which we believe need to be addressed in future studies.

2. Lithospheric structure of the Kenya rift and its implications

An index map of the KRISP teleseismic networks and explosion profiles is shown in Fig. 1. In order to facilitate the comparison of these results, a fence diagram showing the velocity models obtained is shown in Fig. 2. The new picture of crustal structure obtained from the KRISP effort was combined with previous results to revise the contour map of crustal thickness constructed by Keller at al. (1991). Previous results from the Kaptagat and Ngurunit areas indicated a crustal thickness of just over 40 km under portions of the rift flanks (Maguire and Long, 1976; Pointing and Maguire, 1990). In addition, studies of teleseismic waveforms recorded at Nairobi (Bonjer et al., 1970; Herbert and Langston, 1985) indicated a crustal thickness of about 40 km which is consistent with recent xenolith data (Henjes-Kunst and Altherr, 1992). These xenolith data indicate a systematic crustal thickening along the eastern flank of the rift valley from 33 km near Marsabit in northeastern Kenya to 42 km under the Chyulu Hills in southeastern Kenya (Fig. 3). Finally, seismic reflection and gravity data in the Anza rift (Fig. 1) also suggest crustal thinning in northeastern Kenya (Greene et al., 1991; Dindi, 1994). The revised crustal thickness map is shown in Fig. 3. In the vicinity of the rift, the structure in the south near Tanzania is poorly constrained. Away from the rift, particularly in eastern Kenya, the only control is the intuitive interpretation that the crust thins towards the coast and continental margin.

In terms of crustal structure, the most significant discovery from KRISP 89–90 was that the crustal thickness along the rift axis varies from as little as 20 km beneath Lake Turkana to about 35 km under the culmination of the Kenya dome near Lake Naivasha (Fig. 3). The transition in crustal thickness occurs over a horizontal distance of about 150 km (Fig. 3) between Lokori (LKO) and Lake Baringo (BAR) and thus the discrepancy between the model of Henry et al. (1990) and the tentative model of Griffiths et al. (1971) has been resolved. The change in crustal thickness along the rift is accomplished by thinning of all layers, but especially by the thinning of the lowermost crustal layer. The seismic data agree very well with the north to south increase in Bouguer gravity anomaly values (Survey of Kenya, 1982; Swain and Khan, 1978). The area of thickest crust correlates with the apex of the Kenya dome where the elevation of the rift valley floor is highest. As one proceeds northward along the rift valley from the Lake Naivasha area, the physiographic expression of the rift valley widens from its minimum of about 60 km to about 180 km in the area of thinnest crust (Fig. 1). In addition to these observations, recent seismic reflection results near Lake Turkana (Morley et al., 1992; Morley, 1994) indicate that the amount of extension across the rift increases from about 5–10 km in the Naivasha–Nakuru region (Baker and Wohlenberg, 1971; Strecker, 1991; Strecker and Bosworth, 1991) to about 35–40 km in the Lake Turkana area (Morley et al., 1992; Hendrie et al., 1994). The limited KRISP data south of Lake Naivasha suggest that there may be a decrease in crustal thickness southwards along the rift valley towards Lake Magadi. However, the need for additional data in the southern portion of the rift can be identified as a result of KRISP 89–90.

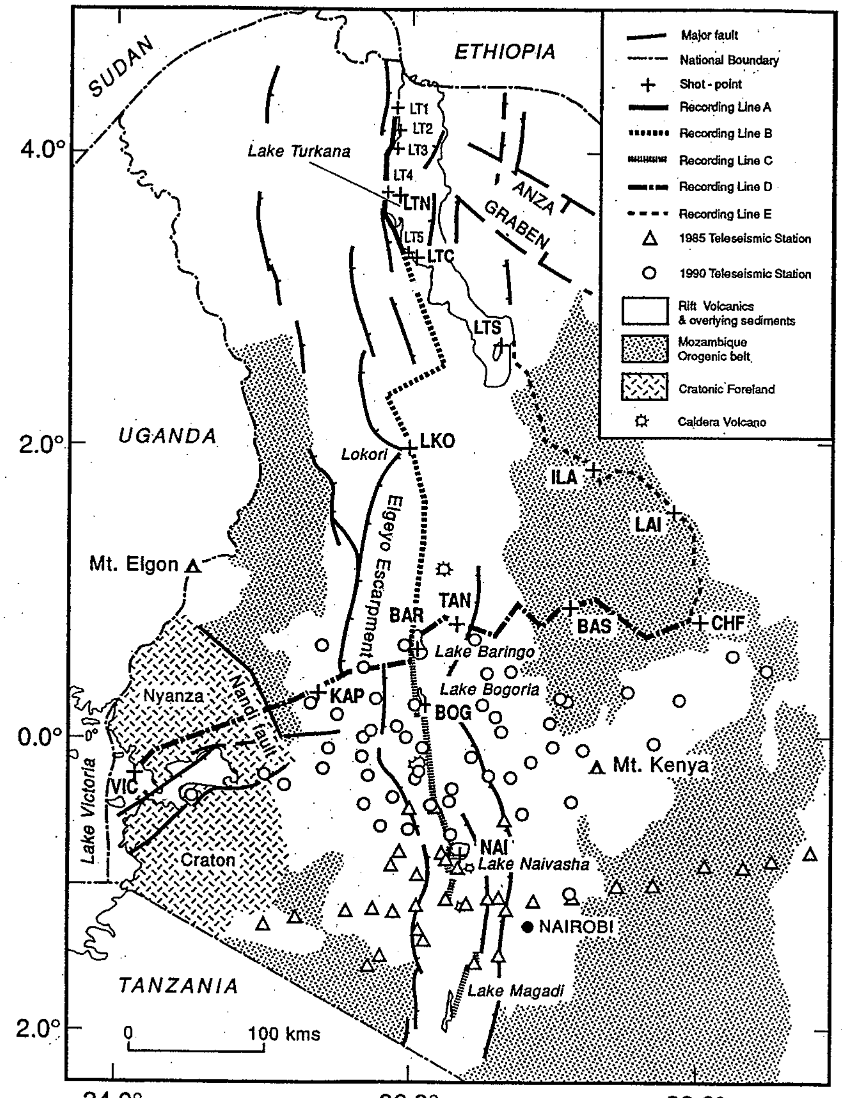

Fig. 1. KRISP 90 location map showing the seismic refraction/wide-angle reflection lines and the configuration of the teleseismic networks in 1985 and 1989–1990. The 1985 refraction/wide-angle reflection lines extended from Lake Baringo to Lake Magadi and across the rift valley just north of the Susua volcano.

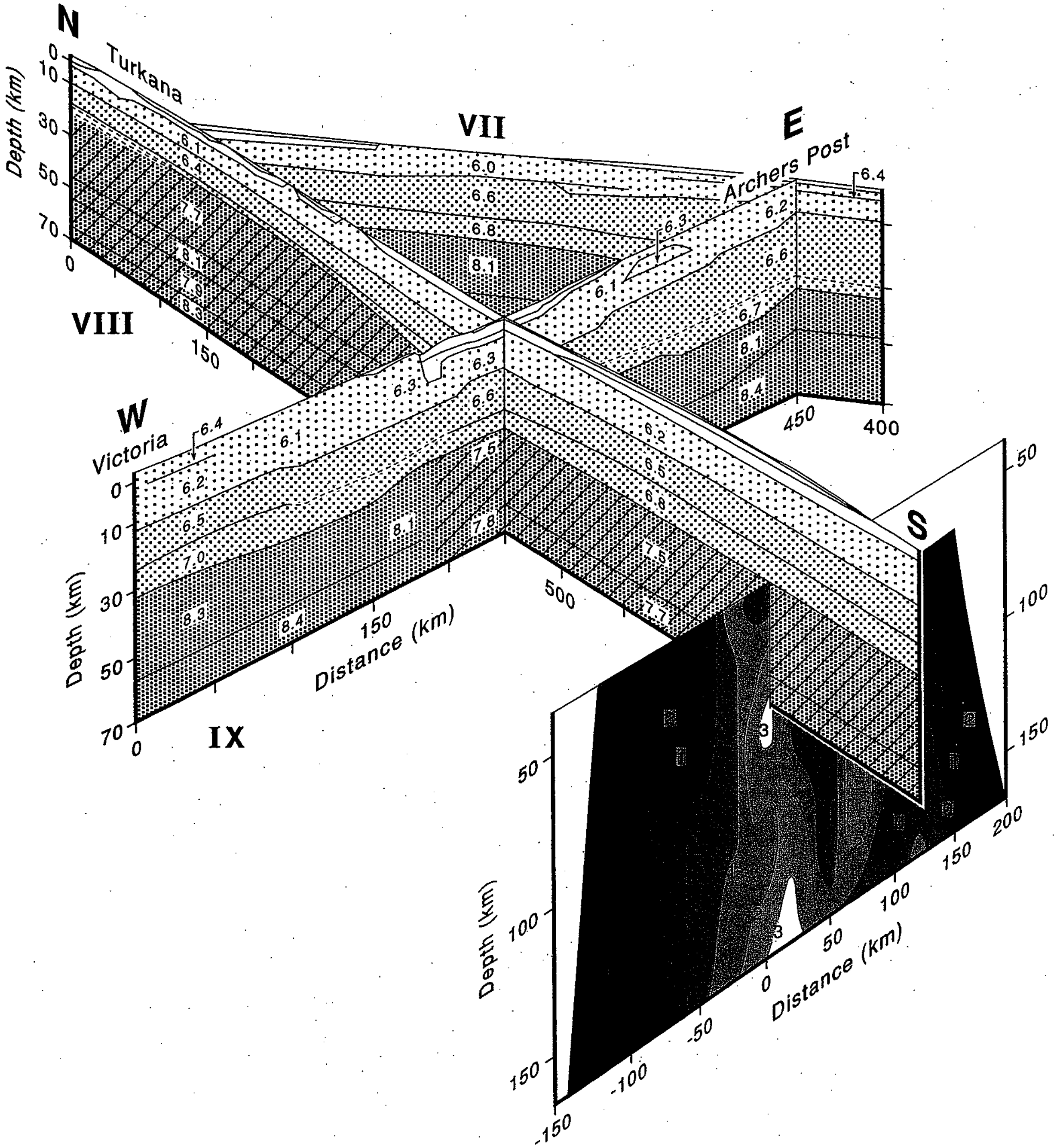

Fig. 2. Fence diagram showing crustal and uppermost-mantle structure of the Kenya rift from Lake Turkana to Lake Naivasha and beneath the neighboring flanks from Lake Victoria to Archers Post. P-wave velocities are shown in km/s. Also shown are relative velocity variations from teleseismic delay time studies across the rift at 1°S (Achauer et al., 1994). For more detailed velocity information for the crustal sections (*VII, VIII,* and *IX*), see corresponding cross-sections in enlarged form in individual papers of this volume (profile *VII*: Mechie et al., 1994a; profile *VIII*: Maguire et al., 1994; Braile et al., 1994; profile *IX*: Prodehl et al., 1994). Teleseismic velocity perturbations are contoured at 1% intervals and shaded at 2% intervals. Slow perturbations are light coloured, fast perturbations are dark coloured.

Another interesting aspect of the crustal structure is that both Maguire et al. (1994) and Braile et al. (1994) show the thickest crust (almost 40 km) beneath the western flank to occur immediately adjacent to the rift valley. Normal continental uppermost mantle velocities of 8.0–8.2 km/s are also found under the flanks. These values of crustal thickness and upper mantle velocity are consistent with the result obtained from the Kaptagat seismic array by Maguire and Long (1976). Beneath the eastern flank, the model of Braile et al. (1994) shows the thickest crust occurring immediately adjacent to the rift valley, while the model of Maguire et al. (1994) shows the thickest crust occurring somewhat more to the east. Nevertheless, on this cross-section the thickest crust beneath both flanks occurs close to the rift. Maguire et al. (1994) have suggested that if this crustal thickening had been present before the rifting started then it could have influenced the position of the rift according to the model of Vink et al. (1984). In this model, rifting tends to occur where the crust is thickest as this creates an overall weaker lithosphere because it contains a higher proportion of weaker minerals (e.g., quartz and feldspar) which are dominant in the crust and a lower proportion of stronger minerals (e.g., olivine) which are dominant in the mantle. Braile

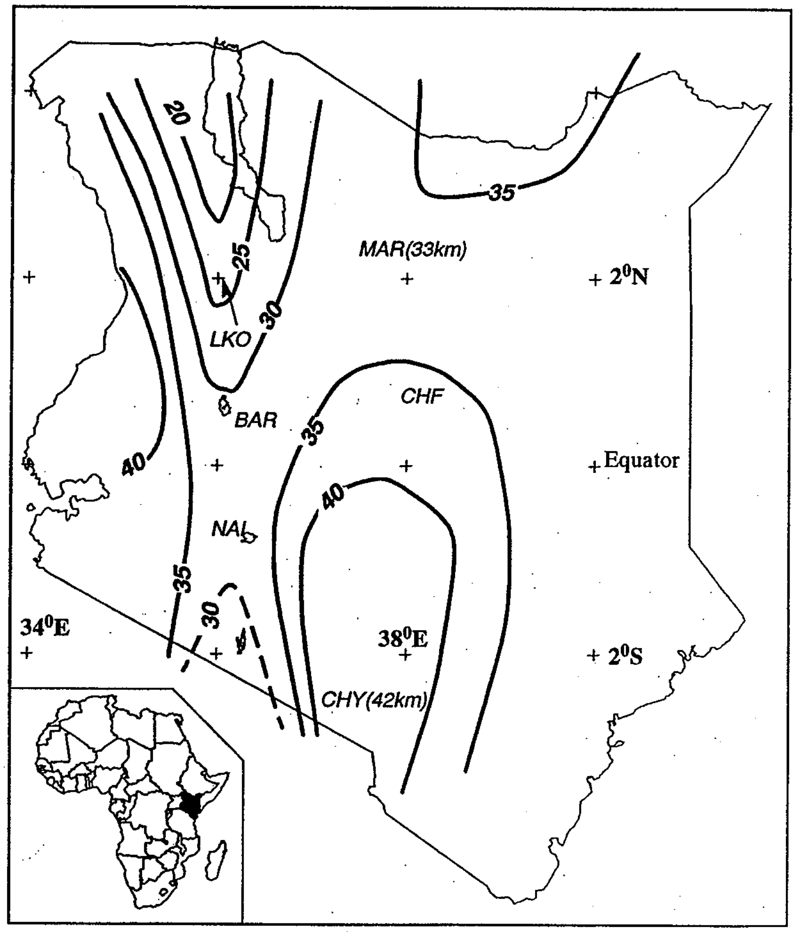

Fig. 3. Crustal thickness map for the Kenya rift region. The contour interval is 5 km. *BAR, NAI, CHF,* and *LKO* identify shotpoints at Lake Baringo, Lake Naivasha, Chanler's Falls, and Lokori, respectively. *MAR* and *CHY* indicate estimates of crustal thickness from xenoliths from the Marsabit and Chyulu Hills areas, respectively.

et al. (1994), on the other hand, suggest that this crustal thickening, which is also manifested by the isostatic effect of rift shoulder uplift, is the result of intrusion/underplating. Of course, the thickening may partly have existed prior to the rift initiation and may be partly due to lower crustal intrusion during rifting isostatically balancing the shoulder uplift. A seismic profile across the rift where little or no shoulder uplift is observed may help to solve the question of to what extent was the crust thicker before rifting began and to what extent has it been thickened below the shoulders due to the rifting process.

East of the eastern flank, the crust thins again to about 33–34 km in the vicinity of Chanler's Falls (Fig. 3). Although Quaternary volcanism is observed in the area, the influence of the Mesozoic to Early Tertiary Anza rift (Fig. 1) in northeastern Kenya (Greene et al., 1991; Bosworth and Morley, 1994; Dindi, 1994) may play some role in this off-rift crustal thinning. Nevertheless, the flank profile lies between the Anza rift and the Kenya rift south of Lake Turkana; thus, we can conclude that the crustal thinning along the Kenya rift with respect to that beneath the flank line must have been caused by the late Cenozoic rifting episode.

The rift infill thickness along the axial and cross-rift profiles is generally 3–5 km. However, large variations occur. For example, the rift infill is zero where Precambrian crystalline basement is exposed on the crests of steeply tilted fault blocks. On the other hand, the rift infill thickness approaches 10 km in the deepest part of the Elgeyo basin at the foot of the Elgeyo escarpment along the cross-rift profile (Fig. 1). The effect of the well documented basins in the Lake Turkana region (Morley et al., 1992; Hendrie et al., 1994) can also be recognized on a band-pass filtered (15–125 km) residual gravity anomaly map in which they are associated with local minima of some tens of milligals. Within the crystalline crust, three principal layers can be recognized (Fig. 2). The top of the upper crystalline crust generally has velocities of 6.0–6.3 km/s. However, "blocks" or "layers" with velocities about 0.2 km/s greater than the average background values have been recognized. Examples are found within the upper crystalline crust beneath the rift flanks on the cross-rift and flank profiles (Maguire et al., 1994; Prodehl et al., 1994) and within the crystalline crust beneath the central part of the rift valley itself (Ritter and Achauer, 1994). These high-velocity zones may represent regions which have experienced significant amounts of mafic intrusive activity. Individual intrusions, most probably dikes, would be too small to be detected by the KRISP data. However, enough of them could raise the average velocity and density of a region of the upper crust enough to be detected. Swain (1992) has made a similar argument in his interpretation of the source of part of the axial gravity high along the rift.

The top of the lower crust is marked by a jump in velocity up to 6.4–6.6 km/s, and the depth to the top of the lower crust varies from 9 to 18 km. Shallower depths are encountered beneath the northern part of the rift valley where extension has been the greatest and beneath the flank profile. The basal 2–10 km of the lower crust is marked by higher velocities generally in the range of 6.7–6.9 km/s. The inversion analysis of Braile et al. (1994) shows that the highest velocities in this basal layer may be restricted to the rift valley region. Beneath the Nyanza craton (Fig. 1), the top of this basal crustal layer is a good reflector. Beneath the Mozambique belt, there is no reflection from the top of this basal layer except within the southern part of the rift valley. Beneath the northern part of the rift valley, this basal layer is too thin for the reflection from it to be distinguished from the reflection from the crust–mantle boundary (Moho), and it is represented as a 2 km thick crust–mantle transition zone beneath the northernmost 100 km of the axial profile (Fig. 2). However, the continuity of this layer from south to north, while convenient for the computer modeling, may not be real.

Crustal velocities within the rift valley itself are not significantly different from those beneath the rift flanks. Thus, it would appear that the expected lowering of seismic velocity due to increased temperature beneath the rift is either masked by pre-rift compositional variations or is compensated for by intrusion of mafic material into the crust beneath the rift itself. Crustal thinning across the rift is accomplished by thinning of both the upper and lower crusts beneath the rift valley. Along the rift, the thinning is most pronounced in the basal ~6.8 km/s layer, and as mentioned above, it may disappear altogether beneath the Lake Turkana area.

While the teleseismic tomography is not able to provide a detailed picture of the crustal velocity–depth distribution it nevertheless (in contrast to the refraction profiles) provides a long-wavelength, 3D velocity image sampling the depth range from 10 to 35 km (Achauer et al., 1994) for the central portion of the rift and neighboring areas. In contrast to the less extensive observations of Green et al. (1991), the most recent results (Achauer et al., 1994) suggest that, for the southern section of the rift, the velocities of the middle and lower crust within the rift are similar to, and in some places slightly lower (1–3%) than, those of the adjacent shoulders. Only in the north near Lake Baringo and in the south near Mt. Susua does there seem to be some spots of higher velocities in the lower crust. As pointed out by Achauer et al. (1994), it is stretching the resolution of these data to make definitive statements about crustal structure. However, we can conclude that the teleseismic data show no evidence for wholesale intrusion of mafic material in the crust.

The entire lithosphere was the target of KRISP 89–90, and good data concerning mantle structures were obtained in both the explosion and teleseismic portions of the experiment. Arrivals from two, at least somewhat, continuous seismic discontinuities were observed in the data from the long profiles extending along and across the rift (Keller et al., 1994; Maguire et al., 1994) and the teleseismic data provided a 3D picture of velocity variations down to depths of about 165 km (Achauer et al., 1994; Slack and Davis, 1994).

Beneath the northern part of the axial profile, two upper mantle high-velocity layers separated by a low-velocity zone have been identified. The upper of these two layers, at 40–45 km depth, has a velocity of 8.05–8.15 km/s and a thickness of at least 8 km. Below the low-velocity zone which has an average velocity of 7.9 km/s, the lower of the two high-velocity layers, at 60–65 km depth, has a velocity of about 8.3 km/s. Beneath the southern part of the axial profile, the upper mantle is characterized by velocities of 7.5–7.6 km/s down to about 60 km depth where the velocity increases to about 7.7–7.8 km/s. Beneath the cross-rift profile, upper mantle reflectors at about 55 km depth have been identified beneath both rift flanks while beneath the flank line an upper mantle reflection at 45–48 km depth has been recognized.

The teleseismic results (Achauer et al., 1994; Slack and Davis, 1994) indicate that below the southern part of the Kenya rift, a low-velocity body exists which extends from the Moho down to a depth of at least 165 km. For the uppermost mantle (35–65 km depth) the lateral extent of this low-velocity body appears to be similar to that of the surface expression of the rift. The overall shape of the low-velocity body is that of a steep-sided wedge of low-velocity material which is more or less confined to the rift and its western prolongation, the Nyanza trough (Fig. 1), and which only broadens at depths greater than 125 km. Within this generally N–S-trending wedge of lower-velocity material, there is a NW–SE lineation along the direction of the old Aswa shear zone of Pan-African age (Nandi fault, Fig. 1). Both trends are most pronounced in the depth range of 65–95 km, but are also visible at other depth ranges. Velocity contrasts in the upper mantle beneath the rift and its flanks are generally about 6–10%, but reach a maximum of up to 12% for some small pockets. The strong variation of velocities even within the low-velocity body reflects either areas with quite large differences in temperature or pockets which contain a somewhat larger fraction of partial melt.

At first glance, the models of crustal and upper mantle structure across the rift are similar to those earlier drawn as schematic cross-sections of rifts (e.g., Baker and Wohlenberg, 1971). However, in both the teleseismic and explosion results, the abruptness of the variations from the rift flanks to the rift valley region are surprising. This abruptness is particularly evident in the teleseismic results which are from depths where the diffusion of heat would be expected to produce broader velocity anomalies. A wide variety of starting models and parameterizations was employed in the inversion scheme which was used to interpret the teleseismic data. In all cases, the abrupt velocity anomaly in the upper mantle was derived from the inversion. Such an anomaly clearly must be very young or diffusion would have smeared it out laterally. A young origin for this anomaly is also suggested by the fact that normal heat flow values are found on the rift flanks in Kenya with the high values being restricted to the rift valley region (Morgan, 1973, 1982; Nyblade et al., 1990; Wheildon et al., 1994).

At least in central Kenya, the lithosphere east and west of the rift valley has experienced fundamentally different geologic histories (Nyblade and Pollack, 1992; Smith, 1994). The Tanzanian craton is located to the west, and the Proterozoic/Cambrian Mozambique mobile belt is located to the east. In central Kenya, the rift valley locally does seem to follow the boundary between these provinces (e.g., Baker et al., 1972), and one might expect the deep structure of these areas to be different enough to be reflected in the KRISP results. There are no striking differences between crustal and upper mantle structure east and west of the rift valley (Achauer et al., 1994; Braile et al., 1994; Maguire et al., 1994; Slack and Davis, 1994). The crust is slightly thinner to the east, but this may be due to the proximity of the Anza rift (Prodehl et al., 1994). However, as discussed above, the NW–SE-trending Aswa suture zone west of the rift valley may be associated with an anomaly in the teleseismic inversion results even at depths of about 100 km.

3. The Kenya rift within the framework of the Afro–Arabian rift system

A recent compilation of seismic investigations of lithospheric structure in the East African rift system (Prodehl and Mechie, 1991) has been updated and is shown in Fig. 4. An obvious point is that our knowledge of deep structure in the western rift remains very rudimentary, and this region needs additional study. The need for additional data on deep structure is crucial from the standpoint of rift processes because the eastern and western branches of the rift are so different in terms of volcanism. The volumes of surface volcanics differ by as much as a factor of 100 and the chemistry of the extrusives varies considerably (e.g., Williams, 1982). At present, we are unable to discern if these dissimilarities are due to variations in the mechanism of rifting or if they are simply the result of the response of different pre-rift lithospheric structures.

Bram (1975) gathered data from earthquakes to compile two refraction "profiles" for the western rift region (Fig. 4). These sections show some evidence of crustal thinning beneath the western rift valley and show rift flank crustal structures similar to those observed in Kenya. However, there is not enough information to draw detailed comparisons on such points as the nature of the transition from rift valley to flank or axial variations in structure.

Much more information is available in the Ethiopian–Afar section of the rift system (Ruegg, 1975; Berckhemer et al., 1975). These results have been synthesized by Makris and Ginzburg (1987) and Prodehl and Mechie (1991), and some interesting comparisons with Kenya are evident. Like the Kenya rift, the crust and uppermost mantle of the Ethiopian plateau appears to be similar to that of shield areas, the transition to rifted crust appears to be abrupt, and uppermost mantle velocities under rifted areas are low (< 7.8 km/s). Unlike Kenya, the crust of the Afar has little similarity to that of typical continental areas. A very thin sialic ($V_p \approx 6.1$ to 6.3 km/s) layer and a large thickness of material with velocities of 6.6–7.2 km/s led Berckhemer et al. (1975) and Mohr (1989) to conclude that the original continental crust cannot merely have been stretched and thinned uniformly as proposed by the mechanical stretching model of McKenzie (1978). In addition, material must have been accreted to the base of the lower crust by magmatic processes within the mantle, and Mohr (1989) argues that almost the entire crust in Afar is new igneous material. The 6.8–6.9 km/s lower crust layer beneath the Kenya rift and its flanks could also represent underplated material.

In the vicinity of the major boundaries of the Kenya rift on the cross-rift line, Moho depths

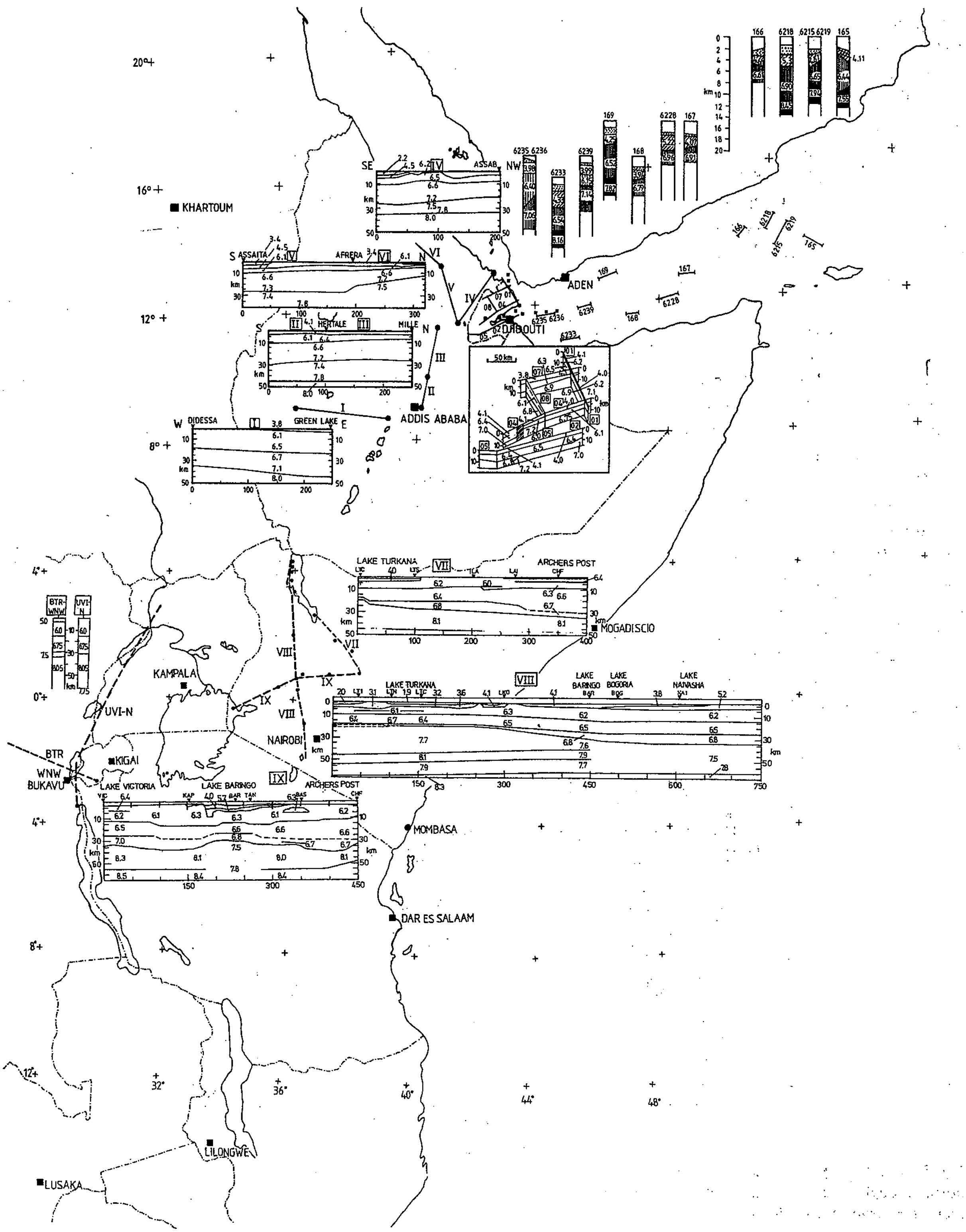

Fig. 4. Map showing the locations of seismic lines and corresponding crustal structure columns or models derived from studies in the area of the East African rift system between the Gulf of Aden and Tanzania. P-wave velocities are shown in km/s.

change quite abruptly by about 5 km over lateral distances of 10–20 km. Abrupt changes in crustal thickness are also known to occur at some places along the western boundary of the Jordan–Dead Sea rift (El-Isa et al., 1987). In southwestern Saudi Arabia, the transition from the Red Sea depression to the Arabian Shield, with a change in Moho depth in excess of 20 km over a lateral distance of a few tens of kilometers (Mechie et al., 1986; Mechie and Prodehl, 1988), appears to be the major lateral discontinuity beneath the region.

As the sequence of crustal columns (Fig. 5) shows, the central portion of the Kenya rift is one of the end members of the Afro–Arabian rift system. The sequence may be viewed as an evolution in space with the thin oceanic crust of the Gulf of Aden and the axial trough of the southern Red Sea at the center. Progressing away from the center, the crustal thickness generally increases until the thickest and closest to "normal" continental type of crusts of the Jordan–Dead Sea and the East African rifts are encountered (Mechie and Prodehl, 1988; Prodehl and Mechie, 1991). This proposed spatial evolution may be supported by the hypothesis of punctiform initiation of sea-floor spreading in the Red Sea between latitudes 22° and 24°N (Bonatti, 1985, 1987), and rift propagation (Vink et al., 1984) as applied to the northern Red Sea–Gulf of Suez–Jordan–Dead Sea rift area (Steckler and ten Brink, 1986). However, the new results from the KRISP-90 experiment show that the northern Kenya rift does not quite fit into the sequence as here the crust is thinner than in the southern Afar depression to the north (Fig. 5). It is obvious that the evolution of the East African rift is more complex than simple propagation southward from Afar.

This sequence (Fig. 5) may also be thought of as an evolutionary progression in terms of intensity of rifting (see also Girdler, 1983), but this would seem most applicable to a stretching (passive) component of rifting. It may not be a useful measure of an active component of rifting as illustrated by the southern Kenya rift. Here geological estimates of extension from structural mapping (Baker and Wohlenberg, 1971; Strecker, 1991) and crustal thickness indications of extension (Mechie et al., 1994a; Maguire et al., 1994) suggest only a small amount of stretching while teleseismic estimates of the depth of the lithosphere–asthenosphere boundary (Achauer et al., 1994; Slack and Davis, 1994) suggest that there is a large active component thinning the lithosphere from below. Morley (1994) further discusses the

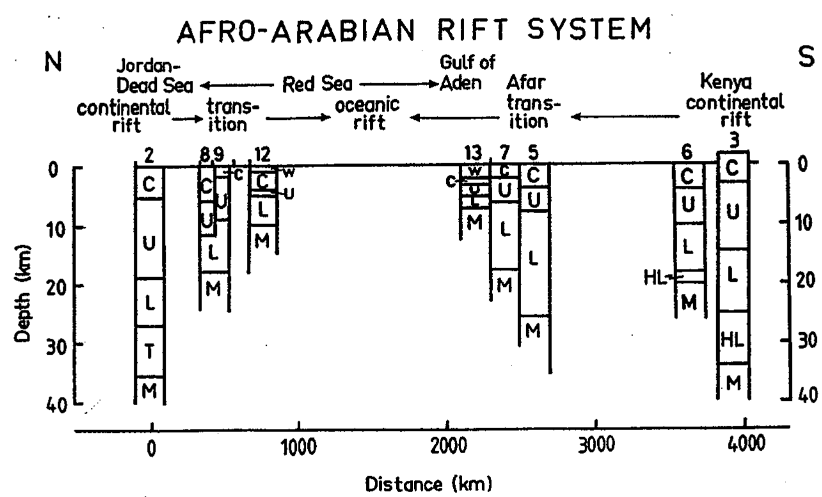

Fig. 5. Evolutionary sequence of crustal structure columns of the Afro–Arabian rift system from the Jordan–Dead Sea rift in the north through Red Sea–Gulf of Aden–Afar triangle to the East African rift in Kenya. W = water, C = cover rocks, U = upper crust, L = lower crust, HL = high-velocity lower crust, M = mantle.

interaction of deep and shallow processes in the evolution of the Kenya rift.

4. Comparison with other rifts

Comparative studies are useful to organize information and to look for key relationships. Since the Kenya rift is the classic continental rift zone, it has often been compared to other features. For example, Keller et al. (1991) conducted a detailed comparison between the Kenya and Rio Grande rifts, and Logatchev et al. (1983) compared the Kenya and Baikal rifts. In any comparison with other Cenozoic rift zones (e.g., Rio Grande, Baikal, Rhine Graben, eastern China, southern Mexico/Colima graben etc.), the Kenya and Ethiopian sections of the East African rift are distinct in terms of volume of volcanic rocks (e.g., Williams, 1982; Karson and Curtis, 1989; Macdonald, 1994). For example, in Kenya, the volume of volcanics is on the order of 50 times greater than that found in the Rio Grande rift. However, several paleorifts (e.g., midcontinent rift system of North America, Hutchinson et al., 1990; Oslo Graben, Neumann et al., 1992) were magmatically active and may have surpassed the Kenyan and Ethiopian rifts in volume of magmatic products per unit length. A major result of the KRISP effort has been to show the extent of the modification of the lithosphere caused by the Kenya rift. Although this modification is extensive, it is similar to that found in some other rift zones. Thus, despite its large volume of volcanics, this rift does not appear to need a special explanation for its evolution, and it can be studied in the context of other rift zones.

One problem in comparative studies is that data for features one would like to compare are often not of comparative quality and quantity. This problem is certainly the case for lithospheric structure. With the completion of KRISP 89–90, the lithospheric structure data bases for the Kenya rift, Rio Grande rift, and Rhine Graben are comparable. The data base for the Lake Baikal region is considerable and is being greatly expanded as a result of new cooperative efforts (e.g., Hutchinson et al., 1992). However, little basic data for the Baikal rift zone has been published in western journals, thus making comparisons difficult. For all of these rifts, one must remember that they present only discrete sections of considerably larger zones of extension (East African rift, western North America, Central European rift zone, Baikal rift zone). None-the-less, some interesting observations can be made regarding the lithospheric structure of these four rifts.

4.1. Physiography and crustal structure

An obvious observation is that all of these rifts have thinned crust as a result of extension. However, as more information has become available, considerable variations in crustal thickness have been observed along these rifts (Fig. 3). In the Kenya and Rio Grande rifts, there is a regular variation in crustal thickness which correlates with the amount of extension suggested by the number and geometry of basins and the physiographic widths of the rifts. The Rio Grande rift experiences a north to south increase in extension and width and a corresponding decrease in crustal thickness and elevation of the flanks and valley floor (Cordell, 1982; Keller et al., 1991). At least from the apex of the Kenya dome (Naivasha area) northward, similar variations occur in the Kenya rift (Morley et al., 1992; Mechie et al., 1994b). In the Rhine Graben area, there is a zone of strong crustal thinning in the south (Edel et al., 1975) which does not correlate with increased extension but does correlate with the highest (not lowest) flank elevations. At the southern end of the Central European rift, at the mouth of the Rhône river, a strong thinning of the crust correlates with increased apparent extension (Sapin and Hirn, 1974; Prodehl, 1981). Crustal structure is not well defined at its northwestern end, the Lower Rhine embayment, which is a center of recent increased seismicity and extension and may be associated with progressive lithospheric thinning (Prodehl et al., 1992; Ziegler, 1992).

Lake Baikal occupies the middle third of the Baikal rift zone (Logatchev and Zorin, 1987), and there is no obvious difference in the structure along the lake or in the physiographic expression of the rift in the region of the lake. Recent reflection data collected in the lake indicate that the basins in the southern portion are larger and perhaps older than the basins in the northern portion of the lake (Hutchinson et al., 1992). The rift zone widens both north and south of the lake, and to the north, the variations are reminiscent of the axial variations found in the Kenya and Rio Grande rifts. However, published geological and geophysical data are not sufficient to evaluate the crustal structure of this region in detail.

From all of these results, we can deduce that there are often correlatable variations in regional topography, width of the rift zone, and crustal thickness which are generally consistent with varying degrees of extension via pure shear. Relations with upper mantle structure and the amount of magmatic activity are less clear as one compares the various rift zones.

4.2. The lower crust

The lower crust is another potential source of information on rift processes, and the KRISP results have provided some unexpected results. One reason the lower crust is important is the role of rift-related magmatic processes in its formation (e.g., Wendlandt et al., 1991). Recent studies (e.g., Lachenbruch and Sass, 1978; Griffin and O'Reilly, 1987; McCarthy and Thompson, 1988; Buck, 1991) have indicated that the Moho is sometimes a transient feature during extension due to magmatic underplating and lower crustal flow. The pre-rift crust is thinned by extension but magmatic additions to the crust and/or flow may cause the Moho to remain at a nearly constant depth.

The axial variations in the lower crust delineated as a result of the KRISP study address the role of the lower crust in rifting. The volume of volcanics associated with the Kenya rift strongly suggests complementary magmatic additions to the lower crust and upper mantle (e.g., Karson and Curtis, 1989; Wendlandt et al., 1991; Macdonald, 1994). However, the observed south-to-north thinning of the lower crust is in excellent agreement with the south-to-north increase in extension. This result does not necessarily argue against magmatic additions but it does indicate that these additions either predate or have not kept up with the mechanical thinning which appears to have been concentrated in the lower crust. In the Rio Grande rift, there is also an axial variation in crustal thickness which correlates with the amount of extension (e.g., Keller et al., 1991), but this variation is not as pronounced as in Kenya. In addition, in the Rio Grande rift, the thinning is not concentrated in the lower crust. The P-wave velocities in the Rio Grande rift lower crust (~ 6.5 km/s) are not unusually high, and thus do not suggest extensive mafic magmatic additions to the lower crust (Wendlandt et al., 1991). For the Central European rift, as far as data are available for the areas of the Rhine Graben and the grabens in the French Massif Central region, no considerable thinning of the lower crust which might be related to extensive magmatism has been detected (Prodehl, 1981; Prodehl et al., 1992). In the Lake Baikal area, existing data are not sufficiently detailed to reveal details of the lower crust.

Thus, the lower crust of the Kenya rift is an enigma in that its velocity structure appears typical of continental areas, its thickness varies in a predictable way with the amount of extension, but its geometry does not reflect the known magmatic history in a straightforward way. Morley (1994) suggests that this enigma indicates that either the amount of magma emplaced in the crust in association with the volcanism is smaller than generally believed or that significant crustal material has been lost into the asthenosphere.

4.3. Deep lithospheric structure

The overall geometry of the lithosphere as a result of rifting is an important issue in efforts to model rift processes (e.g., Buck, 1991). In Kenya, the transitions in the crust and upper mantle from the flanks to the rift valley are very abrupt. In the Rio Grande rift, the transitions in crustal structure are not well known in the northern rift, but in the south, the crust gradually thins into the rifted area (e.g., Keller et al., 1991). All along the Rio Grande rift, the zone of reduced velocities in the upper mantle is broad and only approximately correlates with the rift valley (e.g., Davis, 1991). In addition, the velocity and attenuation anomalies are significant but not as large as those associated with the Kenya rift (Halderman and Davis, 1991). In the Rhine Graben, crustal structure changes abruptly when crossing the graben boundaries, though the overall crustal thickness decreases only gradually when approaching the graben axis (e.g., Brun et al., 1992; Edel et al., 1975; Prodehl, 1981; Prodehl et al., 1992). This result also applies to the French Bresse and Limagne grabens and the southern Rhône Valley (e.g., Sapin and Hirn, 1974; Hirn, 1976; Bergerat et al., 1990; Prodehl et al., 1992). There is no clear relationship of the subcrustal lithospheric structure below 50 km depth to the Rhine Graben proper (Glahn and Granet, 1992). However, under the Rhenish Massif and Lower Rhine embayment, in the seismically active continuation of the Rhine Graben to the northwest, a well pronounced low-velocity zone at 50–150 km depth is clear evidence for progressive lithospheric thinning (Raikes and Bonjer, 1983; Prodehl et al., 1992) which, however, is only observed to the southwest of the Rhine river. In the Lake Baikal area, a low-velocity zone is observed in the upper mantle. However, this zone is also asymmetrical, being offset to the southeast (e.g., Logatchev et al., 1983). The basin geometry, physiographic expression, and heat flow data argue that the crustal thinning beneath Lake Baikal is abrupt (e.g., Ruppel et al., 1992), but more data are needed to confirm this interpretation. The differences between the gradual transitions observed in the southern Rio Grande rift and the abrupt transitions observed in the Kenya, Central European and Baikal rifts may be due to differences in the state of the lithosphere prior to rifting. In the Rio Grande rift, the lithosphere was in a hot back-arc setting, in Central Europe the lithosphere was in a Variscan orogenic terrain setting, and in Kenya and the Baikal area, the lithosphere was in cool cratonal setting.

The new tomographic data from the Kenya rift (Achauer et al., 1994; Slack and Davis, 1994) confirm the observation of Green et al. (1991) that the velocity–depth pattern related to the rift reveals areas with distinct variations which make the Kenya rift a truly 3D feature. The situation is less pronounced but is similar in the Rio Grande rift where a low-velocity upper-mantle anomaly is offset slightly with respect to the direction of the main rift valley (Davis et al., 1993). Davis et al. (1993) have summarized the results of teleseismic studies of the upper mantle structure beneath the southern part of the Kenya rift, the Rio Grande rift and the Rhine Graben. Beneath the southern Kenya rift and the Rio Grande rift, large velocity, density and attenuation anomalies in the upper mantle, interpreted to be caused by partial melting, suggest that the lithosphere has been thinned much more significantly than the crust. The extensional shear processes which caused the limited amount of crustal extension (about 10 km) are unable to explain the magnitude of lithospheric thinning. An active mantle plume thinning the lithosphere from below is a more reasonable source for the observed mantle anomaly. Thus, Davis et al. (1993) place these two rifts in the category of active rifts. In contrast, the Rhine Graben which has no such mantle anomaly (Glahn and Granet, 1992) is placed by Davis et al. (1993) in the category of passive rifts. In the Baikal rift, we unfortunately lack high-resolution data. This situation will hopefully change with the joint Russian–American Baikal project currently underway. (e.g., Scholz et al., 1993).

4.4. Symmetry vs asymmetry

Deep crustal structure is essentially symmetrical in the Kenya rift, Rio Grande rift, Rhine Graben, and Baikal rift in that the maximum crustal thinning is approximately centered in the rift valleys. However, the upper crustal structure as revealed in the basins and tilted horst blocks is generally asymmetrical. The Lake Baikal and Rhine Graben basins are relatively simple and mildly tilted, but in the Rhine Graben the polarity of tilting switches from one flank to the other (e.g., Brun et al., 1992). The Rio Grande rift is distinctive in that it contains a continuous series of large, complex, en-echelon basins along its extent. These basins are generally asymmetrical and flip polarity often (e.g., Keller and Cather, 1994). A lack of drilling and seismic reflection data in Kenya along with the extensive volcanic cover limit our knowledge of its basins. However, the Project PROBE results in Lake Turkana (Dunkelman et al., 1988), the AMOCO reflection data in the Turkana area (Morley et al., 1992; Hendrie et al., 1994) and KRISP results reveal several large generally asymmetrical basins and many smaller complex basins. Gravity data do not suggest that these basins are as extensive as in the Rio Grande rift. However, dense volcanic fill in some basins could mask the negative gravity anomalies expected from large fill thicknesses.

5. Discussion

Cross-sections through the central Kenya rift derived from the KRISP-90 data (Fig. 2) constrain the crustal thinning and anomalously low upper mantle P-wave velocities down to around 165 km depth to be confined essentially to beneath the rift valley itself. In the vicinity of the major boundaries of the rift valley at the surface, the crustal thickness, upper mantle velocity, and heat flow change abruptly to attain normal continental values beneath both rift flanks. Abrupt changes in crustal thickness have also been determined for parts of the western flank of the Jordan–Dead Sea rift which is at a similar stage of development as the Kenya rift and the Arabian margin of the southern Red Sea, which is at a more advanced stage of development. What causes these changes to be so abrupt should be a topic of further research, but they seem to be evidence for the major phase of lithospheric thinning and extension to be young. Otherwise, lateral diffusion of heat would tend to spread the geophysical signatures out.

Crustal thinning is accomplished by thinning of both the upper and lower crusts (Fig. 2). The upper crust is thinned primarily by simple shear along normal faults leading to asymmetric basins and tilted fault blocks. The concentration of the crustal thinning and the anomalously low upper mantle velocities down to about 165 km depth beneath the rift valley itself is consistent with the lower crust and the lithospheric mantle being thinned by pure shear. Models requiring asymmetric development of the rift with the major crustal and lithospheric thinning being significantly offset to one side of the rift (e.g., Bosworth, 1987) are inconsistent with the present seismic results.

Mechie et al. (1994b) and Keller et al. (1994) have provided quantitative petrological interpretations of the seismic P-wave velocities in the uppermost mantle down to 60–70 km depth beneath the rift, while Halderman and Davis (1991) Achauer et al. (1994), and Slack and Davis (1994) have provided estimates of partial melting in the low-velocity body identified by the teleseismic data in the upper mantle down to about 165 km depth. Mechie et al. (1994b) have explained the low Pn velocities of 7.5–7.7 km/s beneath the rift by 3–5% partial melt of basaltic magma rising from greater depths and being trapped below the Moho. At depths between the maximum penetration of the Pn wave and 60–70 km depth, velocities lower than 7.8 km/s can also be explained by a few percent partial melt. The teleseismic results can be explained by low P-wave velocities in the upper mantle down to about 150 km depth which are due to 3–6% partial melt. Thus, the estimates for the amounts of partial melt beneath the rift derived by both the long axial profile and teleseismic data are in agreement with the value of 3% given by Macdonald (1994) from consideration of the petrogenesis of the volcanic rocks.

Mechie et al. (1994b) and Keller et al. (1994) have explained the layers with high velocities (8.0–8.3 km/s) in the uppermost mantle beneath the northern part of the rift in terms of preferred orientation of olivine and depletion of basalt. Depletion alone cannot explain the high velocities as even pure isotropic olivine rock does not have high enough velocities at the high temperatures existing beneath the rift. Depletion does, however, serve to reduce the amount of preferred mineral orientation required to obtain the high velocities. The preferred orientation of olivine could either produce an orthorhombic structure or a transverse isotropic structure. In the case of the orthorhombic structure, and in the absence of depletion, at least 40–55% of the olivine should be oriented with its *a*-axis oriented horizontally along the rift axis. This orientation would result in azimuthal dependence of velocities in the refraction data. In the case of the transverse isotropic structure, there would be no azimuthal dependence of the refraction velocities because the *a*- and *c*-axes of the olivine should be randomly oriented in the horizontal plane and the slow *b*-axis should be oriented vertically. In fact, for the deeper layer (60–65 km), both preferred mineral orientation and depletion are required to explain the high velocity of 8.3 km/s. Mechanisms which cause olivine to show preferred orientation tend to orient the *b*-axis perpendicular to the plane of the flow (Nicolas et al., 1971; Bussod, pers. commun., 1992). Thus, we envisage that a horizontal flow possibly caused by shearing has occurred in the high-velocity mantle layers. Nicolas et al. (1971) also describe the situation in which the flow (shearing) is accompanied by recrystallization. The two processes acting together can yield randomly oriented *a*- and *c*-axes in the plane of the flow (Nicolas et al., 1971).

The plate boundaries surrounding Africa suggest that the whole continent is in a compressional state of stress. The Cenozoic rifting and the tensional state of stress in many areas of Africa can therefore only be explained by a mechanism involving movements within the asthenospheric mantle. Zoback (1992) has stated that the buoyancy force due to the low-velocity, low-density upper mantle anomaly beneath the Kenya rift is more dominant than the ridge push compression and thus produces the present-day NW–SE extension. In contrast, Strecker and Bosworth (1991) and Bosworth et al. (1992) have suggested that the rotation of the stress field in the vicinity of the Kenya rift from an E–W direction to a NW–SE direction during the Quaternary is a result of far-field tectonic stresses (e.g., ridge push forces generated at the Red Sea/Gulf of Aden spreading centers).

In the southern part of the Kenya rift in the vicinity of the Kenya dome, the teleseismic results show that the lithosphere has been thinned considerably, whereas the refraction results and surface structural estimates provide evidence for modest crustal thinning and only 5–10 km extension. This situation is complicated due to the probability of magmatic additions to the crust, but active mantle upwelling seems required. The abrupt lithospheric boundaries between the flanks and rift valley and the low heat flow on the flanks are evidence that the major activity has been recent. The high surface heat flow in the rift valley, the earthquake activity and its concentration in the upper 10–12 km of the crust, and the recent volcanicity all suggest that the rift is active today.

The results of KRISP 90 have shed considerable light on the structure and evolution of the Kenya rift. However, many questions remain unanswered, and this feature is clearly an ideal place to study the relationships between the crust and mantle during extension and the role magmatism plays in the evolution of volcanically active continental rifts.

Acknowledgements

This work was financed by the National Science Foundation Continental Dynamics program in the U.S.A. (University of Texas at El Paso grant EAR-8708388, Purdue University, and Stanford University), the European Community under SCI contract 00064 involving U.K., Denmark, France, Germany, Ireland, and Italy, and the German Research Society (DFG) via the special research project SFB 108 "Stress and Stress Release in the Lithosphere" at the University of Karlsruhe. A small contribution by the N.E.R.C. (Natural Environmental Research Council) in the U.K. helped to establish a local array around Lake Baringo. The support and cooperation of the U.S. Geological Survey is also gratefully acknowledged. The help of the diplomatic representatives of the E.C., U.K., France, U.S.A., and UNESCO (Mr. Driessle) and, in particular, the continuous support of the German Embassy (Mr. Bock and Freiherr von Fritsch) are greatly acknowledged. The sponsorship of the International Lithosphere Program in cooperation with the Kenya Academy of Sciences is also acknowledged.

Special thanks are due to the Kenyan Government and the understanding officials who made the work possible. The University of Nairobi, Egerton College, Kenyatta University, the Ministry of Energy, the Department of Mines and Geology, the Survey of Kenya, the Department of Fisheries, the Ministry of Water Development, the National Environment Secretariat, the National Council for Science and Technology, the Kenya Marine and Fisheries Research Institute, the Kenya Posts and Telecommunications Corporation, the Kenya Police, Contratours, Inside Africa Safaris, Hertz Corporation, Aquadrill, TWIGA Chemical Industries all made vital contributions to the program. Helpful reviews by Paul Morgan and Henry Pollack are also gratefully acknowledged.

References

Achauer, U., 1990. Das Lithosphären–Asthenosphärensystem unter dem Ostafrikanischen Rift, Kenia. Ph.D. Thesis, Univ. Karlsruhe, 173 pp.

Achauer, U., Maguire, P.K.H., Mechie, J., Green, W.V. and the KRISP Working Group, 1992. Some remarks on the structure and geodynamics of the Kenya rift. In: P.A. Ziegler (Editor), Geodynamics of Rifting. Tectonophysics, 213: 257–268.

Achauer, U., and the KRISP Teleseismic Working Group, 1994. New ideas on the Kenya rift based on the inversion of the combined dataset of the 1985 and 1989–90 seismic tomography experiments. In: C. Prodehl, G.R. Keller and M.A. Khan (Editors), Crustal and Upper Mantle Structure of the Kenya Rift. Tectonophysics, 236: 305–329.

Baker, B.H. and Wohlenberg, J., 1971. Structure and evolution of the Kenya Rift Valley. Nature, 229: 538–542.

Baker, B.H., Mohr, P.A. and Williams, L.A.J., 1972. Geology of the Eastern rift system of Africa. Geol. Soc. Am. Spec. Pap., 136, 67 pp.

Berckhemer, H., Baier, B., Bartelsen, H., Behle, A., Burckhardt, H., Gebrande, H., Makris, J., Menzel, H., Miller, H. and Vees, R., 1975. Deep seismic soundings in the Afar region and on the highland of Ethiopia. In: A. Pilger and A. Rösler (Editors), Afar Depression of Ethiopia. Schweizerbart, Stuttgart, pp. 89–107.

Bergerat, F., Mugnier, J.-L., Guellec, S., Truffert, C., Cazes, M., Damotte, B. and Roure, F., 1990. Extensional tectonics and subsidence of the Bresse basin: an interpretation from ECORS data. Mém. Soc. Géol. Fr., N.S., 156: 145–156.

Bonatti, E., 1985. Punctiform initiation of seafloor spreading in the Red Sea during transition from a continental to an oceanic rift. Nature, 316: 33–37.

Bonatti, E., 1987. The rifting of continents. Sci. Am., March 1987, pp. 74–81.

Bonjer, K.-P., Fuchs, K. and Wohlenberg, J., 1970. Crustal structures of the East African rift system from spectral response ratios of long period body waves. Z. Geopys., 36: 287–297.

Bosworth, W., 1987. Off axis volcanism in the Gregory rift, East Africa: implications for models of continental rifting. Geology, 15: 397–400.

Bosworth, W. and Morley, C.K., 1994. Structural and stratigraphic evolution of the Anza rift, Kenya. In: C. Prodehl, G.R. Keller and M.A. Khan (Editors), Crustal and Upper Mantle Structure of the Kenya Rift. Tectonophysics, 236: 93–115.

Bosworth, W., Strecker, M.R. and Blisniuk, P.M., 1992. Integration of East African paleostress and present-day stress data: implications for continental stress field dynamics. J. Geophys. Res., 97: 11851–11865.

Braile, L.W., Wang, B., Daudt, C.R., Keller, G.R. and Patel, J.P., 1994. Modelling the 2-D velocity structure across the Kenya rift. In: C. Prodehl, G.R. Keller and M.A. Khan (Editors), Crustal and Upper Mantle Structure of the Kenya Rift. Tectonophysics, 236: 251–269.

Bram, K., 1975. Zum Aufbau der Kruste und des oberen Mantels im Bereich des westlichen Grabens des ostafrikanischen Grabensystems und im östlichen Zaire-Becken. Ergebnisse einer Untersuchung der Raumwellen von Nah-Erdbeben. Geophys. Abh., Inst. Geophysik, Freie Univ. Berlin, 4, 65 pp.

Brun, J.-P., Gutscher, M.-A., and the ECORS-DEKORP Team, 1992. Deep crustal structure of the Rhine Graben from DEKORP–ECORS seismic reflection data. In: P.A. Ziegler (Editor), Geodynamics of Rifting. Tectonophysics, 208: 139–147.

Buck, W.R., 1991. Modes of continental lithospheric extension. J. Geophys. Res., 96: 20,161–20,178.

Cordell, L., 1982. Extension of the Rio Grande rift. J. Geophys. Res., 87: 8561–8569.

Davis, P.M., 1991. Continental rift structures and dynamics with reference to teleseismic studies of the Rio Grande and East African rifts. In: A.F. Gangi (Editor), World Rift Systems. Tectonophysics, 197: 307–325.

Davis, P., Slack, P., Dahlheim, H.A., Green, W.V., Meyer, R.P., Achauer, U., Glahn, A. and Granet, M., 1993. Teleseismic tomography of continental rift zones. In: H.M. Iyer and K. Hirahara (Editors), Seismic Tomography: Theory and Practice. Chapman and Hall, London, pp. 397–439.

Dindi, E., 1994. Crustal structure of the Anza graben from gravity and magnetic investigations. In: C. Prodehl, G.R. Keller and M.A. Khan (Editors), Crustal and Upper Mantle Structure of the Kenya Rift. Tectonophysics, 236: 359–371.

Dunkelman, T.J., Karson, J.A. and Rosendahl, B.R., 1988. Structural style of the Turkana rift, Kenya. Geology, 16: 258–261.

Edel, J.B., Fuchs, K., Gelbke, C. and Prodehl, C., 1975. Deep structure of the southern Rhinegraben area from seismic refraction investigations. J. Geophys., 41: 333–356.

El-Isa, Z.H., Mechie, J., Prodehl, C., Makris, J. and Rihm, R.,

1987. A crustal structure study of Jordan derived from seismic refraction data. Tectonophysics, 138: 235–253.

Gajewski, D., Schulte, A., Riaroh, D. and Thybo, H., 1994. Deep seismic sounding in the Turkana depression, northern Kenya rift. In: C. Prodehl, G.R. Keller and M.A. Khan (Editors), Crustal and Upper Mantle Structure of the Kenya Rift. Tectonophysics, 236: 165–178.

Girdler, R.W., 1983. Processes of planetary rifting as seen in the rifting and break up of Africa. In: P. Morgan and B.H. Baker (Editors), Processes of Continental Rifting. Tectonophysics, 94: 241–252.

Glahn, A. and Granet, M., 1992. 3-D structure of the lithosphere beneath the southern Rhine Graben area. In: P.A. Ziegler (Editor), Geodynamics of Rifting, I. Case History Studies on Rifts: Europe and Asia. Tectonophysics, 208: 149–158.

Green, W.V., Achauer, U. and Meyer, R.P., 1991. A three-dimensional seismic image of the crust and upper mantle beneath the Kenya rift. Nature, 354: 199–201.

Greene, L.C., Richards, D.R. and Johnson, R.A., 1991. Crustal structure and tectonic evolution of the Anza rift, northern Kenya. In: A.F. Gangi (Editor), World Rift Systems. Tectonophysics, 197: 203–211.

Griffin, W.L. and O'Reilly, S.Y., 1987. Is the continental Moho the crust–mantle boundary? Geology, 15: 241–244.

Griffiths, D.H., King, R.F., Khan, M.A. and Blundell, D.J., 1971. Seismic refraction line in the Gregory rift. Nature, 229: 69–71.

Halderman, T.P. and Davis, P.M., 1991. Qp beneath the Rio Grande and East African rift zones. J. Geophys. Res., 96: 10,113–10,128.

Hendrie, D.B., Kusznir, N.J., Morley, C.K. and Ebinger, C.J., 1994. Cenozoic extension in northern Kenya: a quantitative model of rift basin development in the Turkana region. In: C. Prodehl, G.R. Keller and M.A. Khan (Editors), Crustal and Upper Mantle Structure of the Kenya Rift. Tectonophysics, 236: 409–438.

Henjes-Kunst, F. and Altherr, R., 1992. Metamorphic petrology of xenoliths from Kenya and northern Tanzania and implications for geotherms and lithospheric structures. J. Petrol., 33: 1125–1156.

Henry, W.J., Mechie, J., Maguire, P.K.H., Khan, M.A., Prodehl, C., Keller, G.R. and Patel, J., 1990. A seismic investigation of the Kenya rift valley. Geophys. J. Int., 100: 107–130.

Herbert, L. and Langston, C.A., 1985. Crustal thickness estimates at AAE (Addis Ababa, Ethiopia), and NAI (Nairobi, Kenya) using P-wave conversion. Tectonophysics, 111: 299–327.

Hirn, A., 1976. Sondages sismiques profonds en France. Bull. Soc. Géol. Fr., 18: 1065–1071.

Hutchison, D.R., White, R.S., Cannon W.F. and Schulz, K.J., 1990. Keweenaw hot spot: geophysical evidence for a 1.1 Ga mantle plume beneath the Midcontinent rift system. J. Geophys. Res., 95: 10,869–10,884.

Hutchison, D.R., Golmshtok, A.J., Zonenshain, L.P., Moore, T.C., Scholz, C.A. and Klitgord, K.D., 1992. Depositional and tectonic framework of rift basins of Lake Baikal from multichannel seismic data. Geology, 20: 589–592.

Karson, J.A. and Curtis, P.C., 1989. Tectonic and magmatic processes in the eastern branch of the East African rift and implications for magmatically active continental rifts. J. African Earth Sci., 8: 431–453.

Keller, G.R. and Cather, S.M., 1994. Basins of the Rio Grande rift: structure, stratigraphy and tectonic setting. Geol. Soc. Am., Spec. Paper, in press.

Keller, G.R., Khan, M.A., Morgan, P., Wendlandt, R.F., Baldridge, W.S., Olsen, K.H., Prodehl, C. and Braile, L.W., 1991. A comparative study of the Rio Grande and Kenya rifts. In: A.F. Gangi (Editor), World Rift Systems. Tectonophysics, 197: 355–372.

Keller, G.R., Braile, L.W., Davis, P.M., Meyer, R.P., Mooney, W.D., and the KRISP Working Group, 1992. Kenya Rift International Seismic Project, 1989–1990 experiment. EOS, Trans. Am. Geophys. Union, 73: 345–351.

Keller, G.R., Mechie, J., Braile, L.W., Mooney, W.D. and Prodehl, C., 1994. Seismic structure of the uppermost mantle beneath the Kenya rift. In: C. Prodehl, G.R. Keller and M.A. Khan (Editors), Crustal and Upper Mantle Structure of the Kenya Rift. Tectonophysics, 236: 201–216.

Khan, M.A., Maguire, P.K.H., Henry, W., Higham, M., Prodehl, C., Mechie, J., Keller, G.R. and Patel, J., 1989. A crustal seismic refraction line along the axis of the S. Kenya rift. J. Afr. Earth Sci., 8: 455–460.

KRISP Working Group, 1987. Structure of the Kenya rift from seismic refraction. Nature, 325 (no. 6101): 239–242.

KRISP Working Group, 1991. A compilation of data from the 1990 Kenya Rift International Seismic Project KRISP 90 seismic refraction–wide angle reflection experiment. Open-File Report 91-1, Univ. Karlsruhe, 80 pp.

KRISP Working Party, 1991. Large-scale variation in lithospheric structure along and across the Kenya Rift. Nature, 354: 223–227.

Lachenbruch, A.H. and Sass, J.H., 1978. Models of an extending lithosphere and heat flow in the Basin and Range province. In: R.B. Smith and G.P. Eaton (Editors), Cenozoic Tectonics and Regional Geophysics of the Western Cordillera. Geol. Soc. Am., Mem., 152: 209–250.

Logatchev, N.A. and Zorin, Y.A., 1987. Evidence and causes of the two-state development of the Baikal rift. Tectonophysics, 143: 225–234.

Logatchev, N.A., Zorin, Y.A. and Rogozhina, V.A., 1983. Baikal rift: active or passive? Comparison of the Baikal and Kenya rift zones. Tectonophysics, 94: 223–240.

Macdonald, R., 1994. Petrological evidence regarding the evolution of the Kenya Rift Valley. In: C. Prodehl, G.R. Keller and M.A. Khan (Editors), Crustal and Upper Mantle Structure of the Kenya Rift. Tectonophysics, 236: 373–390.

Maguire, P.K.H. and Long, R.E., 1976. The structure of the western flank of the Gregory rift (Kenya), I. The Crust. Geophys. J.R. Astron. Soc., 44: 661–675.

Maguire, P.K.H., Swain, C.J., Masotti, R. and Khan, M.A., 1994. A crustal and uppermost mantle cross-sectional model of the Kenya Rift derived from seismic and gravity data. In: C. Prodehl, G.R. Keller and M.A. Khan (Editors), Crustal and Upper Mantle Structure of the Kenya Rift. Tectonophysics, 236: 217–249.

Makris, J. and Ginzburg, A., 1987. The Afar Depression: transition between continental rifting and sea-floor spreading. Tectonophysics, 141: 199–214.

McCarthy, J. and Thompson, G.A., 1988. Seismic imaging of extended crust with emphasis on the western United States. Geol. Soc. Am. Bull., 100: 1361–1374.

McKenzie, D.P., 1978. Some remarks on the development of sedimentary basins. Earth Planet. Sci. Lett., 40: 25–32.

Mechie, J. and Prodehl, C., 1988. Crustal and uppermost mantle structure beneath the Afro–Arabian rift system. In: X. Le Pichon and J.R. Cochran (Editors), The Gulf of Suez and Red Sea Rifting. Tectonophysics, 153: 103–121.

Mechie, J., Prodehl, C. and Koptschalitsch, G., 1986. Ray path interpretation of the crustal structure beneath Saudi Arabia. Tectonophysics, 131: 333–352.

Mechie, J., Keller, G.R., Prodehl, C., Gaciri, S., Braile, L.W., Mooney, W.D., Gajewski, D. and K.-J. Sandmeier, 1994a. Crustal structure beneath the Kenya Rift from axial profile data. In: C. Prodehl, G.R. Keller and M.A. Khan (Editors), Crustal and Upper Mantle Structure of the Kenya Rift. Tectonophysics, 236: 179–199.

Mechie, J., Fuchs, K. and Altherr, R., 1994b. The relationship between seismic velocity, mineral composition and temperature and pressure in the upper mantle—with an application to the Kenya Rift and its eastern flank. In: C. Prodehl, G.R. Keller and M.A. Khan (Editors), Crustal and Upper Mantle Structure of the Kenya Rift. Tectonophysics, 236: 453–464.

Mohr, P., 1989. The nature of the crust under Afar: new igneous, not stretched continental. Tectonophysics, 167: 1–11.

Mohr, P., 1992. Nature of the crust beneath magmatically active continental rifts. Tectonophysics, 213: 269–284.

Morgan, P., 1973. Terrestrial Heat Flow Studies in Cyprus and Kenya. Ph.D. Thesis, Univ. of London, 356 pp.

Morgan, P., 1982. Heat flow in rift zones. In: G. Palmason (Editor), Continental and Oceanic Rifts. Am. Geophys. Union, Geodyn. Ser., 8: 107–122.

Morley, C.K., 1994. Interaction of deep and shallow processes in the evolution of the Kenya rift. In: C. Prodehl, G.R. Keller and M.A. Khan (Editors), Crustal and Upper Mantle Structure of the Kenya Rift. Tectonophysics, 236: 81–91.

Morley, C.K., Wescott, W.A., Stone, D.M., Harper, R.M., Wigger, S.T. and Karanja, F.M., 1992. Tectonic evolution of the northern Kenyan rift. J. Geol. Soc. London, 149: 333–348.

Neumann, E.-R., Olson, K.H., Baldridge, W.S. and Sundvoll, B., 1992. The Oslo rift: a review. Tectonophysics, 208: 1–18.

Nicolas, A., Bouchez, J.L., Boudier, F. and Mercier, J.C., 1971. Textures, structures and fabrics due to solid state flow in some European lherzolites. Tectonophysics, 12: 55–86.

Nyblade, A.A. and Pollack, H.N., 1992. A gravity model for the lithosphere in western Kenya and northeastern Tanzania. Tectonophysics, 212: 257–267.

Nyblade, A.A., Pollack, H.N., Jones, D.L., Podmore, F. and Mushayandebvu, M., 1990. Terrestrial the flow in East and Southern Africa. J. Geophys. Res., 95: 17,371–17,384.

Pointing, A.J. and Maguire, P.K.H., 1990. A seismic model for the crust in northern Kenya derived from local earthquake data. J. Afr. Earth Sci., 11: 401–409.

Prodehl, C., 1981. Structure of the crust and upper mantle beneath the central European rift system. In: R. Freund and Z. Garfunkel (Editors), The Dead Sea Rift. Tectonophysics, 80: 255–269.

Prodehl, C. and Mechie, J., 1991. Crustal thinning in relationship to the evolution of the Afro–Arabian rift system—a review of seismic-refraction data. In: J. Makris, P. Mohr and R. Rihm (Editors), Red Sea: Birth and Early History of a New Oceanic Basin. Tectonophysics, 198: 311–327.

Prodehl, C., Mueller, St., Glahn, A., Gutscher, M. and Haak, V., 1992. Lithospheric cross sections of the European Central rift system. In: P.A. Ziegler (Editor), Geodynamics of Rifting, I. Case History Studies on Rifts: Europe and Asia. Tectonophysics, 208: 113–138.

Prodehl, C., Jacob, B., Thybo, H., Dindi, E. and Stangl, R., 1994. Crustal structure on the northeastern flank of the Kenya rift. In: C. Prodehl, G.R. Keller and M.A. Khan (Editors), Crustal and Upper Mantle Structure of the Kenya Rift. Tectonophysics, 236: 271–290.

Raikes, S.A. and Bonjer, K.-P., 1983. Large-scale mantle heterogeneity beneath the Rhenish Massif and its vicinity from teleseismic P-residuals measurements. In: K. Fuchs, K. von Gehlen, H. Mälzer, H. Murawski and A. Semmel (Editors), Plateau Uplift—The Rhenish Shield—A Case History. Springer-Verlag, Berlin, pp. 315–331.

Ritter, J.R.R. and Achauer, U., 1994. Crustal tomography of the central Kenya rift. In: C. Prodehl, G.R. Keller and M.A. Khan (Editors), Crustal and Upper Mantle Structure of the Kenya Rift. Tectonophysics, 236: 291–304.

Ruegg, J.C., 1975. Main results about the crustal and upper mantle structure of the Djibouti region (T.F.A.I.). In: A. Pilger and A. Rösler (Editors), Afar Depression in Ethiopia. Schweizerbart, Stuttgart, pp. 120–134.

Ruppel, C., Kogan, M. and McNutt, M., 1992. Deep structure and active tectonics of Siberia from seismic, structural, and new 2-D gravity data. EOS, Trans. Am. Geophys. Union, 73: 209.

Sapin, M. and Hirn, A., 1974. Results of explosion seismology in the southern Rhône valley. Ann. Géophys., 30: 181–202.

Scholz, C.A., Klitgord, K.D., Hutchinson, D.R., ten Brink, U.S., Zonenshain, L.P., Golmshtok, A.Y. and Moore, T.C., 1993. Results of 1992 seismic reflection experiment in Lake Baikal. EOS, Trans. Am. Geophys. Union, 74: 465–470.

Slack, P.D. and Davis, P.M., 1994. Attenuation and velocity of

P-waves in the mantle beneath the East African Rift, Kenya. In: C. Prodehl, G.R. Keller and M.A. Khan (Editors), Crustal and Upper Mantle Structure of the Kenya Rift. Tectonophysics, 236: 331–358.

Smith, M., 1994. Stratigraphic and structural constraints on mechanisms of active rifting in the Gregory rift, Kenya. In: C. Prodehl, G.R. Keller and M.A. Khan (Editors), Crustal and Upper Mantle Structure of the Kenya Rift. Tectonophysics, 236: 3–22.

Steckler, M.S. and ten Brink, U.S., 1986. Lithospheric strength variations as a control on new plate boundaries: examples from the northern Red Sea region. Earth Planet. Sci. Lett., 79: 120–132.

Strecker, M.R., 1991. Das zentrale und südliche Kenia-Rift unter besonderer Berücksichtigung der neotektonischen Entwicklung. Habil. Schrift, University of Karlsruhe, 182 pp.

Strecker, M. and Bosworth, W., 1991. Quaternary stress-field change in the Gregory rift, Kenya. EOS, 72(17): 21–22.

Survey of Kenya, 1982. Kenya Bouguer gravity anomaly map. Scale 1: 1,000,000. Nairobi.

Swain, C.J., 1992. The Kenya rift axial gravity high: a re-interpretation. In: R. Altherr (Editor), The Afro–Arabian Rift System. Tectonophysics, 204: 59–70.

Swain, C.J. and Khan, M.A., 1978. Gravity measurements in Kenya, Geophys. J.R. Astron. Soc., 53: 427–429.

Swain, C.J., Maguire, P.K.H. and Khan, M.A., 1994. Geophysical experiments and models of the Kenya Rift before 1989. In: C. Prodehl, G.R. Keller and M.A. Khan (Editors), Crustal and Upper Mantle Structure of the Kenya Rift. Tectonophysics, 236: 23–32.

Vink, G.E., Morgan, W.J. and Zhao, W.-L., 1984. Preferential rifting of continents: a source of displaced terrains. J. Geophys. Res., 89: 10072–10076.

Wendlandt, R.F., Baldridge, W.S. and Neuman, E.R., 1991. Modification of lower crust by continental rift magmatism. Geophys. Res. Lett., 18: 1759–1762.

Wheildon, J., Morgan, P., Williamson, K.H., Evans, T.R. and Swanberg, C.A., 1994. Heat flow in the Kenya rift zone. In: In: C. Prodehl, G.R. Keller and M.A. Khan (Editors), Crustal and Upper Mantle Structure of the Kenya Rift. Tectonophysics, 236: 131–149.

Williams, L.A.J., 1982. Physical aspects of magmatism in continental rifts. In: G. Palmason (Editor), Continental and Oceanic Rifts. Am. Geophys. Union, Geodyn. Ser., 8: 193–222.

Ziegler, P.A., 1992. European Cenozoic rift system. In: P.A. Ziegler (Editor), Geodynamics of Rifting. Tectonophysics, 208: 91–111.

Zoback, M.L., 1992. First- and second-order patterns of stress in the lithosphere: the world stress map project. J. Geophys. Res., 97: 11703–11728.