Seismicity, seismotectonics and crustal structure of the southern Kenya Rift—new data from the Lake Magadi area

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SUMMARY

Local seismic activity has been monitored in the southern part of the Kenya Rift in the area around Lake Magadi. An earthquake recording network consisting of 15 station sites was operated for 8 months from November 1997 to June 1998. During this period, the Magadi area proved to be seismically active. Approximately 10 events per day were detected and found to be equally distributed over the rift floor. The hypocentre depth distribution shows surprisingly large depths of up to 27 km in the south and much shallower depths in the northern part of the area. Apart from the background activity, swarm activity with rates of more than 300 events per day was also recorded. The epicentres are clustered and trace a linear structure lying SSW–NNE over a length of 10 km north of Lake Magadi. Hypocentres in the region of the earthquake cluster are shallow and exhibit a sharp cut-off at 9 km depth. A surface crack that occurred during the recording period is connected to the earthquake swarm.

Analysis of the focal mechanisms of selected earthquakes indicates predominantly normal faulting in response to a WNW–ESE-directed tensional stress field. This direction corresponds to the general alignment of the southernmost part of the Kenya Rift and to the surface faulting pattern inside the rift.

The crustal structure of the area has been investigated using local earthquake tomography and the spatial distribution of hypocentres. The main results of the tomography are a linear positive velocity anomaly following the rift axis and a negative anomaly at shallow depth underneath Lake Magadi. The high velocities can be explained by mafic material that has intruded into the upper crust. The negative anomaly is attributed to highly fractured rocks. Maximum hypocentre depths indicate a body of low shear strength centred at the rift axis and a general deepening of the brittle–ductile transition from north to south.

Key words: crustal structure, earthquakes, fault plane solutions, rifts, seismicity, seismotectonics.

1 INTRODUCTION

Ever since geological and geophysical studies were first carried out in the Kenya Rift, the southern part of the rift has been known as a region of high geodynamic activity expressed by recent volcanism, geothermal activity and a high rate of seismicity. During one of the first microearthquake surveys in Kenya, Molnar & Aggarwal (1971) found the area around Lake Magadi (Magadi Rift), located in southern Kenya near the Tanzanian border (Fig. 1), to be seismically the most active section of the Kenya Rift. Later, as part of the Kenya Rift International Seismic Project (KRISP 94), a temporary seismic network in the vicinity of Lake Magadi recorded more than 200 local events in a 2 wk period. The data, however, have not been evaluated (Prodehl et al. 1997). Since 1994, the University of Nairobi has run a regional seismic network in Kenya consisting of five stations. One station is located in Magadi Town and records several local microearthquakes per day (Hollnack & Stangl 1998).

The Magadi Rift is therefore a suitable site to study the tectonic and crustal environment of the rifting process. For this purpose, a temporary seismic network was installed and operated for 8 months. The experiment was designed in order

(i) to undertake a detailed investigation of the local seismicity including precise earthquake location in order to estimate the earthquake size distribution of the region and to identify active fault systems;
(ii) to carry out a stress field analysis using fault plane solutions; and
(iii) to construct a model for the upper crust of the southernmost part of the Kenya Rift using the results of a local earthquake tomography study together with an evaluation of the hypocentre depth distribution.

1.1 Area of investigation

Lake Magadi is a soda lake located in the southern Kenya Rift. It lies 100 km southwest of Nairobi and 20 km north of the Tanzanian border (Fig. 1). Here, the rift valley is 60–70 km wide and flanked by a high fault to the west, the Nguruman escarpment, and several less high fault steps to the east. Further south in Tanzania the character of the rift changes from a narrow graben to a broad depression. The rift floor is a gently southward-dipping plane consisting of flood lava (Baker et al. 1988). Its morphological structure is dominated by north–south-trending ridges and grabens with displacements of up to some tens of metres as well as several volcano cones (Shombole, Lenderut, Shanamu and Olorgesailie). Lake Magadi occupies the deepest depression of the rift (580 m), has no outlet and is fed by numerous (partly hot) springs as well as rainfall. A smaller lake, Little Magadi, is located at its northern end (Fig. 2).

1.2 Seismic network

Fig. 2 is a map of the area of investigation covering approximately 5000 km² and shows the locations of the 15 station sites that formed the Magadi network. The exact positions of the sites were determined by GPS to an accuracy of ~50 m. Recordings began on November 6 1997 and ended on June 17 1998. Not all of the stations were operated during the whole recording period. Stations MG01–MG08 were installed in November 1997. Heavy persistent rains caused some flooding in the area, making it impossible to access new station sites between December 1997 and March 1998. Stations MG09–MG14 were set up in March, April and May 1998, two of them being moved later to another site. In addition to the mobile stations, the data from the permanent station located in Magadi Town (MAG) that operated during most of the recording period could be used for the study. Difficult geographical conditions and logistics resulted in a rather unevenly spaced network, omitting in particular the north and the southeast sections of the area. The centre of the area was well covered by stations with an average receiver spacing of 11 km.

The equipment used for the experiment were digital Lennartz PCM 5800 systems recording on magnetic tape. Because of the rather limited mass storage capacity the stations were run in trigger mode. The absolute time base was provided by GPS time signal receivers and power supply was provided by solar panel charged batteries. Three different sensor types were used: Lennartz LE3D-1Hz, Mark L4-3D-1Hz and Mark L4-3D-2Hz. The sampling frequency was 100 Hz and the anti-alias cut-off frequency was chosen to be 25 Hz. The permanent station MAG is equipped with a Lennartz MARS88 system, recording on optical disk.

2 Data selection and processing

In Fig. 3 the data processing steps are shown. After preparation and set-up of a database, the earthquakes were preliminarily located and different subsets of data were selected. In the next step a set of selected events were used to calculate an optimized 1-D velocity model (M1-D model) for the area of investigation, followed by relocation of all locatable events using this model. The results served to draw an epicentre map and to calculate magnitude statistics. Another data set was applied for a local earthquake tomography that supplies a 3-D crustal velocity model and a more precise hypocentre location. Selected earthquakes were used to calculate fault plane solutions.
For the basic data evaluation the Seisan software package (Havskov 1997) was used. It provides an event database management system and, among other analysis tools, the hypocenter location program (Lienert & Havskov 1995), with which preliminary earthquake location and magnitude determination were performed. The velocity model used for the preliminary location was derived from the results of the KRISP 94 experiment (Prodehl et al. 1997). The mean $V_P/V_S$ ratio of the study area was determined to be 1.74 using a cumulative Wadati diagram.

2.1 Data selection

Location routines and tomography were performed using different subsets of the earthquake data, as shown in Fig. 3. All events that were recorded by the network are represented in Data Set 0. Most of these 5590 events were not strong enough to be recorded by more than one or two of the nearest stations and are therefore not suitable for being located. 1180 local events that were recorded by more than two stations are contained in Data Set 1. Both $P$ and $S$ phases were picked for these events, followed by preliminary location. Events that were recorded by more than four stations and where the epicentres lie within the network were selected for Data Set 2. For these 332 events the azimuthal gap is mostly less than 180° and therefore they were regarded as ‘well-located events’. Approximately 60 per cent of the events of Data Set 2 turned out to form a spatially concentrated earthquake cluster. Since, as is described later, the determination of an initial velocity model for the tomography requires an equally distributed set of earthquakes, another subset of events was selected: in Data Set 3, most of the cluster events were removed and 118 events of uniform distribution remained.

2.2 Tomography

A local earthquake $P$-wave tomography was performed using a method that was originally developed by Thurber (1981) and has been subject to several modifications and enhancements since (e.g. Eberhart-Phillips & Reyners 1997). A current version, simulsps13Q, was used for this study. The routine simultaneously inverts for hypocentral coordinates and velocity structure by an iterative damped least squares method. A detailed description of the inversion process can be found in previous publications (e.g. Eberhart-Phillips & Michael 1998; Haslinger et al. 1999). The velocity model was first kept 1-D to achieve a laterally homogeneous model that best fits the observed $P$-wave traveltimes. In the second step, this model was used as input for the 3-D inversion.

The final result is a laterally heterogeneous $P$-wave velocity model, defined at nodes in a 3-D model space, together with the corresponding hypocentre locations. These locations result from a velocity model that is well adapted to the observed traveltimes and are therefore more reliable than locations obtained for a velocity model that is not optimized.

2.2.1 Minimum 1-D model

The results of a local earthquake tomography are significantly controlled by the starting velocity model. The most appropriate starting model fits the observed traveltimes best and closely reflects a priori information on the crustal segment investigated (Kissling et al. 1994). This model is called the minimum 1-D (M1-D) model. It is achieved by minimizing the squared residuals of a set of earthquake observations by a simultaneous inversion of the velocity parameters of a layered velocity model and the hypocentral parameters. The program velest (Kissling et al. 1995) was used to compute the M1-D model.

The earthquakes for the 1-D inversion have to be carefully selected. First, they should be well located, and second, the ray coverage of the model space should be as uniform as possible to outweigh the influence of possibly densely sampled model regions on the inversion results. Since a high percentage of earthquake sources recorded by the Magadi network form a spatially concentrated earthquake cluster, the second requirement would not have been fulfilled using the whole data set of well-located events. Therefore, most of the cluster events were removed and 118 visually equally distributed events (Data Set 3) with 603 $P$ onset times remained for the 1-D inversion.
During the process of finding the M1-D model, a variety of initial models with variable layer thicknesses and velocities were tested. *A priori* information came from the results of seismic line G, which, during the KRISP 94 project (Prodehl *et al.* 1997), crossed the Magadi area and had a shotpoint near Magadi Town. The crustal model that was derived from this refraction experiment (Birt *et al.* 1997) is shown in Fig. 4. Also shown is the final velocity model of the VELEST inversions that proved to be most stable with respect to changes in the initial model and which is therefore regarded as the M1-D model for the model space.

The KRISP model consists of four layers whereas the M1-D model has six layers. The boundaries of the KRISP model at 2.5 and 11 km represent lithological changes at the base of the rift infill and a mid-crustal boundary, respectively. The introduction of additional layers at these two transitions significantly improved the results of the 1-D inversion. Moreover, the seismic velocities of the M1-D model are approximately 5 per cent lower than those of the KRISP model in the depth range down to 16 km. The differences between the two models might arise because the 3-D heterogeneity of the source–receiver distribution is higher for the method of earthquake traveltime inversion than for refraction experiments, where shotpoints and stations are arranged more or less two-dimensionally.

### 2.2.2 Parameters of the model space

Input data for the 3-D tomography are a 3-D velocity starting model and a set of 332 well-located earthquakes (Data Set 2) with 2098 P onset times that have been relocated with the M1-D model. Since the to morphic inversion requires a continuous model with velocities defined at nodes in a 3-D model space (gradient model), the layered M1-D model has to be translated into a 3-D grid. The model parametrization, that is, the vertical and lateral grid node spacing and the projection of the M1-D velocity–depth function onto the grid, considerably influences the results of the tomography. With regard to the grid spacing, the number of observables (onset times) is crucial because it defines the maximum number of model parameters that can be inverted for. Furthermore, the grid spacing should have the same order of magnitude as the mean receiver spacing and be significantly greater than the signal wavelength. According to these factors, lateral grid spacings greater than 5 km were considered.

Numerous models of varying grid spacing and different representations of the M1-D velocity–depth function were tested. The criterion for selecting the optimal 3-D starting model was the amount of ssqr (sum of squared residuals) reduction obtained after four iterations with the inversion routine. The resulting model using this approach is shown in Fig. 4. It has 10 horizontal node planes and a lateral grid spacing of 7 km × 7 km (indicated in Fig. 2). Except for the rift–basement boundary at 3 km depth, the model is more or less a smoothed representation of the M1-D velocity–depth function.

#### 2.2.3 Damping parameter

The choice of the damping parameter θ of the damped least squares solution strongly influences the behaviour of the inversion: small damping values cause unrealistically high amplitudes in the solution and artefacts that tend to occur in the form of checkerboard patterns, whereas too high a value of θ produces a strongly smoothed (overdamped) solution of only little structure. An optimal value of θ can only be a compromise between these two extremes. An appropriate tool to determine the optimal damping is the so-called trade-off analysis proposed by Eberhart-Phillips (1986). This method examines the relationship between the model variance (which is a measure of the complexity of the model) and the data variance (the variance of the traveltime residuals, which is a measure of the data misfit) with respect to θ. The aim of the trade-off analysis is to identify a damping value, or at least a range of damping values, for which both model and data variance become small.

The trade-off analysis of the data set is shown in Fig. 5. In this plot the pairs of variance values obtained after one iteration form a hyperbolically shaped curve where damping increases from right to left. A strong reduction of model variance is clearly visible with only a minor increase of data variance if $\theta^2$ is changed from 2 to 12. Also, a significant lowering of data variance with only little effect on the model variance
is achieved by decreasing $h^2$ from 1000 to 50. Consequently, a damping value between 12 and 50 should be optimal for the data set. After a detailed examination of the inversion results using this range of damping values, a $h^2$ of 15 proved to be optimal and was chosen for the final $P$ tomography.

2.2.4 Solution quality

The actual framework of a seismic tomography experiment—for example, an inhomogeneous distribution of sources and limited receiver coverage—results in the spatial resolution of the structure varying considerably within the model space. Basically, the resolution of a parameter increases with the number of rays passing through a node cell. Such a hit counter can be used to identify node cells that are poorly sampled. Much more important, however, is the quality of penetration. This means that a node cell needs to be penetrated from different directions to enable an independent evaluation of the single model parameters. The distribution of seismic rays used for the tomography is shown in Fig. 6. It can be seen that the space beneath the most marginal stations is penetrated by rays covering a small angular range. Although the hit counts for these regions are high, the velocity information will be smeared along the mean ray direction here, because no crossing rays occur. Multidirectional ray coverage can be found in the centre of the model space between 36.15° E and 36.45° E and between 1.95° S and 1.65° S in a depth range down to approximately 15 km.

A quantitative measure of the quality of the solution is the elements of the resolution matrix. However, a display of the diagonal elements of the resolution matrix alone does not account for the coupling of adjacent parameters and consequently smearing effects do not become visible. A better way to display the resolution quality is to use the spread function, $Sp$ (Toomey & Foulger 1989), which is a measure of the volume in the model space that a parameter is coupled to. $Sp$ was used as a qualitative measure to estimate smearing effects in the model space and to identify regions of reliable velocity information. In Fig. 7 spread function values are displayed in depth sections running in an E–W direction through the area of investigation. Similar to the display of the ray paths (Fig. 6), the profiles indicate high resolution ($Sp < 2$) for the centre of the model space down to a depth of 10–15 km, while the marginal regions are poorly resolved. The lowest $Sp$ values ($Sp < 1$) are found underneath the central and northern parts of Lake Magadi at depths between 3 and 6 km. This spatial distribution of the solution quality has to be kept in mind when interpreting the velocity model resulting from the tomography (Sections 3.6 and 3.7).

2.2.5 Location errors

An appropriate way to estimate location errors in connection with local earthquake tomography is hypocentre uncertainty tests (Haslinger et al. 1999). In this procedure, the initial hypocentral coordinates are randomly perturbed, followed by a relocation of the disturbed hypocentres. The distribution of the differences in the hypocentral coordinates before and after the disturbance can then be used to estimate location errors. The locations of Data Set 2 were randomly perturbed by equally distributed noise of amplitude ±5 km. The standard errors resulting from this test are given in Table 1. A detailed examination of the test results shows that the mean lateral location error is approximately 300 m. Concerning the determination of focal depths, the location error of deeper events

Figure 6. Horizontal and vertical projections of ray paths for 332 events used for tomographic inversion. Triangles: seismic stations.

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proved to be small (approximately 300 m), whereas the error of events above 10 km is more than 2000 m. This observation can be explained by rather high velocity gradients found in the shallow layers that cause the location routine to converge to local traveltime residual minima. In the case of shallow events, the depth location is therefore strongly dependent on the location starting depth.

3 RESULTS AND DISCUSSION

3.1 The observed activity

Fig. 8 is an epicentre map of the earthquakes of Data Set 1. Also included are epicentres beyond the margins of the area of investigation (dashed box), in order to estimate the influence of the station distribution on the picture of seismicity. Most remarkable is the earthquake cluster north of Lake Magadi, which comprises 75 per cent of the observed events. The cluster has a N–S elongated shape and covers the area between Little Magadi and Shanamu Volcano. The other events are distributed more or less equally over the rift floor. Towards the flanks of the rift the number of earthquakes diminishes.

The concentration of activity seen in space is also found in time. From November 1997 to April 1998 an average of 10 events per day were detected. In January and February 1998, a few short (2–3 days) periods of higher activity were observed. Early in May, the main swarm activity, located in the cluster region, started. It reached rates of up to more than 300 events per day late in May and had not ceased when the network had to be removed in June.

Table 1. Standard errors of hypocentral locations at depths $z$ estimated using hypocentre uncertainty tests.

|        | $z < 10$ km | $z \geq 10$ km |
|--------|-------------|----------------|
| Longitude | 386 m      | 298 m          |
| Latitude | 274 m      | 235 m          |
| Depth   | 2268 m     | 333 m          |

Figure 7. Depth sections of spread function values computed from resolution matrix. Solid circles: grid nodes. For locations of profiles see Fig. 2.

Figure 8. Epicentre map of all earthquakes locatable in the Magadi area. Dashed box: area of investigation; triangles: seismic stations.

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The strongest earthquakes were felt, and they occurred during the main swarm activity in the cluster region. One event caused a surface tensional crack of several kilometres length with an opening width of up to 30 cm and a vertical displacement of the same order of magnitude. The crack is located in the cluster region and strikes approximately 15°NNE (see Fig. 11).

3.2 Magnitude statistics
Local magnitudes $M_L$ were determined from maximum displacement amplitudes $A$ and the epicentral distance $\Delta$ using Hutton & Boore’s (1987) formula:

$$M_L = a \log(A) + b \log(\Delta) + c \Delta + d,$$

with $a = 0.925$, $b = 0.91$, $c = 0.00087$, and $d = -1.31$. The constants $a$, $b$, $c$ and $d$ are taken from the ESARSWG (Eastern and Southern African Regional Seismological Working Group) (Hollnack & Stangl 1998). Fig. 9 shows the frequency–magnitude relation of the 1180 located earthquakes. Magnitudes range from 0.3 to 4.2 with a maximum frequency at $M_L = 1.5$. The $b$-value is 0.77 for the whole data set and 0.75 if only the events of the swarm activity are taken.

The frequency–magnitude relation suggests the recorded data set to be complete for a magnitude range $1.5 < M_L < 3.7$. The $b$-value found for small earthquakes in the southern Kenya Rift reveals that the observed seismicity is of tectonic origin, which, from Mogi (1967), applies to $b$-values between 0.5 and 1.0. It is somewhat lower than $b$-values previously reported for East Africa. Wohlenberg (1969, 1970, 1975) used a 5 yr data set that included larger earthquakes of magnitudes between 4 and 7.5 and found a mean $b$-value of 0.85 for East Africa. A similar $b$-value of 0.865 was derived by Tongue et al. (1992) from data recorded by a temporary network near Lake Bogoria in central Kenya. Hollnack & Stangl (1998) computed a somewhat higher $b$-value of 1.01 for the Rift Valley, although it was determined from only 200 events recorded in a 3 yr period.

3.3 Hypocentre distribution and faulting patterns
From the epicentre maps of the well-located earthquakes (Figs 10 and 11), it is evident that the cluster events follow a narrow, linear structure oriented 14°NNE that has a length of approximately 10 km and a width of approximately 2 km. The surface crack lies parallel to and at the eastern margin of the cluster. The distribution of the other events does not show any significant grouping. The distribution in depth is shown by means of two cross-sections that are constructed in the direction of strike of the cluster and perpendicular to it (Fig. 12). Most of the events outside the cluster region have focal depths between 10 and 26 km. The deepest earthquakes are found in the south, whereas in the northern part of the area the maximum focal depth is approximately 16 km. Shallow events in the range...
0–10 km depth occur almost exclusively in the cluster region. The focal depths of the majority of the cluster earthquakes range from 1 to 6 km. Profiles A and B illustrate the planar character of the spatial distribution of the cluster events. They show its small lateral east–west extent and a steep (>70°) westward dip. The margins of the cluster are remarkably sharp. The crustal segment underneath the cluster region is virtually free of earthquakes.

Fig. 13 is a map compiled from Baker (1963) and Matheson (1966) that shows the pattern of surface faulting in the Magadi region. As is typical for the Kenya Rift, the image is dominated by two groups of faults, the rift boundary faults (bold lines) and the smaller grid faults. Both are parallel to the general alignment of the southern Kenya Rift (NNE–SSW) and relate to the evolution of the rift and its present-day activity. A third group of faults strikes NNW–SSE. This oblique direction is assigned to reactivated pre-rift structures in the basement (Baker et al. 1972).

The major rift boundary faults, in this case the faults of the Nguruman escarpment, date back to the earliest stage of rift evolution (e.g. Baker 1958; Smith 1994). These normal faults are steep at the surface but are assumed to become less steep at depth (Bosworth et al. 1986; Bosworth 1987, 1989): in the model of alternating half-grabens the main rift faults merge into low-angle detachment systems at depths of 10–15 km and extend far beyond the rift margins. Notwithstanding the question about the expected seismic activity of such detachments, Birt et al. (1997) doubted the existence of subhorizontal interfaces beneath the Kenya Rift, referring to the results of the wide-angle refraction experiments during the KRISP project (Maguire et al. 1994; Braile et al. 1994).

The grid faulting is a dense net of small-scale, young, NNE-trending faults that extend all over the rift floor, although it is partly obscured by young volcanic piles and sediments (Baker et al. 1972). From their length, which is mostly less than a few kilometres, the depth extent of the grid faults should also be relatively small.

Comparison of the observed seismicity and the surface faulting pattern leads to the following conclusions.

(i) The epicentre distribution of the cluster events clearly follows the strike direction of the grid faults. The focal depths determined are mainly shallow, therefore it is most likely that the observed swarm activity is an expression of present-day seismic activity in the grid fault system. Further support for this assumption is given by the surface cracking that occurred during the main swarm activity in May 1998, and in which the origin of a new NNE-trending lineament could be directly observed. The observation of activity nucleating in the rift axis on small-scale faults is consistent with the idea that activity has moved from the margins of the rift towards its centre, accompanied by a decrease in fault dimension and an increase in fault density (Smith 1994).

(ii) Regarding the activity that occurred outside the cluster region, it can be established that it covers large areas of the inner rift, most probably at depths that are not reached by the grid faulting. The activity is hence attributed to deep faults located in the basement. The existence of old faults, buried beneath the rift, was also proposed by Baker et al. (1988) and confirmed by Young et al. (1991), who observed some seismic activity deeper than 10 km in the central rift near Lake Bogoria, approximately 200 km north of Lake Magadi.

(iii) Although none of the recorded earthquakes can be clearly correlated to prominent surface lineaments, some of the events appear to have occurred on the faults of the Nguruman escarpment. This becomes evident in depth section B (Fig. 12), which runs perpendicular to the western rift margin. A connection of the most westerly hypocentres (with one exception) results in a plane that dips approximately 65°E and crops out at...
the Nguruman escarpment. It therefore appears that the old boundary faults show present-day microearthquake activity. This observation has also been made by Young et al. (1991) at Lake Bogoria, where most of the activity was located at the older faults on the rift flank, but it contradicts the idea that the rift boundary faults become listric at depth, at least in the depth interval down to 26 km.

3.4 Fault plane solutions and present-day stress field

For a number of events, fault plane solutions were determined from first-motion P-wave polarities. An additional criterion for finding the nodal planes of less well-constrained solutions was the analysis of S-wave polarization angles. These data were not included in the grid search process but used to reduce the number of solutions manually. All earthquakes for which fault plane solutions were attempted belong to the group of well-located events. Azimuths and take-off angles were those computed by the 3-D ray tracer of the tomography program.

The earthquakes located inside the cluster region and those outside were treated separately in order to identify possible differences in source mechanism and stress field orientation. Single fault plane solutions were attempted for 33 earthquakes with at least six polarity readings located outside the cluster, and a composite solution was determined using the polarities of 230 cluster events. Major similarities concerning their waveforms and their occurrence in space and time allow the assumption that the cluster events originate from the same fault or from a set of faults that exhibit similar orientations and mechanisms.

Fig. 14 is a stereonet plot of the composite fault plane solution. The planes indicated in the plot represent the source mechanism for which the number of erroneous polarities is smallest. The solution is rather well constrained. 1336 polarities were used, of which 15 per cent are erroneous. Most errors arise from readings at stations that lie near a nodal plane and can be explained by a heterogeneous crust in which block rotations may also occur. One of the nodal planes strikes 25° and dips 65° NW, whereas the other plane exhibits a similar strike (4°) but lies subhorizontally (26°) and dips eastwards. We consider the steep plane (25°/65° NW) to be the fault plane, because it fits the spatial distribution of the cluster events relatively well. Furthermore, a subhorizontal fault plane at shallow depth in an extensional regime is unlikely. The mechanism is a normal fault with a P-axis that strikes 163° and plunges 68° and a T-axis that strikes 286° and plunges 20°. The axis of minimum compressive stress lies horizontally in a WNW–ESE direction.

Fig. 15 gives an overview of the source mechanisms found outside the cluster region. Most of them are normal faults with varying lateral components. The majority of the fault planes strikes NNE–SSW, as can be seen in the strike rose diagram (Fig. 15a). The maximum frequency is found for the group 10°–20° and a minor group exhibits strike directions between 160° and 170°. Due to the normal fault mechanisms, the plunge of most P-axes is steep, whereas the T-axes lie almost horizontally and are oriented WNW–ESE (Fig. 15b). As a consequence, the direction of extension is also WNW–ESE and can be determined to be between 100° and 110° (Fig. 15c).

The focal mechanism studies give a uniform picture for the pattern of seismotectonic dislocations and the stress field in the Magadi area. Both the fault plane solutions of the normal seismicity and those of the swarm activity indicate predominantly normal faulting in response to a WNW–ESE-directed tensional stress field. The most common strike direction of the fault planes is parallel to the surface fault systems and follows the general alignment of the southern Kenya Rift (NNE–SSW). A minor group of fault planes with strike directions around 165° can be assigned to the fault system that runs obliquely in a NNW–SSE direction through the rift (Fig. 13). The azimuth of 100°–110° found for the minimum compressive stress direction is in good agreement with an investigation by Jestin et al. (1994), who predicted an extension azimuth of 103° for the East African Rift. Their work was based on an inversion of plate motion data. Strecker and Bosworth assumed a rotation of the stress field in Kenya from E–W to SE–NW during the Quaternary, referring to observations made on borehole breakouts and other indicators (e.g. Strecker & Bosworth 1991; Bosworth & Strecker 1997). Concerning the southern Kenya Rift, this theory is not confirmed by our data. However, Strecker and Bosworth also emphasized that in the southern part of
the rift, due to the less advanced stage of lithospheric thinning found there, the crust may not yet react to the rotated stress field.

3.5 Deep earthquakes: evidence for mafic lower crust in the southern Kenya Rift

Models of the deformational behaviour of the Earth's crust mostly describe it in terms of two layers with the upper part of the crust showing brittle behaviour and the lower part being ductile. Stress drop through earthquakes is restricted to the upper, seismogenic crust. Therefore, the change in crustal strength is also reflected in the depth–frequency distribution of earthquakes (Meissner & Strehlau 1982). The depth of the transition between brittle and ductile behaviour is, among other factors, controlled by temperature, strain rate and rock type. In an extensional regime, material of a quartz-based rheology, as is assumed for a normal continental environment, behaves in a brittle manner at depths down to 12–15 km, and earthquakes normally occur only above this depth interval (Chen & Molnar 1983).

Since East Africa is known as a region of continental extension, earthquakes with focal depths of more than 15 km are regarded as unusually deep. Several authors have described deep crustal earthquakes located in the western branch of the East African Rift System, in Tanzania, in Zambia (Luangwa Rift) and in Ethiopia (e.g. Bungum & Nnko 1984; Shudofsky 1985; Shudofsky et al. 1987; Nyblade & Langston 1995; Camelbeeck & Iranga 1996; Nyblade et al. 1996). It has been suggested that a quartz-based rheology does not allow significant stress drops at such depths, therefore a mafic and cool lower crust with stronger rheology, originating either from magmatic underplating related to Cenozoic rifting or from Precambrian dyke swarms, is proposed for East Africa (Shudofsky et al. 1987; Nyblade & Langston 1995; Nyblade et al. 1996). For the Kenya Rift, however, no deep earthquakes have been reported (Foster & Jackson 1998). Depth–frequency distributions of earthquakes in the Lake Bogoria region indicate a peak activity at 10 km depth and a brittle–ductile cut-off depth of 12 km, which is in agreement with a normal, quartz-based rheology in this part of the Rift Valley (Young et al. 1991).

In contrast, the Magadi network located a number of events deeper than 15 km (Fig. 12). As described above, the locations were performed using the 3-D velocity model from the local earthquake tomography. Although the standard errors of the depth locations proved to be small (<500 m) for the deep events, the readings and locations of these events were carefully checked. We tested several velocity models for event relocation and observed the mean time residuals (rms) as a function of focal depth in order to analyse the significance of the depth values. A 1-D, three-layer velocity model and the HYPOCENTER location program were used for the tests. The results for one of the deepest events (located at 25.8 km depth within the 3-D model) are shown in Fig. 16. K1 is a velocity model derived from the results of the KRISP 94 refraction line G (Birt et al. 1997) and consists of a near-surface layer with \( V_p = 4.5 \text{ km s}^{-1} \) and two basement layers with \( V_p = 6.1 \) and 6.5 km s\(^{-1}\). The rms depth curve of this model shows a minimum at \( z = 25 \text{ km} \) (rms = 0.05 s). K2–K5 are modified models with increased and decreased basement velocities. Their rms depth curves show minima with time residuals higher than that of K1. Models K4 and K5 produce significantly higher residuals, and an unreasonably decrease in basement velocity of as much as 0.5 km s\(^{-1}\) is needed to achieve a shift in focus depth of only 1 km from 25 to 24 km. Tests with a number of other events have shown comparable results. We therefore regard the depth determination of our location routines to be reliable.

The frequency–depth distribution of the well-located events in the Magadi area (Fig. 17) shows that there is significant seismic activity down to depths of 22 km and minor activity down to 26 km. This result is surprising when compared to previous investigations in the central Kenya Rift near Lake Baringo and Lake Bogoria. The crustal thickness in the central part is comparable to that in the southern rift, as has been established by the KRISP refraction seismic profiles (e.g. Mechie Figure 16. Mean rms time residuals versus focal depth for a deep event in the Magadi area using different velocity models K1–K5.

Figure 17. Frequency–depth distribution of well-located events in the Magadi area.
et al. 1994), but most of the seismic activity was observed to be not deeper than 12 km depth in the central part (Young et al. 1991; Tongue et al. 1994). The greater hypocentre cut-off depth in the southern rift must therefore be related to a different crustal composition. A quartz-based crustal rheology cannot account for the deep activity in the south, and a stronger, mafic lower crust has to be assumed to explain the significant amount of seismogenic deformation below 15 km depth. Support for this idea is given by Mechlie et al. (1997), who proposed either pre-rift crustal compositional variations or mafic intrusions into the crust to explain unexpected high velocities under the rift.

3.6 Focal depth and model velocity distribution: implications for crustal structure

For a closer look at the spatial distribution of hypocentre depths, contour lines based on the maximum focal depths found in a 2 km × 2 km grid are shown in Fig. 18. Two main features are visible in the plot.

First, a N–S arrangement can be recognized. A transition, indicated by line A, subdivides the area into two parts. Maximum focal depths in the northern part are generally shallower than in the southern part. From the hypocentre depth section A (Fig. 12), a cut-off depth of approximately 15 km can be established for the northern part, whereas in the south it lies some 10 km deeper. Although, due to the limited number of recorded events, an exact transition cannot be given, the N–S arrangement in the hypocentre depth distribution suggests a change of crustal structure in the southernmost part of the Kenya Rift. While the northern cut-off depth of approximately 15 km is similar to that found in the central Kenya Rift, the significant deepening to the south appears to correspond to observations made in Tanzania by Bungum & Nnko (1984), Nyblade et al. (1996) and Zhao et al. (1997), who reported deep earthquakes with focal depths up to 40 km. Dawson (1992) compared the Tanzanian and Kenyan sectors of the Rift Valley and emphasized a major change in the signature of tectonics and volcanicity roughly located at a latitude of 2° S. This is in agreement with the transition of cut-off depths found here. A change in crustal type under the southern Kenya Rift was also proposed by Smith & Mosley (1993) and Smith (1994). In their model, which is based on morphology, structure and volcanic signature, a NW–SE-trending transition zone between the Tanzanian Archaean craton and the Proterozoic Mozambique Belt crosses the rift in southern Kenya, north of Lake Natron.

The second feature in Fig. 18 (represented by the two lines denoted B) is a stripe of generally shallower maximum focal depths that is orientated NNE–SSW and suggests an upwarped brittle–ductile transition beneath the inner rift. Comparison with a band-passed (35–45 km wavelength) Bouguer gravity map of the Magadi region (Fig. 19) reveals that the zone resembles the positive gravity anomaly oriented along the rift axis. This gravity signature is known as the axial gravity high and can be observed along most of the Kenya Rift. Several authors have modelled the anomaly and suggested that it results from mafic intrusions into the crust, be it in the form of a broad intrusion, a single dyke or dyke swarms (e.g. Searle 1970; Baker & Wohlenberg 1971; Fairhead 1976; Achauer 1992; Swain 1992). Whereas the source of the gravity anomaly, due to its short wavelength, is expected to be at shallow depth, the distribution of maximum focal depths offers information on deeper parts.

![Figure 18. Maximum focus depths in the Magadi area. The contour lines are based on the maximum focal depth value found in a 2 km × 2 km grid.](https://academic.oup.com/gji/article-abstract/146/2/439/640618/1)

![Figure 19. Band-pass (35–45 km wavelength) filtered Bouguer gravity anomaly map of the Magadi area, computed on the basis of Swain & Khan’s (1977) gravity catalogue.](https://academic.oup.com/gji/article-abstract/146/2/439/640618/2)
of the crust. Starting from the assumed mafic lithology of the lower crust, the upwarped brittle–ductile transition implies a rise in temperature beneath the inner rift, possibly related to the existence of partial melts. This anomalous hot lower crust would be the source of mafic dyke injections into the upper crust that cause the observed gravity anomaly.

Further constraints on the anomalous crustal structure of the inner rift are found from the results of the tomography. In the resulting $P$-velocity model (Fig. 20) the centre of the rift is characterized by a high-velocity anomaly. In the south it is approximately 15 km wide and starts at approximately 3 km depth. In the two northern profiles the anomaly is less constrained and it appears to be masked by a near-surface negative anomaly located under the northern extension of Lake Magadi. The positive velocity anomaly correlates with the assumed mafic material that has moved into the upper crust. Similar results have been obtained from a tomographic investigation in the central Kenya rift, where Ritter & Achauer (1994) detected intracrustal high-velocity bodies beneath the inner rift that they attributed to magmatic intrusions.

Another detail in the distribution of maximum focal depths is the unusual shallow cut-off depth just beneath the cluster region. This subject will be discussed in the next section.

### 3.7 Earthquake swarm

Earthquake swarms have commonly been found to be associated with volcanic regions, and their occurrence has been related to the movement of magma. Swarms typically have an earthquake size distribution characterized by an unusually large $b$-value between 1 and 2.5. The high $b$-value is explained by a highly fractured crust in connection with a heterogeneous stress system (Lay & Wallace 1995). Earthquake swarms are not unusual for the Kenya Rift. Tongue et al. (1994) observed swarm activity in the vicinity of Lake Baringo and interpreted it as resulting from possible opening of cracks caused by dyke intrusion.

The $b$-value found for the Magadi swarm activity is only 0.75 and does not differ significantly from the $b$-value of the background seismicity. Supported by the fact that the epicentres of the cluster events follow the general strike direction of the rift.

![Figure 20. Results of local earthquake tomography for $V_p$. Left: horizontal slices at different depth levels $z$ showing $V_p$ anomalies $dV_p$ with respect to starting velocity $V_0$. Right: W–E-oriented depth sections showing absolute $V_p$. The map indicates location of the profiles.](https://academic.oup.com/gji/article-abstract/146/2/439/640618)
faulting pattern, it can be assumed that the seismic activity is of tectonic origin. Moreover, the well-defined composite fault plane solution of the cluster events indicates tectonic activity in response to a homogeneous stress field. It is a matter of debate whether the observed swarm activity might have been triggered by magmatic activity in its vicinity: the absence of seismic activity below 9 km depth just beneath the closely restricted cluster region strongly suggests a lower crustal anomalous body of significantly reduced shear strength. A lower crustal magmatic intrusion that applies upward-directed pressure onto the upper crust could therefore explain the observed situation. The upper crust itself responds to the regional stress field with normal faulting and NNE-striking fault planes.

All activity in the cluster region is characterized by a migration of hypocentres in time from south to north, as illustrated in Fig. 21. The initial release of strain in the south may have caused accumulation of strain further north. This process could explain the shift of earthquake activity in a direction that is in accordance with the regional stress field. Another possible explanation for the migration of hypocentres is a movement of the earthquake trigger mechanism. In this case, a propagation of the above discussed magmatic intrusion in a northerly direction at an average rate of 300 m day$^{-1}$ would be proposed.

The southern margin of the cluster region located at the northern end of Lake Magadi is remarkably sharp. In this area numerous hot springs indicate a high subsurface fluid concentration. The existence of fluids is assumed to be responsible for the absence of seismic activity further south: the tomograms (Fig. 20) for this region exhibit a distinct negative velocity anomaly underneath the northern extension of the lake at shallow depths down to 5 km. The amplitude of the anomaly reaches values of more than 10 per cent. Thermal, lithological or pore pressure causes alone are not capable of decreasing velocities in the observed range and a high microcrack porosity is preferred to explain the anomaly (Schön 1996). The highly fractured subsurface could then enable fluids to penetrate the uppermost parts of the crust, where they act as lubricants and prevent strain accumulation.

**Figure 21.** Migration of cluster events. The positions of the epicentres of the cluster events are plotted along a line that follows the orientation of the cluster versus time.

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## 4 CONCLUSIONS

The seismic activity of the southernmost part of the Kenya Rift has been monitored for eight months using a network of 15 stations. The data have been analysed and interpreted in terms of seismicity, seismotectonics and crustal structure. The results can be summarized as follows.

### 4.1 Seismicity

During the observation period, an average of 10 micro-earthquakes per day were recorded in the Magadi area. This background seismicity nucleates almost exclusively at depths >10 km and can therefore be assigned to faults that are buried underneath the volcanic rift infill. Indications for possible present-day seismic activity of the rift boundary faults are obvious as well. Besides the background activity, extensive swarm activity was observed. The clustered hypocentres of the earthquake swarm trace a steep, NNE–SSW-striking plane located in the centre of the rift with a depth extent of 9 km. The swarm activity may be connected to the grid fault system. Frequency–magnitude relations of both the background and the swarm activity result in $b$-values of approximately 0.75, thus indicating a strong crust and a tectonic origin for the earthquakes.

### 4.2 Seismotectonics

The stress field of the southern Kenya Rift is directed WNW–ESE. The fault plane solutions indicate predominantly normal faults. Most fault planes are aligned parallel to the graben axis and a minor group follows an oblique direction that is controlled by pre-rift structures.

### 4.3 Crustal structure

Interpretation of focal depth distributions and a velocity model derived from local earthquake tomography allow statements to be made concerning the crustal structure of the southern Kenya Rift. The resulting model is illustrated by two depth sections (Fig. 22). The profiles are arranged perpendicular to the graben axis and sketch the crustal conditions in the northern and southern parts of the investigated area.

The distribution of maximum focal depths indicates a brittle–ductile transition depth of ~15 km in the north and ~25 km in the south. It is assumed that this change in thickness of the seismogenic crust might be related to the possible location of the craton margin between the Proterozoic Mozambique Belt and the Archaean Tanzania Craton north of Lake Natron. In the axial direction, a locally upwarped brittle–ductile transition points to the existence of ductile material underneath the rift axis, starting at a depth of 15 km (in the south) and 10 km (in the north). It remains unclear whether this linear body of low shear strength represents a magmatic intrusion that at the same time could have triggered the swarm activity. Indications for mafic material that has moved into the upper crust can be derived from the results of the tomography: the $P$-wave velocities show a positive anomaly in the rift centre. Furthermore, the interpretation of gravity measurements supports the assumption of mafic intrusions.

A strong negative $P$-wave velocity anomaly located at shallow depth underneath Lake Magadi is attributed to highly fractured
crust south of the earthquake cluster. This region of high micro-crack porosity explains the numerous hot springs that can be found here, as well as the absence of seismic activity beyond the southern margin of the cluster.

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