Seismic Structure across the Kenya Rift Valley:
Data Analysis and Geodynamic Implications

Thesis Submitted for the Degree of
Doctor of Philosophy
at the Department of Geology
Leicester University

by
Roberta Masotti
Laurea in Scienze Geologiche

March, 1995
"...and then we entered a valley, the character of which is different, both in structure and scenery, from any that we had previously seen. Its eastern wall was a straight precipitous cliff, due to a dislocation of the type known to geologists as 'faults'... We descended a few hundred feet, and then a wonderful prospect burst upon us. We were on the face of a cliff 1400 feet in height, broken only by a platform 500 feet above the floor of the valley. From the foot of the cliff a level plain extends thirty miles to the west, to the foot of the scarp of 'Mau'...

...We stopped there, lost in admiration of the beauty of and in wonder at the character of this valley..."

The Great Rift Valley, J.W. Gregory (1896).
The Nguruman escarpment from Lake Magadi
Abstract

Seismic structure across the Kenya Rift Valley: data analysis and geodynamic implications

During the 1990 Kenya Rift International Seismic Project (KRISP 90) a 450 km E-W crustal refraction profile was undertaken across the Kenya Rift, a late Tertiary to Recent extensional feature associated with extensive volcanic activity. The P-wave data have been analysed using 2-D ray-tracing, finite difference and reflectivity dynamic modelling. A simultaneous velocity and travel time inversion has been applied to the forward model to test its uniqueness and resolution. The analyses show an asymmetric sedimentary basin which is thickest against the rift’s major western boundary fault. The crustal velocities vary from 6.2 km/s in the Archaean craton to the west of the rift to about 6.0 km/s in the Proterozoic orogenic belt along the remainder of the profile. The crustal thickness outside the rift varies from 38 ± 3 km adjacent to the rift’s western margin and 34 ± 2 km to the east. Beneath the rift itself the thickness is only 30 ± 2 km. The upper mantle velocity is generally about 8.0 km/s except beneath the rift where it is consistently low at 7.6 – 7.8 km/s. This anomalously low velocity suggests a 5 – 6% partial melt. The combined seismic and gravity model supports the contention that convective processes in the mantle are dynamically supporting the uplifted East African Plateau. Kinematic and dynamic modelling of the S-wave field show that upper crustal phases have been recorded only outside the rift. Mid and lower crustal S arrivals do not seem to have been attenuated underneath the rift axis, precluding an extensive hot regime at lower crustal depth. A reflected phase is observed from an interface within the mantle beneath the western flank of the rift. Detailed analyses of this phase confirm the presence of a high velocity layer (≥8.4 km/s) below 60 km: compositional anomalies as well as crystal orientation have been suggested as an explanation for the observed velocity structure. This evidence may delimit the lateral extent of the upper mantle low velocity zone underneath the graben itself. A model of extension via simple shear in the upper crust and pure shear in the lower crust and upper mantle is suggested. The presence of a small diapir under the Kenya Rift, radiating from a ‘weak’ plume seated under the East African Plateau, is envisaged; the diapir appears to have spread asymmetrically towards the Proterozoic lithosphere to the east of the rift.
Acknowledgement

KRISP 90 was funded by the DFG (Germany), the NSF (USA), the EC and the NERC (UK) and was fully supported by participants and resources as well as the permission of the Kenya Government. An European Commission grant (Ref. No. 900009) first, and later a British Council fellowship (Ref. ITA/2281/235/A) provided funds for my studies in UK.

My thanks to Dr. Peter Maguire for becoming my supervisor. His support, guidance and enthusiasm helped me through the most difficult times of my research and my life experiences in the UK. To Val Maguire, for her caring presence.

Thanks to Dr. Jim Mechie for his help and patience during the PhD and particularly during my stay at the Geophysical Institute in Karlsruhe.

The following colleagues from Leicester University are gratefully acknowledged for their help and advice at various stages of the work: Prof. M.A. Khan, Dr. I. Hill, Dr. M. Meju, Dr. J. Luetgert, Dr. A. Saunders, Dr. S. Rigby, A. Parker, P. Denton, P. Gibson and C. Abbott.

On a personal level, thanks to the G3 group: Nick, Lieve, Mike, Nikos, and especially to my friend Christina, always there when needed; to Bipasa, for some late night phone calls; to Dickson, for editing and support in the final stages; to the international community at Leicester University, that helped me during my adaptation to English life; to my warm Spanish (and Spanish convert) friends for the time we spent together; to my Italian friends, for the long chats in my native language and some well cooked pasta; to my innumerable e-mail friends for the cyberspace communications; to all the people that made my visit to Karlsruhe productive and pleasant; to my friends who, even from Italy, always showed their care; to John, for being there, somehow.

This thesis is dedicated to my fantastic family that never failed to be with me.

....................grazie
# Contents

## 1 The Kenya Rift

1.1 Introduction ....................................................... 1

1.2 The Kenya Rift ...................................................... 2

1.2.1 Continental rifts and models of Origin ............... 5

1.3 The Geology of Kenya along the KRISP 90 E-W Profile .... 12

1.3.1 Stratigraphy ....................................................... 12

1.3.2 Xenoliths .......................................................... 22

1.3.3 Tectonics .......................................................... 24

1.4 Geodynamics of Rifts and Plate Kinematics ............... 29

1.4.1 Geodynamic Evolution of the Kenya Rift ............ 29

1.4.2 Rift Development in Kenya ................................. 31

1.5 Geophysical Studies of the Kenya Rift ..................... 34

1.5.1 Gravity Surveys in Kenya ................................. 34

1.5.2 The Deep Structure of the Kenya Rift from Teleseismic Studies ... 40

1.5.3 Seismic Wide-angle and Reflection Surveys .......... 43

1.5.4 Seismicity of the Kenya Rift ............................. 46

1.5.5 Heat Flow Measurements in Kenya .................... 48

1.5.6 Magnetotelluric and Geomagnetic Studies .......... 49

1.6 Summary .......................................................... 49

1.6.1 Thesis Outline ................................................... 50

## 2 KRISP 90

2.1 Introduction ....................................................... 55

2.2 KRISP 90 Objectives .............................................. 55
2.3 The Refraction Experiment, the Teleseismic Experiment, the Microseismic
   Network .................................................................................................................. 58
2.4 Seismic Velocities of Rock Samples ..................................................................... 65
2.5 Seismic Refraction Program - The Cross-rift Profile ......................................... 69
   2.5.1 Set up and Shooting ....................................................................................... 69
   2.5.2 Topography ................................................................................................... 72
   2.5.3 Pre-site Survey ............................................................................................. 73
   2.5.4 Communication and Timing ......................................................................... 73
   2.5.5 Recording Equipment .................................................................................... 73
   2.5.6 Data Processing .......................................................................................... 74
   2.5.7 The Seismic Data ........................................................................................ 79
   2.5.8 Profile Geology ............................................................................................. 80
2.6 Summary .................................................................................................................. 82
3 Kinematic Modelling of P-waves ........................................................................... 87
   3.1 Introduction ......................................................................................................... 87
   3.2 Picking .................................................................................................................. 87
   3.3 Record Section Description ............................................................................... 90
   3.4 Seismic Modelling ............................................................................................. 106
      3.4.1 Methods of Kinematic Modelling ............................................................. 106
      3.4.2 Model Description ....................................................................................... 126
   3.5 Gravity Modelling .............................................................................................. 136
   3.6 Summary ............................................................................................................. 141
4 Kinematic Modelling of S-waves ........................................................................... 142
   4.1 Introduction ......................................................................................................... 142
   4.2 Picking .................................................................................................................. 145
   4.3 S Model for the Cross-rift Line ......................................................................... 147
      4.3.1 Record Section Description ....................................................................... 147
      4.3.2 Ray-tracing Results and Phase Analysis .................................................. 157
      4.3.3 Model Description ....................................................................................... 165
      4.3.4 Discussion ................................................................................................. 169
   4.4 S Model for the Flank Line ................................................................................. 173
| 4.4.1 Record Section Description                                      | 173 |
| 4.4.2 Ray-tracing Results and Phase Analysis                         | 180 |
| 4.4.3 Model Description                                              | 183 |
| 4.5 Summary                                                          | 184 |
| 5 Dynamic Modelling                                                  | 186 |
| 5.1 Introduction                                                     | 186 |
| 5.2 Ray Method                                                       | 187 |
| 5.3 Reflectivity                                                     | 192 |
| 5.3.1 Amplitude Distribution                                         | 200 |
| 5.3.2 Modelling the z Phase                                          | 212 |
| 5.3.3 Modelling the S-waves                                         | 216 |
| 5.4 Finite-difference                                                | 227 |
| 5.5 Summary                                                          | 246 |
| 6 Analysis of the Mantle Phases                                      | 248 |
| 6.1 Introduction                                                     | 248 |
| 6.2 The Seismic Data                                                 | 249 |
| 6.3 Methods                                                          | 252 |
| 6.4 Procedure                                                        | 254 |
| 6.4.1 Processing                                                     | 254 |
| 6.4.2 CANC Filtering                                                 | 254 |
| 6.5 Interpretation                                                   | 261 |
| 6.5.1 Ray-Transing                                                   | 261 |
| 6.5.2 Synthetic Seismogram Modelling                                 | 262 |
| 6.5.3 Amplitude Ratio versus Offset Calculation                      | 265 |
| 6.5.4 Petrological Modelling                                         | 267 |
| 6.6 Discussion                                                       | 274 |
| 6.7 Summary                                                          | 278 |
| 7 Inverse Modelling                                                  | 280 |
| 7.1 Introduction                                                     | 280 |
| 7.2 Inverse Modelling of Line D                                      | 287 |
List of Figures

1.1 The Afro-Arabian Rift System. EAP = East African Plateau; CAR = Central African Republic; RB = Ruanda-Burundi; MW = Malawi. ................................................................. 3
1.2 Top: the Kenya Dome and Kenya Rift. Bottom: simplified topographic profile from the Western Rift to the Indian Ocean along 0 – 5°S (modified from Maguire and Khan, 1989). 4
1.4 a: Tectonic interpretation of western Kenya; b: Schematic cross-section through the Kenya Rift from Mount Elgon on the Uganda border to Mount Marsabit in northern Kenya (after Bosworth, 1987). c: formation of continental flood basalts (CFB) in the model with low-angle uniform-sense normal simple shear (after Kazmin, 1991) .................................................... 8
1.5 a: Model of an active rifting mechanism. Ascending mantle convection thins the lithosphere causing doming and lithospheric failure. b: Model of a passive rifting mechanism. Tensional stresses cause the failure of the continental lithosphere. A mantle diapir penetrates to the base of the crust causing crustal thinning (after Turcotte and Emmerman, 1983). .......... 9
1.6 Proposed hypothesis of continental rifting in eastern Africa, shown in cross-section (vertical and horizontal scales equal). a: protorift downwarp, with associated flood basalts fed from newly initiated asthenospheric diapir intruding base of lithosphere. b: rifting and penetration of lower crust by asthenolith. Profuse silicic volcanism derived from fractional melts at top of asthenolith. with additional remelting of underplated crust. c: rapid ascent of asthenolith. Underplating causes crustal thickening under rift margins (after Mohr, 1987). 11
1.7 Simplified geological map of the area surrounding the KRISP 90 cross-rift line. .......... 13
1.8 Interpretative diagrammatic cross-section across the Alps. IF: Inseabed line strike-slip fault (after Mattauer, 1986). ................................................................. 14
1.9 Principal Proterozoic structures in Kenya with locations (after Moseley, 1993). .......... 17
1.10 Geological map of the Kenya Rift volcanics (locality names in fig 1.11). After Baker et al. (1971). ................................................................. 20
1.11 Locality names for fig 1.10. After Baker et al. (1971). ................................................................. 21
1.12 Lithospheric structures underneath Quaternary volcanic fields. Textures and composition of xenoliths from Marsabit suggest a two-stage evolution with decompression and melting (after Henjes-Kunst and Altherr, 1992). ................................................................. 23
LIST OF FIGURES

1.13 Structural framework of the Kenya Rift zone. (a) Map view. (b) N-S profile of estimated depth to basement along line of rift between X and X' in Fig. 1.13a; top of basement after P-wave velocity model of Henry et al. (1990) and MOE (1987). Vertical exaggeration x8. Structures in basement are as for Fig. 1.9. After Smith and Mosley (1993). .............................. 25

1.14 Frequency of K-Ar ages for basalts from the Samburu Hills. Ages of felsic volcanism and tectonic events are also shown. Major basaltic volcanism occurred at 15 and 6 Ma and were followed by felsic volcanism and domal uplift. After Tatsumi et al. (1991) .......................................................... 27

1.15 Model for Archaean and Proterozoic crustal evolution emphasizing differences in the chemical properties of the uppermost mantle. Archaean crust developed above initially hotter mantle; magmatic underplating, if present, is ultramafic and is isotopically indistinguishable from normal mantle. A cold lithospheric keel may also act as a thermal boundary against crustal underplating by basaltic melts from asthenosphere (after Durrheim and Mooney, 1991). .............................................................................................................................................. 28

1.16 Simplified cartoon of basalt sources under a continental rift. EM=enriched lithospheric mantle; PM=OIB-type plume mantle; LAB=lithosphere-asthenosphere boundary. a: Lithospheric cross-section, prior to thermal perturbation from arriving plume head. b: At end of flood basalts episode, lava filled subsided basin is underlain by numerous sill-like intrusions that have thickened the rift crust. Lower part of lithospheric mantle has been both convected away and converted into 'active' lithosphere. c: Plume mantle head penetrates into the crust; prolific silicic magma production above ponded mantle melts (modified from Mohr, 1992). ..................................................................................................................... 30

1.17 Top: Successive block diagrams of the lithosphere corresponding to the different stages of rift evolution (after Chorowicz et al., 1987). Bottom: Diagram showing the tectonic and volcanic evolution of the Kenya Rift. The generalised spatial relation of the volcanism to the rift axis is shown (after Keller et al., 1991). ................................................................................. 32

1.18 Bouguer gravity anomaly map of Kenya, after Khan and Swain (1978). Contour intervals=10 mGals. .............................................................................................................................................. 35

1.19 Residual Bouguer gravity map of Kenya, after Khan and Swain (1978). Contour intervals=10 mGals. ............................................................................................................................................. 36

1.20 Gravity models for the Kenya Rift. A: model by Fairhead (1976) at 1.8°S showing lithospheric thinning beneath the rift valley, as well as beneath the MOB and ANC. B: model by Baker and Wohlenberg (1971) near the Equator showing thinning localised beneath the rift and a low-density sheet at the surface. C: model by Nyblade and Pollack (1992) at 0.5°N showing the 'rift' and 'suture' components. D: model by Swain (1992) at 0.5°N showing a broad deep zone of slightly increased density to explain the Kenya Rift axial gravity high. ................................................................................................................................................... 38

1.21 Crustal velocity-depth functions from earthquake studies: AFRIC (S): for shield areas from Rayleigh wave dispersion and body wave arrival times (Gumper and Pomeroy, 1970); AAE-NAl: from surface wave dispersion along the path AAE-NAl; crustal velocities and thicknesses of the first two layers are adopted from AFRIC (Long et al., 1972); Kaptagat: from body wave arrivals observed on the western flank of the rift (Maguire and Long, 1970); AFRIC (P): derived from AFRIC (S) (Gumper and Pomeroy, 1970); NAI Conversions: from P to S conversion at the Moho beneath NAI, AFRIC values for crustal velocities and thicknesses of first two layers (Herbert and Langston, 1985). ................................. 40
## LIST OF FIGURES

<table>
<thead>
<tr>
<th>Figure</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.22</td>
<td>KRISP 85 teleseismic model (after Green and Meyer, 1992), showing a cross-section (after Bosworth, 1987), travel-time residual profile, and Bouguer gravity profile at 0.5°S.</td>
<td>42</td>
</tr>
<tr>
<td>1.23</td>
<td>Map showing locations of seismic refraction lines and temporary local networks; KRISP 90 is also included.</td>
<td>44</td>
</tr>
<tr>
<td>1.24</td>
<td>Interpretation of KRISP 75 first arrival data and geological section. Velocity section: mean velocities in km/s. The zig-zag lines do not represent actual structure (after Swain et al., 1981).</td>
<td>45</td>
</tr>
<tr>
<td></td>
<td>a: Interpretation of KRISP 85 north-south profile. Shallow velocity model.</td>
<td></td>
</tr>
<tr>
<td></td>
<td>b: Interpretation of KRISP 85 north-south profile. Shallow velocity model.</td>
<td></td>
</tr>
<tr>
<td>1.26</td>
<td>Flow chart showing the seismic data analysis procedure.</td>
<td>51</td>
</tr>
<tr>
<td>2.1</td>
<td>KRISP 90 location map showing the seismic refraction/wide angle reflection lines and the configuration of the teleseismic network (trapezoidal area) (after Prodehl et al., 1994b).</td>
<td>56</td>
</tr>
<tr>
<td>2.2</td>
<td>a: 2-D ray-trace model for the KRISP 90 axial line (segments A, B and C). Velocities are accurate to the nearest 0.05 km/s. The structure north of LTN has been taken from the interpretation of line A by Gajewski et al. (1994).</td>
<td>59</td>
</tr>
<tr>
<td></td>
<td>b: 2-D ray-trace model for the KRISP 90 flank line E. Vertical exaggeration 5x. Depths are related to sea level. Dashed line is a change of velocity gradient (after Prodehl et al., 1994a).</td>
<td></td>
</tr>
<tr>
<td>2.3</td>
<td>a: Top view of the schematic map showing the 'rift' model used to calculate the theoretical residuals to test several structural elements. Stippled area=north-south oriented rift; TS=NW-SE-oriented narrow 'shear zone', with lower velocities relative to the lithospheric upper mantle; D=small diapirs within the low-velocity upper mantle beneath the rift.</td>
<td>61</td>
</tr>
<tr>
<td></td>
<td>b: E-W cross-section through the model shown in (a). Numbers indicate the velocities in km/s.</td>
<td></td>
</tr>
<tr>
<td>2.4</td>
<td>Inversion result for KRISP 90 teleseismic data. Velocity perturbations per layer with respect to the starting model are shown. Layer thickness, block size and initial velocity model are similar to the real data case. Major rift faults as heavy lines; elements of the ANC-MOB suture as dashed lines.</td>
<td>62</td>
</tr>
<tr>
<td>2.5</td>
<td>Contoured inversion results (after Ritter and Achauer, 1994).</td>
<td>64</td>
</tr>
<tr>
<td>2.6</td>
<td>Map of sample locations documented in Fig. 2.7. Samples 8, 14, 15, 16: granitic gneisses; 19: dioritic gneisses; 20: tonalite gneisses. Samples 2a, 3: basalt; 7: basaltic trachyandesite; 17a, 17b: tephriphonolites; 21a, 21b: tephrite basalts (after Mooney and Christensen, 1994).</td>
<td>66</td>
</tr>
<tr>
<td>2.7</td>
<td>Comparison of the average velocity/depth structure beneath the southern portion of the axial rift profile with metamorphic rocks (top) and volcanic rocks (bottom).</td>
<td>67</td>
</tr>
<tr>
<td>2.8</td>
<td>Composition of the crust beneath the Kenya Rift at the Equator from crustal seismic velocity structure and laboratory measurements on Kenyan rock samples (after Mooney and Christensen, 1994).</td>
<td>68</td>
</tr>
<tr>
<td>2.9</td>
<td>Examples of power spectra of P and S phases from VIC shot point.</td>
<td>73</td>
</tr>
<tr>
<td>2.10</td>
<td>Summary of spectral analysis for P and S phases from the cross-rift line D. Continuous lines: P phases; dashed lines: S phases; thick lines: reflected phases; thin lines: diving phases. Distances are marked E and W from S-P. The rift graben is indicated by the shaded area on the distance axes.</td>
<td>77</td>
</tr>
<tr>
<td>2.11</td>
<td>Observation scheme for the KRISP cross-rift line D.</td>
<td>80</td>
</tr>
<tr>
<td>2.12</td>
<td>Schematic geological section for the KRISP 90 cross-rift line D.</td>
<td>81</td>
</tr>
</tbody>
</table>
LIST OF FIGURES

3.1 Basic traveltime diagram. The numbers on both axes indicate approximate values. The curves can be shifted as much as 30-50 km and up to ±2-3 s (after Prodehl, 1979). ... 89
3.2 Trace-normalized band-pass filtered (1-25 Hz) record section for shot-point VIC recorded to the east along the cross-rift line D emphasizing phase a. Note the display gain increased by four and the time scale by two with respect to Fig. 3.3. $V_r=6 \text{ km/s}$ (ray-traced phase notation for Fig. 3.2, 3.3, 3.4: $a=P_g$, $b_1=P_{i1}P$, $b_2=P_{i2}P$, $c=P_mP$, $d=P_n$, $d_1=$mantle reflection). ... 92
3.3 Trace-normalized band-pass filtered (1-25 Hz) record section for shot-point VIC recorded to the east along the cross-rift line D. $V_r=6 \text{ km/s}$ (ray-traced phase notation: see Fig. 3.2 caption). ... 93
3.4 Trace-normalized band-pass filtered (1-25 Hz) record section for shot-point VIC recorded to the east along the cross-rift line D emphasizing the mantle phases $d$ and $d_1$. $V_r=6 \text{ km/s}$ (ray-traced phase notation: see Fig. 3.2 caption) ... 94
3.5 Trace-normalized band-pass filtered (1-15 Hz) record section for shot-point KAP recorded to the west along the cross-rift line D. $V_r=6 \text{ km/s}$ (ray-traced phase notation for Fig. 3.5, 3.6: $a=P_s$, $b=P_nP$, $b_1=P_{i1}P$, $c=P_mP$, $d=P_n$, $d_1=$unidentified phase discussed in the text). Note: dashed line identifies observed but not ray-traced phase $d$ (see text). ... 98
3.6 Trace-normalized band-pass filtered (1-15 Hz) record section for shot-point KAP recorded to the east along the cross-rift line D. $V_r=6 \text{ km/s}$ (ray-traced phase notation: see Fig. 3.5 caption) ... 99
3.7 Trace-normalized band-pass filtered (5-15 Hz) record section for shot-point BAR recorded to the west along the cross-rift line D. $V_r=6 \text{ km/s}$ (ray-traced phase notation for Fig. 3.7, 3.8: $a=P_s$, $b=P_nP$, $b_1=P_{i1}P$, $c=P_mP$, $d=P_n$, $d_1=$unidentified phase discussed in the text). ... 100
3.8 Trace-normalized band-pass filtered (5-15 Hz) record section for shot-point BAR recorded to the east along the cross-rift line D. $V_r=6 \text{ km/s}$ (ray-traced phase notation: see Fig. 3.7 caption) ... 101
3.9 Trace-normalized band-pass filtered (1-20 Hz) record section for shot-point BAS recorded to the west along the cross-rift line D. $V_r=6 \text{ km/s}$ (ray-traced phase notation for Fig. 3.9, 3.10: $a=P_s$, $b=P_nP$, $b_1=P_{i1}P$, $c=P_mP$, $d=P_n$, $d_1=$unidentified phase discussed in the text). ... 102
3.10 Trace-normalized band-pass filtered (1-20 Hz) record section for shot-point BAS recorded to the east along the cross-rift line D. $V_r=6 \text{ km/s}$ (ray-traced phase notation: see Fig. 3.9 caption) ... 103
3.11 Trace-normalized band-pass filtered (1-20 Hz) record section for shot-point BAS recorded to the west along the cross-rift line D. $V_r=6 \text{ km/s}$ (ray-traced phase notation for Fig. 3.11, 3.12: $a=P_s$, $b_1=P_{i1}P$, $c=P_mP$, $d=P_n$). Note: dashed line identifies observed but not ray traced phase $a$. ... 104
3.12 Trace-normalized band-pass filtered (1-25 Hz) record section for shot-point BAS recorded to the east along the cross-rift line D. $V_r=6 \text{ km/s}$ (ray-traced phase notation: see Fig. 3.11 caption) ... 105
3.13 Trace-normalized band-pass filtered (1-25 Hz) record section for shot-point CHF recorded to the west along the cross-rift line D. $V_r=6 \text{ km/s}$ (ray-traced phase notation: $a=P_s$, $b_1=P_{i1}P$). ... 106
3.14 1-D velocity depth functions for shot VIC, KAP and BAR. A = to the west, B = to the east ... 109
3.15 1-D velocity depth functions for shot TAN and BAS. A = to the west, B = to the east ... 110
LIST OF FIGURES

3.16 Plus-minus depth determined from the α phase. Different lines result from the combination of different data sets. ........................................... 112
3.17 Time-term depths for the α phase. Numbers give velocities in km/s. Solution variances are given in brackets. ................................................................. 114
3.18 α and φ phases ray-traced final model (for velocity distribution see Fig. 3.23. Number of rays for each phase limited to a maximum of 20 for display). Vertical exaggeration 10:1. ........................................... 118
3.19 β, δ, and γ phases ray-traced final model (for velocity distribution see Fig. 3.22. Number of rays for each phase limited to a maximum of 20 for display). Vertical exaggeration 5:1. ........................................... 121
3.20 c and d phases ray traced final model (for velocity distribution see Fig. 3.22. Number of rays for each phase limited to a maximum of 20 for display). Vertical exaggeration 5:1. ........................................... 123
3.21 Merged, trace-normalized band-pass filtered (5-15 Hz) record section for shots BAR4, BAX1, BAX2 and BAX3 recorded to the west along the cross-rift line D. Vp=6 km/s (ray-traced phase notation: see Fig. 3.7 caption) ........................................... 125
3.22 Final 2-D ray-trace model for KRISP 90 cross-rift line D. Dashed lines indicate transitions where there is a change in velocity gradient but no significant discontinuity. P-wave velocities in km/s, gradients in s^-1 in italic. ........................................... 127
3.23 Upper crustal final 2-D ray-trace model for KRISP 90 cross-rift line D. P-wave velocities in km/s, gradients in s^-1 in italic. ........................................... 129
3.24 The observed and calculated gravity profiles along the cross-rift line D showing the initial density model derived directly from the final ray-trace seismic model. The gravity effect of the deep compensation as discussed in the text is also shown. (Densities in g cm^-3). ........................................... 139
3.25 Final density model of a cross-section through the crust and upper mantle beneath the Kenya Rift along the KRISP 90 east-west line D. (Densities in g cm^-3). ........................................... 140
4.1 a: selected ray-paths of converted waves; b: reflection coefficient for the pS converted wave at the free surface as a function of the angle of incidence (after Fertig, 1984). ........................................... 143
4.2 a: correlation between Vp/Vs, and K/μ. Note the relative linear relation between Vp/Vs and K/μ especially at large Vp/Vs (after Tatham, 1985); b: left: Vp/Vs vs porosity; right: interval transit time for P- and S-wave velocity trends vs porosity in water saturated sandstones (after Tatham, 1985); c: observed Vp/Vs vs porosity for gas and water saturated well consolidated sedimentary rocks and confining pressure between 0 and 8.9x10^8 Pa (after Gregory, 1976). ........................................... 146
4.3 Trace-normalized band-pass filtered (3-10 Hz) S-wave record section for shot-point VIC recorded to the east along the cross-rift line D. Vp=3.46 km/s (observed phase notation: s=S, t=ST(S), t2=SL(S), u=SM(S)). ........................................... 149
4.4 Trace-normalized band-pass filtered (4-10 Hz) S-wave record section for shot-point KAP recorded to the west along the cross-rift line D. Vp=3.46 km/s (observed phase notation for Fig. 4.4, 4.5: s=S, t=ST(S), u=SM(S)). ........................................... 150
4.5 Trace-normalized band-pass filtered (4-10 Hz) S-wave record section for shot-point KAP recorded to the east along the cross-rift line D. Vp=3.46 km/s (observed phase notation: see Fig. 4.4 caption). ........................................... 151
4.6 Trace-normalized band-pass filtered (1-8 Hz) S-wave record section for shot-point BAR recorded to the west along the cross-rift line D. Vp=3.46 km/s (observed phase notation for Fig. 4.6, 4.7: s=S, t=ST(S), t2=ST(S), u=SM(S), u2=unidentified phase discussed in the text). ........................................... 153
LIST OF FIGURES

4.7 Trace-normalized band-pass filtered (1-8 Hz) S-wave record section for shot-point BAR recorded to the east along the cross-rift line D. $V_o = 3.46$ km/s. (observed phase notation see Fig. 4.6 caption). ................................................................. 154

4.8 Trace-normalized band-pass filtered (3-10 Hz) S-wave record section for shot-point TAN recorded to the west along the cross-rift line D. $V_o = 3.46$ km/s (observed phase notation for Fig. 4.8, 4.9: $s=S_g$, $t_1=S_{nl}S$, $t=S_mS$, $u=S_mS$, unidentified phase discussed in the text). ........................................................................................................ 155

4.9 Trace-normalized band-pass filtered (3-10 Hz) S-wave record section for shot-point TAN recorded to the east along the cross-rift line D. $V_o = 3.46$ km/s (observed phase notation see Fig. 4.8 caption). ........................................................................................................ 156

4.10 Trace-normalized band-pass filtered (3-12 Hz) S-wave record section for shot-point BAS recorded to the west along the cross-rift line D. $V_o = 3.46$ km/s (observed phase notation for Fig. 4.10, 4.11: $s=S_g$, $t_1=S_{nl}S$, $u=S_mS$). ........................................................................ 158

4.11 Trace-normalized band-pass filtered (3-12 Hz) S-wave record section for shot-point BAS recorded to the east along the cross-rift line D. $V_o = 3.46$ km/s (observed phase notation: see Fig. 4.10 caption). ........................................................................................................ 159

4.12 Phase ray-traced final model (for velocity distribution see Fig. 4.17. Number of rays for each phase limited to a maximum of 10 for display). Vertical exaggeration 10:1. ......................................................................................... 161

4.13 $t_1$ and $t_2$ phases ray-traced final model (for velocity distribution see Fig. 4.17. Number of rays for each phase limited to a maximum of 10 for display). Vertical exaggeration 5:1. ......................................................................................... 162

4.14 S phase ray-traced final model (for velocity distribution see Fig. 4.17. Number of rays for each phase limited to a maximum of 10 for display). Vertical exaggeration 5:1. ......................................................................................... 164

4.15 Amplitude of observed P diving waves in the uppermost mantle and noise amplitude in the S time window. The $S_n$ theoretical amplitudes are calculated from $P_n$ using amplitude ratios derived from reflectivity modelling. ......................................................................................... 166

4.16 Top: S ray trace coverage. Bottom: P ray trace coverage. ........................................................................................................ 167

4.17 Final 2-D ray-trace model for KIRIS 90 cross-rift line D. Dashed lines indicate transitions where there is a change in velocity gradient but no significant discontinuity. S-wave velocities in km/s; Poisson's ratio in italic. ......................................................................................... 168

4.18 a: Histograms of cross-sectional area of lower crust versus $\sigma$ in rifts and shields; darker shading indicates higher quality factor; bars at the top indicate velocity range; b: Comparison of field- and laboratory-measured $P$-velocities and $\sigma_x$. Field measured data from several areas (cross-hatched ovals) are plotted on fields of lab-measured data from a compilation by Holbrook et al. (1992). R=Rift Zone; P=Fissure; P=Paleozoic crust. Numbers are laboratory measured data from lower-crustal rock types: 1=Quartzite (granulite); 2=Felsic amphib. gneiss; 3=Felsic granulite; 4=Quartz-mica schist; 5=Intermediate granulite; 6=Anorthosite; 7=Felsic granulite; 8=Amphibolite; 9=Metagranite (granulite); 10=Pyroxenite; 11=Eclogite; 12=Dunit/Pegmatite. ........................................................................................................ 172

4.19 Trace-normalized band-pass filtered (2-10 Hz) S-wave record section for shot-point CHF recorded along the flank-rift line E. $V_o = 3.46$ km/s (observed travel times notation: $s=S_g$, $t_1$=reflection from an upper crustal reflector, $t=S_{nl}S$, $u=S_mS$). ........................................................................................................ 174

4.20 Trace-normalized band-pass filtered (2-10 Hz) S-wave record section for shot-point ILA recorded to the west along the flank-rift line E. $V_o = 3.46$ km/s (observed travel times notation for Fig. 4.20, 4.21: $s=S_g$, $t_1$=reflection from an upper crustal reflector, $t=S_{nl}S$, $u=S_mS$). ........................................................................................................ 176
LIST OF FIGURES

4.21 Trace-normalized band-pass filtered (2-10 Hz) S-wave record section for shot-point ILA recorded to the east along the flank-rift line E. $V_s = 3.46$ km/s (observed travel times notation: see Fig. 4.20 caption). ........................................ 177

4.22 Trace-normalized band-pass filtered (1-10 Hz) S-wave record section for shot-point LTS recorded to the east along the flank-rift line E. $V_s = 3.46$ km/s (observed travel times notation: $t = S_1$, $u = S_2, S_3$). ........................................................................................................ 178

4.23 Trace-normalized band-pass filtered (3-12 Hz) S-wave record section for shot-point LTC recorded to the east along the flank-rift line E. $V_s = 4.62$ km/s (observed travel times notation: $t_2 = S_1$, $u = S_2, S_3$). ........................................................................................................ 179

4.24 $t$, $t_1$, and $t_2$ phase ray-traced final model (for velocity distribution see Fig. 4.26. Number of rays for each phase limited to a maximum of 10 for display). Vertical exaggeration 10:1. 181

4.25 $t_2$ and $u$ phase ray traced final model (for velocity distribution see Fig. 4.26. Number of rays for each phase limited to a maximum of 10 for display). Vertical exaggeration 5:1. 182

4.26 Final 2-D ray-trace model for the KRISP 90 flank line E. Dashed lines indicate transitions where there is a change in velocity gradient but no significant discontinuity. S-wave velocities in km/s; Poisson's ratio in italic. .................................................... 183

5.1 Function $f(t)$ versus $2\pi f M t$, for various $\gamma$ and $\psi$. For $\psi = 0$, the impulse is symmetrical; for $\psi = \pm \pi/2$ the impulse is antisymmetrical (after Červeny et al., 1977). ........................................ 190

5.2 Synthetic seismograms calculated with SEIS81. $f_M = 5$ Hz, dominant frequency; $\gamma = 4$, no of extrema in the source wavelet; $t_0 = 0$, time; $\psi = \pi/2$, phase shift. $V_r$ is 8.0 km/s, the time shift of the envelope maximum is 0.3 sec. .................................................... 191

5.3 a: Point source response over a layered medium (after Fuchs, 1980). b: Point source Q and observation point P over a layered medium. The compressional reflection from the reflecting zone suffers elastic transmission losses and time shifts in the layers 1 to m (after Fuchs and Müller, 1971). .................................................... 193

5.4 Frequency-angle window (after Fuchs, 1980). ........................................................................ 195

5.5 Comparison of synthetic seismograms computed with the reflectivity method and the SRM. The model is two layered: $V_{p1} = 1.8$ km/s, $V_{s1} = 1.2$ km/s, $h_1 = 0.6$ km, $\rho_1 = 1.8$ g/cm$^3$, $V_{p2} = 2.3$ km/s, $V_{s2} = 1.35$ km/s, $h_2 = \infty$, $\rho_2 = 2.0$ g/cm$^3$. a: reflectivity section; b: SRM section. Both sections are NMO corrected; the data are corrected for spherical divergence. c: amplitude versus distance plot. .................................................... 197

5.6 Reduced travel time diagram for the reflecting zone of an arbitrary model. The theoretical seismograms have to be computed inside the rectangle (after Fuchs and Müller, 1971).................................................... 190

5.7 Observed and calculated $a$ amplitude distribution from shot point VIC. The vertical bar indicates a factor of four variation. The amplitudes in the current figure are corrected to an absolute particle velocity with units of cm/s. .................................................... 201

5.8 Observed and calculated $a$ amplitude distribution from shot point KAP. The vertical bar indicates a factor of four variation. The amplitudes in the current figure are corrected to an absolute particle velocity with units of cm/s. .................................................... 202

5.9 Observed and calculated $a$ amplitude distribution from shot point BAR. The vertical bar indicates a factor of four variation. The amplitudes in the current figure are corrected to an absolute particle velocity with units of cm/s. .................................................... 203
LIST OF FIGURES

5.10 Observed and calculated amplitude distribution from shot point TAN. The vertical bar indicates a factor of four variation. The amplitudes in the current figure are corrected to an absolute particle velocity with units of cm/s. .............................................. 204

5.11 Observed and calculated amplitude distribution from shot point BAS. The vertical bar indicates a factor of four variation. The amplitudes in the current figure are corrected to an absolute particle velocity with units of cm/s. .............................................. 205

5.12 Observed and calculated amplitude distribution from shot point CHF. The vertical bar indicates a factor of four variation. The amplitudes in the current figure are corrected to an absolute particle velocity with units of cm/s. .............................................. 206

5.13 Amplitude distance curve for model shown in the inset upper right with 0.0, 0.2, 2.0 and 5.0 km of low-velocity material overlying the basement (after Banda et al., 1982). 208

5.14 1-D velocity depth functions used for the a amplitude modelling. .......................... 210

5.15 a: Ray-trace diagram for the x phase showing the possible fit with a multiple in the midcrust (1) and with a reflection from the uppermost mantle (2). b: Ray-trace diagram for the x phase showing the fit with a multiple from the Moho interface in the basement cover. 213

5.16 a: Space-time window from the finite-difference record section calculated from TAN showing the theoretical x phase. b: Space-time window from the reflectivity record section calculated from TAN showing the theoretical x phase. .................................................. 214

5.17 Top: synthetic seismograms calculated with the reflectivity method for the 1-D VIC model in the upper right. Bottom: trace-normalized band-pass filtered (5-10 Hz) S-wave record section for shot-point VIC recorded to the east along the cross-rift line D. $V_p=3.46$ km/s. 219

5.18 Top: synthetic seismograms calculated with the reflectivity method for the 1-D KAP model in the upper right. Bottom: trace-normalized band-pass filtered (4-10 Hz) S-wave record section for shot-point KAP recorded to the west along the cross-rift line D. $V_p=3.46$ km/s. 220

5.19 Top: synthetic seismograms calculated with the reflectivity method for the 1-D BAR model in the upper left. The arrow indicates an unwanted numerical phase generated by reflectivity. Bottom: trace-normalized band-pass filtered (1-6 Hz) S-wave record section for shot-point BAR recorded to the west along the cross-rift line D. $V_p=3.46$ km/s. 222

5.20 Top: synthetic seismograms calculated with the reflectivity method for the 1-D TAN model in the upper left. Bottom: trace-normalized band-pass filtered (3-10 Hz) S-wave record section for shot-point TAN recorded to the west along the cross-rift line D. $V_p=3.46$ km/s. 223

5.21 Top: synthetic seismograms calculated with the reflectivity method for the 1-D BAS model in the upper left. Bottom: trace-normalized band-pass filtered (3-12 Hz) S-wave record section for shot-point BAS recorded to the east along the cross-rift line D. $V_p=3.46$ km/s. 225

5.22 Finite-difference synthetic seismogram section for shot point VIC recorded to the east along the cross-rift line D. Grid spacing 40 m. Dominant frequency 4.0 Hz. $V_p=6$ km/s. Correlated phases from ray-tracing as for Fig. 3.2 caption. ................................. 232

5.23 Finite-difference synthetic seismogram section for shot point VIC recorded to the east along the cross-rift line D. Grid spacing 40 m. Dominant frequency 4.0 Hz. $V_p=8$ km/s. Correlated phases from ray-tracing as for Fig. 3.4 caption. ................................. 234

5.24 Finite-difference synthetic seismogram section for shot point KAP recorded to the west along the cross-rift line D. Grid spacing 40 m. Dominant frequency 4.0 Hz. $V_p=6$ km/s. Correlated phases from ray-tracing as for Fig. 3.5 caption. ................................. 235
5.25 Finite-difference synthetic seismogram section for shot point KAP recorded to the east along the cross-rift line D. Grid spacing 40 m. Dominant frequency 4.0 Hz. $V_p=6$ km/s. Correlated phases from ray-tracing as for Fig. 3.6 caption.

5.26 Finite-difference synthetic seismogram section for shot point BAR recorded to the west along the cross-rift line D. Grid spacing 40 m. Dominant frequency 4.0 Hz. $V_p=6$ km/s. Correlated phases from ray-tracing as for Fig. 3.7 caption.

5.27 Finite-difference synthetic seismogram section for shot point BAR recorded to the east along the cross-rift line D. Grid spacing 40 m. Dominant frequency 4.0 Hz. $V_p=6$ km/s. Correlated phases from ray-tracing as for Fig. 3.8 caption.

5.28 Finite-difference synthetic seismogram section for shot point TAN recorded to the west along the cross-rift line D. Grid spacing 40 m. Dominant frequency 4.0 Hz. $V_p=6$ km/s. Correlated phases from ray-tracing as for Fig. 3.9 caption.

5.29 Finite-difference synthetic seismogram section for shot point TAN recorded to the east along the cross-rift line D. Grid spacing 40 m. Dominant frequency 4.0 Hz. $V_p=6$ km/s. Correlated phases from ray-tracing as for Fig. 3.10 caption.

6.1 Record section for the Lake Victoria (VIC) shot point. $V_p=8.0$ km/s. Band pass filter 1-20 Hz. The traces in the shaded areas are stacked in Fig. 6.5, 6.6.

6.2 NMO corrected record section for the Lake Victoria (VIC) shot point. NMO constant velocity correction=6.9 km/s.

6.3 Seismic records and stacked traces from the axial line. a: $d_1$ refraction and $d_1$ reflection; b: $d_1$ and $d_2$ reflections (after Keller et al., 1994a).

6.4 Correlated Adaptive Noise Cancelling Concept (CANC); modified from Hattingh (1988).

6.5 a: CANC output seismograms for location numbers 31 to 42 and their stack for the $d_1$ phase. b: CANC output seismograms for location numbers 31 to 42 and their stack for the $d_1$ phase.

6.6 a: CANC output seismograms for location numbers 68 to 74 and their stack for the $d_1$ phase. b: CANC output seismograms for location numbers 68 to 74 and their stack for the $d_1$ phase. It should be noted that the filter length may be different and in some cases the traces used are not the same for $d_1$ and $d_2$.

6.7 Example of CANC output stacked seismograms $d_1$ cross-correlated with a time shifted $d_1$ (location number 27 to 45) top: $d_1$ with a normal polarity $d_1$; bottom: $d_1$ with an inverted polarity $d_1$. The maximum amplitude of the cross-correlation function is indicated.

6.8 Ray-trace diagram for the intra-mantle reflection $d_1$ for a model with a LVZ (bottom) and corresponding travel time diagram (top).

6.9 Ray-trace diagram for the intra-mantle reflection $d_1$ for a model without a LVZ (bottom) and corresponding travel time diagram (top).

6.10 Synthetic seismograms for $d_1$ and $d_2$ phases for the model with the LVZ (top) and without the LVZ (bottom) calculated by using a finite-difference technique (Sandmeier, 1990).
6.11 Comparison of observed and theoretical amplitude ratios for the model with and without the LVZ ................................................................. 268

6.12 a: Ternary diagram showing the relationship between P-wave velocity and mineralogical composition P1 for: a: a pure isotropic model at 63 km depth. b: a transverse isotropic model at 63 km depth. Note that although P1 does not have velocities exceeding 8.35 km/s compositions with 75-85% olivine have velocities of about 8.6 km/s. c: a transverse isotropic model at 59 km depth. Note the similarity in the velocity for P1 between Fig. 6.12b and 6.12c ............................................................................................................................................ 271

6.13 Ternary diagrams showing the relationship between P-wave velocity and mineralogical composition P3 for: a: a pure isotropic model at 63 km depth. b: a transverse isotropic model at 63 km depth. Note that although P3 does not have velocities exceeding 8.45 km/s compositions with 70-80% olivine have velocities of around 8.6 km/s. c: a transverse isotropic model at 59 km depth. Note the similarity in the velocity for P3 between Fig. 6.13b and 6.13c ...................................................................................................................................... 272

6.14 Velocity model of the cross-rift (top) and axial (bottom - distance scale halved) profiles derived from ray-trace and synthetic seismograms modelling. P-wave velocities in km/s. The reflector at a depth of c.60 km beneath the axis of the rift on the cross-rift line is obtained from the axial line model .................................................................................................. 275

7.1 A simplified example of a linear travel time inversion scheme, using a ray-trace forward step and a least squares minimization technique. ................................................................. 283

7.2 Trade-off curve resolution and variance for a given discretization of a continuous function (modified after Menke, 1984). The damping factor should be chosen to optimize the trade-off between parameter resolution and model stability ................................................................ 284

7.3 Inversion routine, iteration sequence and file structure ...................................................................................................................... 286

7.4 Velocity model for the upper crust resulted from the travel-time inversion of line D. Only selected velocity values are included in the diagram. (Velocities in km/s) ................................................................. 292

7.5 Traveltime inversion modelling for shallow velocity model for the a and b phases: a: observed (vertical bars are picking errors) and calculated (lines) travel-times. b: raypath coverage for the phases used in the travel-time inversion ......................................................................................................... 294

7.6 Velocity model for the upper and lower crust resulting from the travel-time inversion of line D. Only selected velocity values are included in the diagram. (Velocities in km/s) ................................................................................................................ 295

7.7 Traveltime inversion modelling for mid-crustal velocity model for the b1 and b2 phases. Shallow velocities were fixed using the model in Fig. 7.4. a: observed (vertical bars are picking errors) and calculated (lines) travel-times. b: raypath coverage for the phases used in the travel-time inversion ................................................................................................................................. 297

7.8 Final velocity model resulted from the travel-time inversion of KRISP 90 cross-rift line D. Only average velocities are reported for the crust, and selected velocity values in the upper mantle (Velocities in km/s). ................................................................................................................ 298

7.9 Traveltime inversion modelling for the deep crustal velocity model for the c and d phases. Shallow velocities were fixed using the model in Fig. 7.6. a: observed (vertical bars are picking errors) and calculated (lines) travel-times. b: raypath coverage for the phases used in the travel-time inversion ........................................................................................................................................................................ 300

7.10 The observed and calculated gravity profiles along the cross-rift line D showing the density model B3 derived from the final inverted seismic model. (Densities in g cm^-3). ................................................................................................................ 304
LIST OF FIGURES

7.11 The observed and calculated gravity profiles along the cross-rift line D showing the density model derived from the model in Fig. 3.25 (model A) without the high velocity block beneath BAS. (Densities in g cm$^{-3}$). .............................................. 306

8.1 Fence diagram summarizing the models from the KRISP 90 seismic studies. For line locations see Fig. 2.1; relative velocity variations from teleseismic delay time studies across the rift at 1°S (after Keller et al., 1994b). .............................................. 314

8.2 Different styles of extensional deformation expected with fast and slow extension rates (after Kuszniir and Park, 1987). .............................................. 318
## List of Tables

1. **Summary of the tectonothermal events of the MOB around KRISP 90 cross-rift line.** ......................................................... 15
2. **Cenozoic volcanic associations in the area of interest.** ................................................................. 19
3. **Line D shot point locations.** .................................................................................................................. 71
4. **Line D shot point characteristics.** ........................................................................................................... 72
5. **Phase analysis results; the numbers in the columns are the ranges in km where each phase is observed.** ......................................................... 105
6. **Velocity versus density values used in the gravity modelling.** ......................................................... 138
7. **Phase analysis results; the numbers in the columns are the ranges in km where each phase is observed.** ......................................................... 157
8. **Velocity versus density values used in modelling the gravity data in the forward model A and inverted model B.** ......................................................... 306
List of Plates

Photo 1: The Nandi escarpment; view from the SW. .......................................................... 54

Photo 2: Nyanzian intrusives at c.80 km along line D. ......................................................... 84

Photo 3 Elgeyo escarpment from the Rift Valley floor. ....................................................... 84

Photo 4 Hornblende-biotite gneisses of the basement system at the Elgeyo escarpment. .......................................................... 85

Photo 5 Miocene Uasin Gishu Phonolites from the Rift Valley floor, at c.0.5° N. .............. 85

Photo 6: Miocene Elgeyo volcanics formations at the Elgeyo escarpment. ....................... 86

Photo 7: Pliocene Kabarnet Trachytes at the Kamasia Range. ........................................ 86
### List of Abbreviations

<table>
<thead>
<tr>
<th>Abbreviation</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>ANC</td>
<td>Archaean Nyanza Craton</td>
</tr>
<tr>
<td>BAR</td>
<td>Lake Baringo shot point</td>
</tr>
<tr>
<td>BAS</td>
<td>Barsalanga shot point</td>
</tr>
<tr>
<td>BGS</td>
<td>British Geological Survey</td>
</tr>
<tr>
<td>CANC</td>
<td>Correlated Adaptive Noise Cancellation</td>
</tr>
<tr>
<td>CFB</td>
<td>Continental Flood Basalts</td>
</tr>
<tr>
<td>CHF</td>
<td>Chanlers Falls shot point</td>
</tr>
<tr>
<td>DLS</td>
<td>Damped Least Squares</td>
</tr>
<tr>
<td>EAP</td>
<td>East African Plateau</td>
</tr>
<tr>
<td>ftp</td>
<td>file transfer protocol</td>
</tr>
<tr>
<td>FD</td>
<td>Finite-difference</td>
</tr>
<tr>
<td>GBM</td>
<td>Gaussian Beam Method</td>
</tr>
<tr>
<td>GPS</td>
<td>Global Positioning System</td>
</tr>
<tr>
<td>HF</td>
<td>High Frequency</td>
</tr>
<tr>
<td>HVB</td>
<td>High Velocity Block</td>
</tr>
<tr>
<td>ILA</td>
<td>Illaut shot point</td>
</tr>
<tr>
<td>KAP</td>
<td>Kaptagat shot point</td>
</tr>
<tr>
<td>KRISP</td>
<td>Kenya Rift International Seismic Project</td>
</tr>
<tr>
<td>IPG</td>
<td>Institute de Physique du Globe</td>
</tr>
<tr>
<td>LAB</td>
<td>Lithosphere-asthenosphere boundary</td>
</tr>
<tr>
<td>LAI</td>
<td>Laisamis shot point</td>
</tr>
<tr>
<td>LTC</td>
<td>Lake Turkana Centre shot point</td>
</tr>
<tr>
<td>LTS</td>
<td>Lake Turkana South shot point</td>
</tr>
<tr>
<td>LU</td>
<td>Leicester University</td>
</tr>
<tr>
<td>LVZ</td>
<td>Low Velocity Zone</td>
</tr>
<tr>
<td>MOB</td>
<td>Mozambique Orogenic Belt</td>
</tr>
<tr>
<td>NMO</td>
<td>Normal Move Out</td>
</tr>
<tr>
<td>N, S, E, W</td>
<td>North, South, East, West</td>
</tr>
<tr>
<td>NXC</td>
<td>Normalised cross-correlation</td>
</tr>
<tr>
<td>prm</td>
<td>parameter resolution matrix</td>
</tr>
<tr>
<td>PRM</td>
<td>Paraxial Ray Method</td>
</tr>
<tr>
<td>Qp, Qs</td>
<td>P-wave, S-wave Quality Factor</td>
</tr>
<tr>
<td>rms</td>
<td>root-mean-square</td>
</tr>
<tr>
<td>SEG</td>
<td>Society of Exploration Geophysicists</td>
</tr>
<tr>
<td>SM</td>
<td>Standard Ray Method</td>
</tr>
<tr>
<td>TAN</td>
<td>Tangulbei shot point</td>
</tr>
<tr>
<td>USGS</td>
<td>United States Geological Survey</td>
</tr>
<tr>
<td>VIC</td>
<td>Lake Victoria shot point</td>
</tr>
<tr>
<td>VMS</td>
<td>Virtual Memory System</td>
</tr>
<tr>
<td>Vp, Vs</td>
<td>P-wave velocity, S-wave velocity</td>
</tr>
<tr>
<td>1-D, 2-D, 3-D</td>
<td>One, Two, Three Dimensional</td>
</tr>
</tbody>
</table>

Standard SI unit abbreviations have been used throughout.
Chapter 1

The Kenya Rift

1.1 Introduction

This thesis concentrates on the interpretation of data from KRISP 90. The survey involved three seismic refraction/wide angle reflection profiles along the axis, across the margins, and on the north-eastern flank of the Kenya Rift Valley. The project aimed to resolve questions concerning processes associated with the rifting of continents.

The interpretation of KRISP 90 data requires the understanding of continental rifting processes. Continental rifts are characterized by regional elongation of a zone where the lithosphere has been altered during extension and there has been asthenospheric modification or upwelling beneath the rift axis. Seismic profiling can give information on the structure, seismic velocities and petrological condition of the lithosphere.

Four major Cenozoic Rift Systems are recognised and most studied: the Rhine graben in Central Europe, the Rio Grande Rift in South-western United States, the Baikal Rift in Eastern Siberia and the Kenya Rift in East Africa. The Kenya Rift is regarded as the typical, perhaps the exemplary, active continental rift, implying that the extensional process is presently in action.

Since Gregory's (1896) first geological expedition in the Great Rift Valley generations of scientists have studied this fascinating feature of the earth's crust. A thorough summary of the present state of study on the Kenya Rift is therefore necessary. Current models on rifting processes, geodynamic, geological and tectono-magmatic evolution of the Kenya
Rift are described. The most recent geophysical data and their interpretation are also mentioned.

Particular emphasis is given to the structure of the crust (especially the Precambrian basement), and uppermost mantle across the Rift around equatorial latitudes.

1.2 The Kenya Rift

The Kenya Rift forms the eastern branch of the East African Rift System, itself part of the Afro-Arabian Rift System which extends c.6500 km from Turkey to Mozambique and includes the Dead Sea (Levantine), Red Sea, and Gulf of Aden Rifts (Fig. 1.1).

The eastern and western African branches of the East African Rift System split north of Lake Victoria and rejoin south of the Lake through a broad zone of faults in Tanzania. This bifurcation has been attributed to the presence of the Tanzanian or Nyanza Craton, which acted as a resistant shield to the propagation of the continental fracture (Fig. 1.2).

The volcanic evolution of the East African Rift produced significantly greater amount of volcanic products in the eastern branch than in the western branch. About 220000 km$^3$ of igneous material has been extruded over the last 30 Ma (144000 km$^3$ in Kenya since Oligocene-Miocene time) in the eastern branch, of the 500000 km$^3$ of rift-related volcanics (Barberi et al., 1982; Williams, 1972; 1982).

The eastern branch begins in the north at the Afar triple junction where it joins the Red Sea and Gulf of Aden Rifts and than extends southwards across Ethiopia, Kenya, and northern Tanzania (Baker et al., 1972). The Malawi (Nyasa) Rift to the south has more in common with the western branch and it has been considered part of it (Ebinger, 1989). The two branches are apparently not connected in structure, morphology and seismicity, (Ebinger, 1989), although a number of subsidiary troughs are directed into the Nyanza Craton, such as the Nyanza (Kavirondo) Rift (King, 1978).

The Kenya Rift traverses Kenya from Lake Turkana in the north to Lake Magadi in the south and dies out in northern Tanzania. The graben transects the Kenya Dome, a topographic high rising from 300 m in Turkana to more than 1900 m between Nakuru and Naivasha and decreasing in altitude to 500 m around Lake Magadi. It is considered to be
Figure 1.1: The Afro-Arabian Rift System. EAP=East African Plateau; CAE=Central African Republic; RB=Ruanda-Burundi; MW=Malawi.
Figure 1.2: Top: the Kenya Dome and Kenya Rift. Bottom: simplified topographic profile from the Western Rift to the Indian Ocean along $0 - 5^\circ S$ (modified from Maguire and Khan, 1980).
a local culmination on the eastern rim of the East Africa Plateau elliptical in plan and about 1000 km wide (Fig. 1.2).

The central sector of the Kenya Rift is well defined by a 50-70 km wide half graben, with surface displacements of up to 1600 m on the bounding faults. The faulted structure extends over a width of about 200 km and loses its graben like appearance to the north in Turkana and to the south in northern Tanzania (King, 1978).

1.2.1 Continental Rifts and Models of Origin

There is agreement that rifts are dominated by extensional processes. It is possible to estimate the magnitude of the deviatoric stress required to form a graben in the Earth’s crust. It is more difficult to constrain the cause and effect relationship of the lower lithosphere/asthenosphere system. Two major questions are still under debate concerning the processes involved in continental rifting.

One concerns the mode of extensional strain: is it either irrotational pure shear (Baker and Wohlenberg, 1971; McKenzie, 1978; Davis, 1991) or rotational simple shear (Wernicke, 1981; 1985), or both (Kuszmir et al., 1991)?

The second question involves the role of the asthenosphere. Is the asthenosphere active or passive in the mechanism of rifting (Şengör and Burke, 1978; Turcotte and Oxburgh, 1973)?

**Pure or Simple Shear?**

Fig. 1.3 shows the models of strain geometry in rifts.

A) - In the pure shear model, crust and upper mantle are uniformly attenuated along any given vertical reference line (McKenzie, 1978). There are two main stages:

1. rapid stretching which produces thinning, block faulting and subsidence, and
2. thickening due to the heat conduction to the surface followed by slow subsidence.

It should be noted that there is a close relationship between heat flux and subsidence and that the asthenosphere plays a passive role during the lithospheric thinning.

B) - In the simple shear model the relative extension of crust and mantle lithosphere along any given vertical line is non uniform. The main stages are:

1. a low angle normal fault penetrates most of the crust (and possibly the mantle),
Figure 1.3: Top: end-member models of strain geometry in rifts. a: 'Pure-shear' model. b: 'Simple shear' model (after Wernicke, 1985). Bottom: A schematic representation of lithosphere extension by simple shear in the upper crust and pure-shear in the lower crust and mantle. A- listric fault, B- planar fault (after Kusznir et al., 1991).
developing a surface sedimentary basin,

2- the upper-plate may extend relative to the surrounding region, and

3- isostatic rebound occurs in the unloaded terranes. To note are: (a) the diachronosity between lower plate ductile tectonism and upper plate translation and/or distension and (b) the scale of the shear zone, which may be several hundreds of kilometres wide.

C)- In the flexural-cantilever simple shear-pure shear model faulting is dominant in the upper crust, and distributed plastic deformation occurs in the lower crust and mantle. This model includes:

1- crustal thinning during extension by simple shear in the upper crust and pure shear in the lower crust and mantle,

2- the lithosphere temperature field is perturbed during extension and re-equilibrated after, and

3- a flexural-isostatic response of the lithosphere to crustal thinning and thermal loads both syn- and post- the rifting event.

The Kenya Rift’s evolution has been described in terms of model B:

Bosworth (1987) suggested that simple shear with low-angle detachment systems in the lithosphere could explain a) the asymmetry of sub-basins alternating in polarity along the rift axis and being joined by accommodation zones and b) the off-axis volcanism, occurring where the detachments cut to the base of the lithosphere (Fig. 1.4a-b).

Kazmin (1991) analysed the asymmetric distribution of volcanics and again suggested that Wernicke’s simple shear model could be a suitable model for the Kenya Rift (Fig. 1.4c); nevertheless he disagreed with Bosworth’s idea of accommodation zones, arguing that the distance from Turkana to Elgeyo, part of the same polarity accommodation zone, is too short.

Active or Passive?

The active mechanism (Fig. 1.5a) is related to direct intervention of asthenospheric upwelling due to gravitational instability of the less dense asthenosphere under the more dense mantle lithosphere, and flow occurs in both the lithosphere and asthenosphere. In other words it associates rifting with mantle convection in the form of axisymmetric mantle plumes or linear upwelling associated with mantle convection cells.
Figure 1.4: a: Tectonic interpretation of western Kenya; b: Schematic cross-section through the Kenya Rift from Mount Elgon on the Uganda border to Mount Marsabit in northern Kenya (after Bosworth, 1987). c: formation of continental flood basalts (CFB) in the model with low-angle uniform-sense normal simple shear (after Kazmin, 1991).
Figure 1.5: a: Model of an active rifting mechanism. Ascending mantle convection thins the lithosphere causing doming and lithospheric failure. b: Model of a passive rifting mechanism. Tensinal stresses cause the failure of the continental lithosphere. A mantle diapir penetrates to the base of the crust causing crustal thinning (after Turcotte and Emerman, 1983).

The passive mechanism (Fig. 1.5b) results from plate interaction moving under the influence of large-scale convective flows in the mantle, i.e. the rifting is due to convection at a distance. That is to say it associates rifting with extensional stresses in the lithosphere that extends at zones of weakness; in the passive case the driving force is the plate motion, and the rifting could induce lithospheric thinning and asthenospheric upwelling.

An alternative model has been proposed by Dunbar and Sawyer (1988). It explains the different surface expression of ‘active’ and ‘passive’ rifts as due to the mode of failure of pre-existing weaknesses in the continental lithosphere rather than difference in the nature of driving forces involved.

Models in relation to the Kenya Rift

An ‘active’ scenario includes crustal upwarp; the volcanism precedes the rifting; the asthenospheric intrusion initiating the upwarp is limited in width. For the ‘passive’ hypothesis, the crustal doming and volcanism are secondary processes.

The amount of extension suffered by the crust in Kenya during the rifting process is considered to be about 5%-20% (Keller et al., 1991). The evidence from surface geology
CHAPTER 1. THE KENYA RIFT

is estimated at a maximum of 10 km in the central part of the Kenya Rift (Baker and Wohlenberg, 1971; King, 1978); from geophysics, plate tectonics, and mathematical modelling the estimated extension is generally greater (McKenzie et al., 1970; KRISP Working Party, 1991), and the differences have been used to develop different models for the rifting in Kenya.

Bosworth et al. (1986) and Bosworth's (1987; 1989) model could be considered 'passive', involving rotational simple strain as a response to intraplate forces. Break-up of the African plate, stationary during the last 25 Ma, is controlled by low-angle sub rift detachments and not vertical cracks. Although important, hotspots and domal uplift are not essential to the rifting process. The central Kenya hot spot is responsible for the large volume of volcanics in the rift but does not directly control its location (Fig. 1.4a).

Girdler (1983) proposes a model in which rifting induces thinning and not vice versa. He observes (1) from a study of the distribution of faulting and seismically active areas in relation to lithospheric thinning (deduced from Bouguer anomaly maps) that there are seismically active areas in Africa where there is no lithospheric thinning, and (2) uplift following faulting in the Gulf of Aden. He concludes that horizontal stresses cause brittle fractures allowing hot material to rise from the asthenosphere.

Mohr's (1987) asthenolithic intrusion model results from a thermal anomaly within the lithosphere. It can be described as 'active' rifting. The forced diapiric intrusion would activate reverse décollements in the lithospheric mantle and subsequently in the crust and at the Moho (Fig. 1.6).

Most authors prefer the 'active' hypothesis, in the light of magmatic (Sengör and Burke, 1978; Karson and Curtis, 1989; Tatsumi and Kimura, 1991; Mohr, 1992); and teleseismic studies (Davis, 1991) and stratigraphic/structural evidence (Smith, 1994). These will be considered in more detail later.
Figure 1.6: Proposed hypothesis of continental rifting in eastern Africa, shown in cross-section (vertical and horizontal scales equal). a: protorift downwarp, with associated flood basalts fed from newly initiated asthenospheric diapir intruding base of lithosphere. b: rifting and penetration of lower crust by asthenolith. Profuse silicic volcanism derived from fractionates at top of asthenolith, with additional remelting of underplated crust. c: rapid ascent of asthenolith. Underplating causes crustal thickening under rift margins (after Mohr, 1987).
CHAPTER 1. THE KENYA RIFT

1.3 The Geology of Kenya along the KRISP 90 E-W Profile

Different lithologies and structural patterns influence the seismic velocities and amplitude content of the different phases. A detailed description of the geological history of the central Kenya Rift is presented.

1.3.1 Stratigraphy

The geology of the area surrounding the KRISP 90 cross-rift line will be described in detail. The zone is located on either side of the equator from Lake Victoria in the west to the northern slopes of Mt. Kenya in the east (from 34° to 38°E in longitude) (Fig. 1.7).

The area includes outcrops ranging in age from the Precambrian to Recent. The basement is a complex of metamorphic and igneous rocks of mainly Precambrian age. A cratonic area, the Archaean Nyanzian Craton (ANC) at the western end of the line, includes both metamorphic and intrusive rocks.

The ANC is overthrust by the Mozambique Orogenic Belt (MOB) of Proterozoic age, at the Nandi fault. The MOB migmatites and high grade metasediments, with a dominant north-south trend, form the basement to the eastern end of the seismic line (Shackleton and Ries, 1984; Mosley, 1993; Smith and Mosley, 1993).

The volcanic activity related to the rifting process started in this portion of the rift in Early Miocene time and continues to the present. The oldest volcanic outcrops in this area are represented by the main phonolitic plateaux of Middle Miocene; these now outcrop on the rift shoulders. Pliocene to Recent volcanics are visible on the rift floor.

Voluminous volcanism infilled the depression created by the rifting almost as fast as the floor subsided. This resulted in the Kenya graben being characterized by shallow, ephemeral lake basins.

Sediments can be found associated with early tectonic events, of basement provenance, or interbedded in the volcanics, deposited during the subsidence related to the rifting.

Sediments are exposed in the western end of the line (Lake Victoria) and in the basins on the rift floor.
Figure 1.7: Simplified geological map of the area surrounding the KRISP 90 cross-rift line.
CHAPTER 1. THE KENYA RIFT

The Precambrian Basement

Archaean rocks outcrop at the western end of the cross-rift line, although the limit of the buried craton might extend further to the east (Chorowicz et al., 1987; Smith and Mosley, 1993).

The ANC is composed of low-grade volcano-sedimentary/ophiolitic terranes (c.2.9-2.5 Ga); lithologies range from basic to intermediate and acid and mainly consist of granulites, greenstone associations, and granitoids including narrow metasedimentary fold belts.

The Nandi fault zone involves the interslicing of lithologies from both the ANC and MOB and marks their tectonic contact (Photo 1). The contact may involve a crustal-scale thrust system and has been compared to the Insubric strike-slip fault marking the contact between the Alpine and African plates in the Alps (Fig. 1.8 after Mattauer, 1986; Mosley, 1993).

The history of the MOB terrane in this area is summarised in Table 1.1 (modified from...
CHAPTER 1. THE KENYA RIFT

Hackman et al., 1989). The MOB is suggested to include Archaean rocks (2.5 Ga) and younger nappes (900 Ma).

The MOB has been affected by a complex sequence of metamorphic episodes as a consequence of the collision between the ANC in the west and a Kibaran craton in the east. This is interpreted as a continent-continent collision. Large scale recumbent structures indicate a complete collisional sequence. The NS suture zone is oblique to the stretching lineations, striking NW-SE. An initial NW-SE transpressive plate motion was followed by displacement parallel to the suture zone (Shackleton and Ries, 1984; Berhe, 1990).

Ultramafic and ophiolitic bodies are present in the MOB along zones of N-S and NNW-SSE sinistral shearing. Ophiolites are interpreted as remnants of a late Precambrian ocean in Kenya (Vearncombe, 1983; Berhe, 1990), or alternatively as completely allochthonous and emplaced during orogenesis by southerly directed thrust sheets (Mosley, 1993; Smith and Mosley, 1993).

Linked metamorphic and deformational phases can be recognised. Collision is related to a tectonothermal (Samburuan-Sabachian) event. The Samburuan phase involves high metamorphic grade, extensive feldspathisation and migmatisation. The Sabachian (850-820 Ma) represents regional horizontal tectonism, with emplacement of granites and metabasic dykes.

A series of "Pan-African" events (Baragoian-Barsaloian and Loldaikan between 630-550 Ma) and post- orogenic uplift is recorded. Localised variable melting of lower crustal rocks is recorded, and high-grade metaigneous rocks are emplaced in the upper crust. Linked metamorphic and deformational phases can be recognised. Collision is related to a tectonothermal (Samburuan-Sabachian) event. The Samburuan phase involves high metamorphic grade, extensive feldspathisation and migmatisation. The Sabachian (850-820 Ma) represents regional horizontal tectonism, with emplacement of granites and metabasic dykes.

Table 1.1: Summary of the tectonothermal events of the MOB around KRISP 90 cross-rift line.

<table>
<thead>
<tr>
<th>Events</th>
<th>Accompanying activity</th>
<th>thermal-igneous activity</th>
<th>Tectonic level-control</th>
</tr>
</thead>
<tbody>
<tr>
<td>KIPINGIAN</td>
<td>High tectonic levels; post orogenic uplift.</td>
<td></td>
<td></td>
</tr>
<tr>
<td>LOLDAIAN</td>
<td>Local melting; minor felsic veins</td>
<td>High tectonic levels; same stress as Barsaloian.</td>
<td></td>
</tr>
<tr>
<td>BARSALOIAN</td>
<td>High tectonic levels; same stress as Barsaloian.</td>
<td></td>
<td></td>
</tr>
<tr>
<td>BARAGIOIAN</td>
<td>High tectonic levels; ENE-WNW compression to develop en echelon shear zones preferentially on Baragoian fold limbs.</td>
<td></td>
<td></td>
</tr>
<tr>
<td>SABACHIAN</td>
<td>High tectonic levels; E-W tension across the orogen with gravity controlled deformation.</td>
<td></td>
<td></td>
</tr>
<tr>
<td>SAMBURUAN</td>
<td>High tectonic levels; E-W tension across the orogen with gravity controlled deformation.</td>
<td></td>
<td></td>
</tr>
<tr>
<td>MUROGODO</td>
<td>High tectonic levels; E-W tension across the orogen with gravity controlled deformation.</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Table 1.1: Summary of the tectonothermal events of the MOB around KRISP 90 cross-rift line.
Ma) represents post-collision deformation with ductile shear and intrusion of syntectonic granites. This results in a structural pattern continuous with the lower grade volcanic-sedimentary sequences more typical of ANC.

The final phase (550-520 Ma) reflects general cooling throughout the orogen, with folding and brittle shear deformation (Shackleton, 1986; Vail, 1988; Key et al., 1989; Mosley, 1993). Isotopic data, P-T-t paths, and structural styles indicate that up to 430 Ma a slow uplift of the lower part of a tectonically thickened crust occurred. Proterozoic rocks at the surface originated from depths in excess of 20 km within an isostatically exhumed crustal sequence (Smith and Mosley, 1993).

Recent works have subdivided the MOB rocks in three sectors (Western, Central and Eastern) representing structural/mechanical differences 'most likely of lithospheric extent' (Fig. 1.9 after Mosley, 1993). The contacts between these sectors are marked by late-stage steep, upright, ductile shear zones often with a sinistral shear sense, although there is a general concordance in the metamorphic and structural history of the sectors.

The Nandi fault and more generally the Aswa shear zone marks the western boundary of the MOB Western sector, which is composed of a W or NW verging sequence of tectonically repeated lithological units; kinematic indicators are representative of an initial W or SW directed thrusting and shearing episode, culminating in the overthrusting of large nappes onto, and possibly tectonically incorporated into, parts of the western foreland (ANC).

The Central sector is proposed as 'the most mobile part' of the orogen. It is a complex of near horizontal thrust stacks with axial traces trending generally ESE-WNW (part of the Sabachian event) and involving Baragoian-Barsaloian events. On the eastern shoulder of the rift a relatively massive 'basement' ridge exists. At structurally lower levels migmatites and granitoids gneisses are common, and form a N-S trending block that acts as a rigid buttress to the main Barsaloian ductile shear juxtaposed with its eastern boundary. This differential response to deformation suggests this block might be a 'lenticle of older crust'. The adjacent Mukogodo migmatite, of Kibaran age, supports this hypothesis. To the east of the main Barsaloian ductile zone W verging, E dipping thrusts involve paragneissess and Mukogodo migmatites (Table 1.1); structurally beneath...
Figure 1.9: Principal Proterozoic structures in Kenya with locations (after Mosley, 1993).
this stack a granodioritic body is emplaced. Further east, outside the area considered here the vergence changes to the E; W dipping units with abundant limestones and more extensive metapsammitic/gneisses occur.

The Eastern sector is beyond the eastern end of the cross-rift line. It is composed of E-verging nappes of MOB association and post-tectonics intrusions.

A major late-stage structural event affects all three sectors with a sinistral NW-SE trending shear event and a conjugate NE-SW weaker shear set.

These NW-SE faults gave a strong segmentation to the MOB (and early Proterozoic and Archaean gneisses in the western craton) at the post-orogenic stage following the major N-S ductile shearing event of Shackleton and Ries (1984), and are suggested to indicate the major stress transmission at the final stage of the collision. They might extend to lithospheric depths, acting as stress guides during the Tertiary, when their reactivation affected the inception of rifting within the Kenya Rift (Smith and Mosley, 1993; Hetzel and Strecker, 1994).

Cenozoic Volcanics

The volcanic activity connected with the rifting process began in Oligocene-Miocene times. The Kenya Rift is characterised by bimodal volcanism and evolution from strongly sodic-alkaline and basaltic volcanics through more evolved and felsic products (Barberi et al., 1982).

Both mafic and felsic rocks show a general decrease in silica undersaturation with time. The volcanic activity shows an eastward shift from the Uganda-Kenya border in the Lower Miocene towards the rift axis in the Miocene-Pliocene and a concentration of igneous eruptives in and near the rift axis (Keller et al., 1991).

The Cenozoic volcanic activity in the area of interest is summarised in Table 1.2, modified from Hackman (1988); subordinate components are in brackets.
Table 1.2: Cenozoic volcanic associations in the area of interest.

Details on the thickness of the various formations will be given in the geological interpretation of the seismic cross-section (Section 3.4.2).

The Rift Valley and adjacent shoulder plateaux include volcanic complexes from Miocene to Recent age involving lava flows and pyroclastics of central volcanoes and plateau lavas, with minor associated intrusives and interbedded sediments (Baker et al., 1972; Chapman and Brook, 1978; Williams, 1982; Hackman et al., 1990; Fig. 1.10-1.11).

In general the chemistry of the volcanics was sodic-alkaline in the Miocene. The extrusives range from nepheline-normative basalts in proto-rift depressions (Early Miocene) to the voluminous plateau phonolites which filled and overflowed the existing rift structure.
Figure 1.10: Geological map of the Kenya Rift volcanics (locality names in fig 1.11). After Baker et al. (1971).
CHAPTER 1. THE KENYA RIFT

Figure 1.11: Locality names for fig 1.10. After Baker et al. (1971).
between about 14-11 Ma on the western side. Overspill on the eastern side is not demonstrated, but phonolites of Middle-Late Miocene are present also on the eastern plateau.

In the Pliocene the trend was somewhat less alkaline, ranging from alkali basalts with some tholeiitic affinities (6.5-5 Ma) to trachyphonolites, phonolitic trachytes and soda-trachytes (7-5 Ma). From about 5-2 Ma flood trachytes and ignimbrites were erupted, overspilling the rift structure for the second time (Keller et al., 1991).

Quaternary volcanic activity included trachytic caldera volcanoes along the Rift Valley floor.

Quaternary Sediments

In the Kenya Rift the great volume of volcanics infilled the developing depression nearly as fast as it subsided. This resulted in the presence of shallow and ephemeral lake basins different to the deep, long-lived lakes of the Western Rift (Baker, 1986). However sediments outcrop in different areas along the KRISP 90 cross-rift line.

Quaternary sediments cover the area surrounding Lake Victoria.

Fluvial deposits occur on the basement outcrop on both sides of the Rift. Sedimentary fill of unknown depth (Chapman et al., 1978) lies at the base of the Elgeyo escarpment in the Kerio Valley.

Sub-volcanic sediments of basement provenance outcrop immediately to the north of the area of interest (Tambach sediments). Voluminous intra-volcanic waterlain sediments are present in Kamasia, again north of the seismic line (Williams and Chapman, 1986).

The Baringo-Hannington Lake Beds surround the depression in which Lake Baringo lies.

1.3.2 Xenoliths

Mantle xenoliths yield important information on the structure and composition of the lithosphere, on the crust mantle transition, and on the apparent thickness of the lithosphere.

In the area considered, ultramafic xenoliths have been found in Nyanzian terrane (Ito, 1986): they consist of garnet-free harzburgite and dunite, with a small amount of
hornblendite. Their emplacement is related to kimberlites (kimberlite-borne xenoliths) of upper mantle origin, structurally controlled by the development of the Nyanza Rift.

On the eastern flank, Quaternary volcanic fields brought up pyroxenitic and peridotitic xenoliths whose analysis allowed Henjes-Kunst and Altherr (1992) to model the lithospheric structure using geothermobarometric data (Fig. 1.12).

Marsabit lithospheric structure represents a decompressed and cooling lithosphere. Some of the xenoliths collected at Marsabit are better modelled by a structure similar to that in the Chyulu, representing an early stage of evolution when the Moho was deeper. The Lashaine model is representative of a stable cratonic setting. The basement beneath Lashaine is part of the Western Sector of the MOB characterized by pre-Mozambiquan terranes reworked by the Mozambique orogeny. For location of the volcanic fields see Fig. 1.11.
1.3.3 Tectonics

Morpho-tectonic Pattern

An outline of the main patterns (Baker et al., 1972; Williams, 1982; Morley et al., 1992; Smith and Mosley, 1993; Fig. 1.13) defines three different sectors along the Kenya Rift.

A NNE-SSW trend defines the northern sector, where the eastern shoulder is composed of box-fault systems and ramps (Griffiths, 1980), and the western by a series of large E-dipping faults. In the central sector, where the displacements on the bounding faults are greatest, the dominant trend is NW-SE, while in the southern sector it is predominantly NNE-SSW. The main fault scarps vary between 300 and 1600 m in height. The elevation of the plateaux at the rift flanks is between 2000 and 3000 m.

The character of the structural framework of the Kenya Rift is still debated (Morley, 1988; Bosworth, 1989; Strecker and Bosworth, 1991). Information on timing of the development of the various basins is still insufficient to discriminate between:

(a) - an asymmetrical model of basin development (Morley et al., 1992) and (b) - an alternating polarity rift basin model in which the sub-basins are connected by ‘accommodation zones’ (Bosworth et al., 1986; Bosworth, 1987).

The asymmetric graben model can be applied with confidence to the central sector, especially in the Elgeyo-Baringo area, where deep dissection and good stratigraphic and geophysical control allow a number of basins to be distinguished (Smith, 1994).

Recent studies have revealed the close relationship between rift bounding faults, such as the important western bounding Elgeyo fault, and basement structures (Hetzel and Strecker, 1994); a brittle NW trending fault responsible for the abrupt change in rift trend at the Elgeyo escarpment may act as transfer fault. The displacement on the Elgeyo fault is suggested to be between 3 km (Jones, 1988) and 9 km (Morley, 1988).

A listric fault model of continental extension has recently been suggested for the Kerio Basin; up to 5 km of sediment fill and a structure more symmetric than surface data can predict have been proposed (Hendrie et al., 1994).

The fault structures of the Nyanza Rift are radial to the Kenya Dome and probably result from regional stress release during crustal uplift (Fairhead and Walker, 1980).
Figure 1.13: Structural framework of the Kenya Rift zone. (a) Map view. (b) N-S profile of estimated depth to basement along line of rift between X and X' in Fig. 1.13a; top of basement after P-wave velocity model of Henry et al. (1990) and MOE (1987). Vertical exaggeration x8. Structures in basement are as for Fig. 1.9. After Smith and Mosley (1993).
Tectono-magmatic Evolution

The rifting process determined the tectonic evolution with a change from the E-dipping faults striking NNE and NNW from 12-7 Ma (Middle Miocene) to a W-dipping fault system with a NW strike from 5.5-4 Ma (Pliocene) (defining the full graben), to a W-dipping fault system with a NNE and NNW strike after 3.3 Ma (Hackman et al., 1990; Smith and Mosley, 1993).

Overall the zone of rifting has narrowed with time and migrated inwards from the rift margins to the floor of the inner trough. This has been accompanied by an increase in fault density and a decrease in fault length and displacement (Smith, 1994).

'Magmatic evolution for active rifting mode'

Wendlandt and Morgan (1982) have interpreted the magmatic evolution as being consistent with decreasing depth of magma with time, as predicted for magmagenesis associated with an upwelling source region.

Latin et al. (1993) confirmed that the volume and composition of basaltic melt beneath the Kenya Rift over the last 30 Ma supports the model of a plume, decompressing beneath thinned lithosphere: the top of the melt region is within 70 km of the surface beneath the axis and 80 km beneath the flanks of the Rift.

Tatsumi et al. (1991) and Tatsumi and Kimura (1991) recognise two pulses of mantle upwelling following the onset of mantle melting, at 14-15 and 5-6 Ma, and expressed by the basaltic eruptions. Each pulse is followed by expansion of the crust and domal uplift, cracking of the brittle surface, and rifting during the thermal relaxation period (Fig. 1.14).

Smith (1994) suggests that the scale of uplift and the limited amount of lava (compared to the large flood basalts generated by plumes elsewhere in the world - (White and McKenzie, 1989)) is indicative of the presence of a small convective cell with an initial diameter of only 100-150 km.

Lithospheric Structure

An idealised cross-section of the Kenya Rift at lithospheric depth would show important lateral changes in lithospheric structure.

The Kenya Rift is superimposed on the boundary between Archaean and Proterozoic
Figure 1.14: Frequency of K-Ar ages for basalts from the Samburu Hills. Ages of felsic volcanism and tectonic events are also shown. Major basaltic volcanism occurred at 15 and 6 Ma and were followed by felsic volcanism and domal uplift. After Tatsumi et al. (1991).

terranes; there is evidence (White, 1988; Menzies, 1990) that the sub-Phanerozoic and sub-Archaean lithosphere have different chemical and physical properties. This could determine different melt composition and flexural response to imposed loads (Fig. 1.15).

From xenolith analysis fundamental chemical differences have been found in sub-Proterozoic lithosphere (lherzolite) and sub-Archaean lithosphere (harzburgite). Cratonic lithosphere beneath the Archaean crust is depleted in komatiitic elements, while post-Archaean lithosphere is depleted in basaltic melts. Successive overprint, such as upwelling of the asthenosphere and melt addition, could determine local enrichments throughout Proterozoic and Phanerozoic (Menzies, 1990). These differences can be seen in seismic structure: Proterozoic crust is thicker (40-55 km vs 27-40 km) and has a thicker layer of underplated basalt (20-30% vs 5-10%) than Archaean crust (Durrheim and Mooney, 1991). This result is obtained from a world-wide dataset that does not include the East African area.

A cross-section of the MOB by Smith (1994) shows that the depth of the lithosphere-asthenosphere boundary (LAB), where asthenosphere is mantle containing partial melt - (White, 1988; 1993) rises from 250 km underneath the craton into the lower crust beneath
Figure 1.15: Model for Archaean and Proterozoic crustal evolution emphasizing differences in the chemical properties of the uppermost mantle. Archaean crust developed above initially hotter mantle; magmatic underplating, if present, is ultramafic and is seismically indistinguishable from normal mantle. A cold lithospheric keel may also act as a thermal boundary against crustal underplating by basaltic melts from asthenosphere (after Durrheim and Mooney, 1991).
CHAPTER 1. THE KENYA RIFT

the axis of the rift. The relation of this feature to the rising plume will be discussed in Section 1.4.1.

White (1993) has studied the melt production of the Kenya plume, concluding that the LAB, being the base of the mechanical boundary layer, has risen to a minimum of about 100 km.

1.4 Geodynamics of Riffs and Plate Kinematics

In plate tectonic terms, rifting is considered to be the first stage of continental splitting, and could lead to the formation of a new ocean. The spreading plates are part of the lithosphere, the cool and rigid outer shell of the earth, overlying hot and less rigid asthenosphere (White, 1993). Many authors (Section 1.5.2) on Kenya extend the definition of asthenosphere to include any portions of the upper mantle containing partial melt (Section 1.3.3).

In the following section the spatial and temporal evolution of the lithosphere-asthenosphere system underneath the Kenya Rift is summarised, the surface expression of the rifting stages reconstructed, and finally there is a discussion on how the plate movements relate to the asthenospheric flow.

1.4.1 Geodynamic Evolution of the Kenya Rift

The first sign of a thermal perturbation in the mantle underneath Kenya occurs in the form of diatremes at about 45 Ma. The spatial-temporal-compositional variation of the eruptives suggests a magma source shallowing with time.

Mohr (1992) describes the evolution of the Kenya Rift, based on isotopic data from rift lavas, in the following way (Fig. 1.16): the impingement on the base of the lithosphere of a hot mantle plume head facilitates uplift and generates flood basalts; the lateral stresses and uplift cause rifting of the brittle crust; asthenospheric material intrudes the lithospheric mantle. He envisages the wedge penetration to at least the Moho base occurring episodically (Section 1.3.3).

White and McKenzie (1989) relate the quantity of basaltic lava generated during rifting
Figure 1.16: Simplified cartoon of basalt sources under a continental rift. EM=enriched lithospheric mantle; PM= ocean island type plume mantle; LAB=lithosphere-asthenosphere boundary. 

a: Lithospheric cross-section, prior to thermal perturbation from arriving plume head. 
b: At end of flood basalt episode, lava filled subsided basin is underlain by numerous sill-like intrusions that have thickened the rift crust. Lower part of lithospheric mantle has been both convected away and converted into 'active' lithosphere. 
c: Plume mantle head penetrates into the crust; profuse silicic magma production above ponded mantle melts (modified from Mohr, 1992).
to the amount of stretching and the temperature of the mantle underneath the extended lithosphere. In the northern part of the Kenya Rift the extension and the amount of basaltic lavas is greater than in the south; it has been suggested that the mantle temperature is higher under the apex of the Kenya Dome than under the northern part beneath Lake Turkana (Achauer et al., 1992).

1.4.2 Rift Development in Kenya

Mesozoic rifting and early Tertiary rifting will not be considered here; according to Bosworth et al. (1992) the main pre-Cenozoic rift in Kenya, the Anza Rift (Fig. 1.9), was not significantly reactivated during development of the presently active East African Rift System. Archaean and Proterozoic basement features (ductile shears and mantle sutures) however do influence Cenozoic structures (Smith and Mosley, 1993; Smith, 1994; Hetzel and Strecker, 1994).

Two schools of thought have developed. One suggests that the East African System has formed during rift-normal extension (Baker et al., 1972; Morley, 1988; Bosworth et al., 1992; Smith and Mosley, 1993), and the other invokes oblique extension parallel to the NW-SE striking Aswa shear system (Chorowicz, 1983; Rosendahl, 1987).

In general, the Neogene rifts align N-S, upon a basement with strong NW-SE trends and less pronounced N-S anisotropy (Bosworth et al., 1992); different elements of the inherited tectonic mosaic came into play during successive Cenozoic tectonic episodes.

**Stages of Rifting evolution (Fig. 1.17)**

A schematic description of the Kenya Rift's progressive development is presented from Chorowicz et al. (1987) and Hackman et al. (1990); for a description of the volcanic and tectonic evolution in the study area see Section 1.3.3:

- Mid-Cenozoic (Eocene ?): a monoclinal flexure develops on the western flank of the Kenya Dome.
- Oligocene: flood basalts in the northern rift predate extensive extrusions in the Kenya and Nyanza Rift to the south, suggesting a southward extension of the activity from Ethiopia where it started in Eocene time.
- Late Oligocene-Early Miocene (up to 20 Ma): on the sub-Miocene surface a pre-rift
CHAPTER 1. THE KENYA RIFT

Figure 1.17: top: Successive block diagrams of the lithosphere corresponding to the different stages of rift evolution (after Chorowicz et al., 1987). Bottom: Diagram showing the tectonic and volcanic evolution of the Kenya Rift. The generalised spatial relation of the volcanism to the rift axis is shown (after Keller et al., 1991).
depression is formed by downwarping and minor faulting, corresponding to an upwarping of the Kenya Dome near the Kenya-Uganda border. Horizontal slip movement with a divergent component occurs along fractures inherited from older tectonic phases; the first movement is NW-SE (not perpendicular to previous structures). Tension gashes evolve into transcurrent faults. Very thick successions of tholeiitic volcanics and sediments accumulate; dykes and plugs are emplaced.

- Miocene (from 20 to 7 Ma) the volcanic and tectonic activity continues in the Nyanza Rift (Jones and Lippard, 1979). True half grabens are developed in the Kenya Rift after 12 Ma (Middle Miocene) with a broad uplift of about 300 m and a culmination of the tectonism at about 7 Ma (Elgeyo fault). The tectonism is represented by faulting with some strike slip component and the segmentation of the rift into basins separated by basement highs. The extension strikes NW-SE. Trachytic and basaltic volcanism is of alkaline and peralkaline chemistry.

- Late Miocene-Pliocene: the faults become mainly normal, the extension horizontal and often striking perpendicular to the main scarps. Tension gashes lie parallel to the rift axis. Important subsidence and uplift of the shoulders occur (up to 1400 m). The faulting tends to migrate to the centre of the rift (Keller et al., 1991). Widespread basaltic volcanism marks this phase. The resulting basalt platforms are then tilted and covered by trachytic shield complexes. The faulting close to the rift axis allowed the deposition of sediments in the half graben structures in the Elgeyo-Baringo sector. Apart from local eruptions the Nyanza Rift seems inactive since 7 Ma ago.

- Pleistocene-Recent: the Pliocene trends continue. The rift subdivision into basins have almost disappeared. Alkaline volcanism and dyke-swarms are currently intruding to high levels in the crust (Smith and Mosley, 1993).

Plate Kinematics

Pollitz (1991) synthesized information from hotspot tracks, geophysical constraints and the record of the present plate motions using the relative motion model of DeMets et al., 1990 and unpublished results from DeMets. The absolute motion of the African plate changed between 8 and 4 Ma, with a counter-clockwise rotation. One of the consequences of this change could be the initiation of the East African Rift. The model involves mantle
material flowing out horizontally in a radial pattern from the Afar hotspot, which could account for the south-westerly direction of motion of the African Plate.

1.5 Geophysical Studies of the Kenya Rift

Geophysical data acquired in the Kenya Rift in the past 30 years have helped to elucidate questions concerning its development and present structure; data acquired during the KRISP 90 project will be discussed in Chapter 2.

Heat flow, seismicity and neotectonic measurements demonstrate the Kenya Rift is an active feature. Gravity and seismic data are consistent with an upwelling of the asthenosphere beneath the axis of the rift. Teleseismic results suggest a degree of melting, and magnetotelluric and geomagnetic deep sounding data indicate anomalously high conductivity in the upper mantle. The presence of partial melt resulting from rising geotherms could also explain the high conductivities in the upper mantle.

1.5.1 Gravity Surveys in Kenya

Gravity data have been collected in this area over several years by workers from the University of Leicester (Khan and Mansfield, 1971; Swain and Khan, 1977; Khan and Swain, 1978) and elsewhere (for full reference list see Swain and Khan, 1977). The Bouguer Gravity map of Kenya published by the Survey of Kenya in 1982 is shown in Fig. 1.18.

The regional field shows that the East African Plateau and the Kenya Dome are characterised by negative Bouger values (up to 250 mGal over a circle of 1000 km radius), suggesting a thick (35-40 km) continental crust, lithosphere uplift along the axis of the rift and its replacement by low-density asthenosphere (Fig. 1.19).

A prominent feature in the Kenya Rift Valley is a discontinuous positive gravity lineation of wavelength 50-80 km and of amplitude 30-35 mGal, which follows the spatial distribution of the seismic, volcanic and geothermal activity of the Rift Valley floor (Searle, 1970; Baker and Wohlenberg, 1971; Baker et al., 1972; Fairhead, 1976).

A large negative anomaly dominates the Nyanza Rift; Fairhead and Walker (1980) interpreted this in terms of a granite intrusion underlying the rift. Smaller wavelength
Figure 1.18: Bouguer gravity anomaly map of Kenya, after Khan and Swain (1978). Contour intervals=10 mGals.
Figure 1.19: Residual Bouguer gravity map of Kenya, after Khan and Swain (1978). Contour intervals=10 mGals.
negative anomalies are caused by sedimentary basins.

The broad negative anomaly

Over the Kenya Dome the negative anomaly is of about 170-180 mGal. Searle (1970) modelled this anomaly with low-density material at the base of the lithosphere, rising into the lower crust beneath the rift. Low density rift volcanics at the surface are added to account for the negative anomaly. Baker and Wohlenberg (1971) suggest that the negative anomaly could be explained by 3000 m thick low density volcanic products in the rift floor (Fig. 1.20b).

With the assumption that the Dome is isostatically compensated, Fairhead’s model (1976) included a broad low density structure extending more than 100 km to the west of the Rift and replacing the normal lithospheric mantle (1976, Fig. 1.20a). Raised geotherms in the lithosphere provide low-density material that gives the buoyancy forces supporting the uplifted topography. To explain the negative anomaly a density contrast of $0.05 \text{ g/cm}^3$ is required; the velocity decrease necessary to explain the delay in teleseismic arrival is about 10%. Because the value of $\delta V_p/\delta \rho$ is larger than for the Nafe-Drake relationship Fairhead (1976) suggested for the first time the presence of partial melt in the upper mantle.

Khan and Swain (1978) represented the regional field by a seventh order trend surface with a minimum at the culmination of the Kenya Dome. The surface correlates inversely with the topography and rises to the north of the Rift Valley. It is suggested that the anomalous mantle rises and widens south of Turkana.

In the 1960s the Kenya Dome was considered a region of domal uplift, but opinions diverge on the timing and stages of uplifting. Integration of seismic and gravity studies with apatite fission-track studies during the 1980s recognised the Kenya Dome to be a topographic feature resulting from the combined effect of Cenozoic volcanic accumulation and domal uplift (King, 1978; Bechtel et al., 1987; Wagner et al., 1992).

Bechtel et al. (1987) showed that the longest wavelength component of the topography is locally compensated; this represent a substantial portion of the Dome relief at the eastern edge of the East African Plateau. The short wavelength topography is mainly due to volcanic accumulation or surface loading.
Figure 1.20: Gravity models for the Kenya Rift. A: model by Fairhead (1976) at 1.8°S showing lithospheric thinning beneath the rift valley, as well as beneath the MOB and ANC. B: model by Baker and Wohlenberg (1971) near the Equator showing thinning localised beneath the rift and a low-density sheet at the surface. C: model by Nyblade and Pollack (1992) at 0.5°N showing the 'rift' and 'suture' components. D: model by Swain (1992) at 0.5°N showing a broad deep zone of slightly increased density to explain the Kenya Rift axial gravity high.
CHAPTER 1. THE KENYA RIFT

More recently Ebinger et al. (1989) analysed a data set covering the whole East African Plateau and stated that wavelengths longer than 1000 km are overcompensated. A mechanism of dynamic uplift and associated heating of the thermal lithosphere above a convecting region in the asthenosphere was used to satisfy the mechanism of isostatic compensation supporting the topography of the uplifted regions.

Nyblade and Pollack (1990) emphasize the importance of the boundary between the MOB and the ANC in contributing to the broad negative anomaly (Fig. 1.20c). The overall anomaly across the rift is the result of a wide asthenospheric intrusion ('rift' signature resulting from low density material beneath the axis) and a 'suture' signature between the craton and the MOB producing a negative anomaly over the boundary and a positive anomaly over the younger tectonic block. Their low-density material is therefore confined to the width of the rift valley suggesting that at the earliest stage of rifting no large-scale perturbation is present at the base of the lithosphere.

The axial gravity high

The axial positive anomaly is too large to be explained by dense volcanic rocks extruded at the surface. Searle (1970) suggested the presence of a dense intrusion beneath the rift, approximately 16-20 km wide, coming to within 2-3 km from the surface.

A similar solution was presented by Baker and Wohlenberg (1971) but in their model the low density asthenospheric material is only 10 km wide in the upper crust (Fig. 1.20b).

A 10 km wide single dyke or a dense infill body approximately 2.5 km thick is modelled by Fairhead (1976; Fig. 1.20b) and an 8-14 km wide high density intrusion in the lower crust is modelled by Nyblade and Pollack (1990: Fig. 1.20c).

Recently Swain (1992) reinterpreted an E-W profile across the Lake Baringo. In his model a basic dyke swarm intrusion in the basement gneiss increases the basement density (from 2.70 to 2.76 g/cm$^3$) over a 40 km span across the graben down to 22 km depth (Fig. 1.20d). The depth is constrained by a high velocity reflector identified during KRISP 85 (Henry et al., 1990) and interpreted as a basic sill, although its width is less well constrained (Section 1.5.3).
CHAPTER 1. THE KENYA RIFT

Figure 1.21: Crustal velocity-depth functions from earthquake studies: AFRIC (S): for shield areas from Rayleigh wave dispersion and body wave arrival times (Gumper and Pomeroy, 1970); AAE-NAI: from surface wave dispersion along the path AAE-NAI; crustal velocities and thicknesses of the first two layers are adopted from AFRIC (Long et al., 1972); Kaptagat: from body wave arrivals observed on the western flank of the rift (Maguire and Long, 1976); AFRIC (P): derived from AFRIC (S) (Gumper and Pomeroy, 1970); NAI Conversions: from P to S conversion at the Moho beneath NAI, AFRIC values for crustal velocities and thicknesses of first two layers (Herbert and Langston, 1985).

1.5.2 The Deep Structure of the Kenya Rift from Teleseismic Studies

Data from the World-Wide Standardized Seismograph Network (WWSSN) provided almost the first evidence for a ‘gap’ in the lithosphere beneath the Rift (Fig. 1.21). Gumper and Pomeroy (1970) outlined a model for Africa’s shield areas where a 37 km crust can be subdivided into three layers with apparent velocities of 5.8, 6.2, 6.7 km/s above an upper mantle of 8.0 km/s velocity. $S_n$ propagates poorly across the rift north of the equator. Consequently, the presence of an anomalous upper mantle down to depths of 400 km has been suggested by Nolet and Mueller (1982). Rayleigh wave dispersion studies indicate that the shear wave velocity under the rift is 10% less than under the shield areas.

Analysis of long period body waves (Bonjer et al., 1970) determined:

- normal crustal structure under the NAI station (Nairobi), involving a 40-42 km thick two-layer crust, with an intermediate high-speed layer in the lower crust;
- the presence of low-velocity mantle material in the upper mantle under the northern part of the rift valley being thicker than under the southern part.

Mueller and Bonjer (1973) integrating the long period body wave with surface wave analysis modelled a 48 km thick two layer crust with a 20 km depth upper-lower crust boundary, and detected a low velocity zone for S-waves between 80 and 210 km depth.

An analysis of first arrivals from local and regional earthquakes in the Kaptagat area indicated that beneath the western flank of the Kenya Rift the crustal thickness is about 44 km and thinning of the crust appears limited to the rift valley, with an anomalously low P and S velocity zone beneath the rift axis (the P-wave velocity in the upper crust, lower crust and upper mantle beneath the western flank being respectively 5.8, 6.5 and 7.9 km/s (Maguire and Long, 1976)).

Teleseismic data has been used to identify an anomalously low velocity body (LVZ) in the upper mantle, modelled by a symmetrical upwarp in the lithosphere/asthenosphere boundary and possibly penetration into the crust from 200 km depth. This body thins both to the north and to the south of Kaptagat (Long and Backhouse, 1976). A 400 km linear array orientated SE-NW through the central part of the rift supported the existence of a LVZ centred under the rift (Savage and Long, 1985; Long, 1987).

Dahlheim et al. (1989) and Davis (1991) studied data from a 600 km long EW array at 1°S deployed during KRISP 85. They postulated the presence of a LVZ down to 200 km depth, dipping to the west. Combined analysis of teleseismic studies and gravity modelling required a 5% partial melt in the upper mantle to explain the 4% slow velocity perturbation in relation to the required density distribution to explain the gravity anomaly (Section 1.5.1).

An estimate of Q in the asthenosphere on the KRISP 85 data show that the LVZ and low Q values coincide beneath the Kenya Rift. Low densities associated with these anomalies could support regional uplift. A 3% melt fraction is hypothesized for the upper mantle under the Kenya Rift (Halderman and Davis, 1991).

A more detailed picture results from the 2-D array study undertaken during KRISP 85 in the central rift zone (Green et al., 1991; Green and Meyer, 1992; Achaner, 1992; Fig.1. 22).
Figure 1.22: KRISP 85 teleseismic model (after Green and Meyer, 1992), showing a cross-section (after Bosworth, 1987), travel-time residual profile, and Bouguer gravity profile at 0.5°S.
The inversion of these data shows evidence of high velocities in the lower crust beneath the rift axis (10-35 km depth), whereas in the upper mantle (65-105 km depth) a N-S linear LVZ below the rift suggests an asthenosphere upwarp. The discrepancy in magnitude between the decrease in velocity and the density contrast can not be completely explained with a temperature anomaly under the rift. 5% melt is required, within a dry peridotitic upper mantle. The asymmetry of the rift is explained via the rejuvenation of a zone of weakness along the craton-mobile belt boundary. An active mechanism of rifting is supported by these recent results (Achauer, 1992). However, the crustal thinning and the asymmetric shape of the sedimentary basins suggest that passive extensional forces are also acting within the lithosphere (Green et al., 1991).

1.5.3 Seismic Wide-angle and Reflection Surveys

**Wide-angle Surveys**

The first seismic refraction experiment in the Kenya Rift was carried out from Lake Turkana to Lake Bogoria, in 1968 (Fig. 1.23; KRISP 68: Griffiths et al., 1971).

A crustal layer of $V_p$ around 6.4 km/s at a depth of about 3 km was related to a 'shallow axial intrusion' of basic rocks producing the gravity high. Asthenospheric penetration of the lithosphere to within 15-20 km of the surface was identified by P-wave velocities of about 7.5 km/s.

The Kaptagat array of Maguire and Long (1976) confirmed the existence of anomalous material beneath the rift axis with a velocity between 7.1 and 7.5 km/s, possibly underlying the 6.4 km/s zone detected by Griffiths et al. (1971).

KRISP 75 involved a 50 km E-W profile across the western half of the Rift Valley, from the base of the Elgeyo escarpment to Lake Baringo. The aim was to map the structure of the 6.4 km/s layer detected in 1968, and to constrain the gravity data. The velocity section is shown in Fig. 1.24a. There is no direct evidence of the axial intrusion but the presence of high density material in the upper crust is implied by the combined interpretation of the gravity and seismic data (Swain, 1979; Swain et al., 1981).

Two seismic refraction lines were deployed for KRISP 85, along the rift axis from Lake Baringo to Lake Magadi (with stations between Lake Bogoria and Suswa) and across the
Figure 1.23: Map showing locations of seismic refraction lines and temporary local networks; KRISP 90 is also included.
Figure 1.24: a: Interpretation of KRISP 75 first arrival data and geological section. Velocity section: mean velocities in km/s. The zig-zag lines do not represent actual structure (after Swain et al., 1981). b: Interpretation of KRISP 85 north-south profile. Shallow velocity model. Velocities in km/s (after Henry, 1987).
axis of the rift in its central part (from 36°E to 36°3'E) (Henry, 1987; KRISP Working Group, 1987; Khan et al., 1989; Henry et al., 1990; Keller et al., 1991). The results are shown in Fig. 1.24b. The depth of the basement is very variable (1.5-1.6 km), but Precambrian rocks of velocity c.6.05 km/s underlie the Rift. There is no evidence of the 'massive axial intrusion'. Nevertheless Swain (1992) suggested that some relatively dense material should exist to explain the positive isostatic anomalies in a region with several kilometres of low density volcanics (Fig. 1.20d).

It is suggested that beneath the rift there is a pervasive high-velocity intrusion which can explain the axial gravity anomaly. The brittle-ductile transition (Young et al., 1991) may coincide with a change in \( V_p \) from 6.2 to 6.45 km/s beneath the rift. A high velocity layer at a depth of about 25 km may indicate the presence of a sill-like basic intrusion. The crustal thickness under the central part of the rift is about 35 km. The \( V_p \) velocity in the upper mantle, from unreversed data, seems to be less than 7.7 km/s at a depth of 34 km.

Near Vertical Surveys

Five seismic reflection lines were undertaken in 1989 in the Kerio Valley in the Kenya Rift. Interpretation depicted high variability in the shallow subsurface along the Valley. Two episodes of faulting are seen in the development of the Elgeyo fault. The results suggest the presence of a full graben against the western margin of the rift (Pope, 1992). The development of the basin might be related to an Eocene-Oligocene rifting episode, and be part of the Anza Rift system (Hendrie et al., 1994).

1.5.4 Seismicity of the Kenya Rift

The Rift Valley is an area of concentrated seismicity of low magnitude. Microearthquake studies show good correlation of the seismicity with surface faults, hot springs and geysers, and more generally with geothermal and volcanic regions and the large scale structural features of the main rift and the E-W trending Nyanza trough (Pointing, 1985; Pointing et al., 1985; Maguire et al., 1988; Young, 1989) (Fig. 1.25).

Combined studies of the geotherm, from heat flow data, and from earthquake focal
Figure 1.25: Seismicity distribution in Kenya from temporary local networks (after Tongue et al., 1992).
depth in the Eastern Rift suggest a depth of $10 \pm 2$ km for the brittle-ductile transition (Fadale and Ranalli, 1996; Young et al., 1991). Aseismic creep may occur at relatively shallow depths where strain is accommodated along low-angle faults, or by the intrusion of magma bodies.

Analysis of source mechanisms (Shudofsky, 1985), shows predominantly normal faulting with the least compressive stress direction normal to the rift axis.

There is no agreement on the direction of the palaeostress. Strecker (1991) suggests that kinematic and stratigraphic evidence shows an early extension striking E-W. A rotation of the extension direction to NW-SE during the Pleistocene (in the last 0.5 Ma), radially away from Afar plate junction is derived from regional fault kinematic and borehole break-out data (Strecker et al., 1990; Strecker and Bosworth, 1991; Bosworth et al., 1992) and analysis of shear wave data from Lake Baringo (Tongue et al., 1994). Chorowicz et al. (1987) prefer a pre-rift NW-SE strike, related to horizontal opening movements, deduced from observations of fault striations. The present state of regional deformation, from earthquake focal mechanism studies, is compatible with a NW-SE to WNW-ESE extension.

1.5.5 Heat Flow Measurements in Kenya

Heat flow data have been published by Morgan (1982, 1983) and Crane and O'Connell (1983); Nyblade and Pollack (1990) and Nyblade et al. (1990) have summarised some of the previous data sets. Their conclusions are consistent with a model of anomalous thermally and tectonically active lithosphere under the Kenya Dome and its associated volcanic province, and localised thermal upwelling in the mantle. On a regional scale the Archaean craton has lower heat flow values than the Proterozoic mobile belt, with values increasing away from the centre of the craton.

Revised and integrated data have been published by Wheeldon et al. (1994). The following pattern can be identified:

- In the rift floor values range from 50 to 100 mWm$^{-2}$; the highest values are spatially associated with fluid circulation (considerable hydrothermal activity is observed within the rift), Quaternary volcanism and faulting.
- on the western flank values of 40-50 mWm$^{-2}$ are measured; higher values are associated with the Nyanza Rift.

- on the eastern flank values of 40-60 mWm$^{-2}$ are obtained, with the lower values associated with ground water infiltration. Their conclusions do not directly support 'active' rifting, as the thermal anomaly does not seem to have reached the surface. However, the lack of a broad thermal anomaly accounts for the youth of the rifting process, and excludes models of asymmetric rifting at a lithospheric scale as proposed by Bosworth (1987).

1.5.6 Magnetotelluric and Geomagnetic Studies

High values of conductivity measured from a magnetotelluric traverse across the Rift Valley in the vicinity of the equator were explained by magmatically convected heat locally distributed by fluid circulation at depths of less than 20 km.

High conductivities 100 km east of the rift below 40 km probably result from partial melting (Rooney and Hutton, 1977). It is estimated that less than 5% melt in a mantle composed predominantly of olivine would be sufficient to produce the observed conductivity values.

Geomagnetic variation studies also suggest that the high conductivity in the lower crust and upper mantle results from a zone of melting within the subcrustal lithosphere. Another conductive layer in the upper crust beneath the rift axis may also be due to partial melt or electrolytic conduction (Banks and Ottey, 1974; Banks and Beamish, 1979).

In 1987 an aeromagnetic survey was undertaken over the entire rift. The aim was to detect anomalies of the type found at Olkaria (SW of Lake Naivasha) associated with a geothermal field. One target anomaly was discovered on the rift margin to the SE of Lake Baringo (Compagnie Générale de Géophysique, 1987).

1.6 Summary

Many questions remain to be answered concerning the Kenya Rift and continental rifts in general. In the last 30 years the amount of information and the integration of different approaches to the study of the Kenya Rift have helped to focus attention on the major
problems.

Mathematical simulations of continental processes have eliminated some of the unrealistic possibilities; combined models have reduced the contrast between end-members in geodynamic models.

Detailed geological mapping, petrological studies and rock dating define a generally accepted sequence of Rift evolution. Nevertheless, even for the crustal structure there are still obscure points, especially concerning the depth and shape of the rift related basins.

A big effort has been put into the collection of geophysical data: the combined interpretation of heat-flow, magnetic, gravity and seismic data outlines the lithospheric structure of the rift. The term 'Kenya Dome' is now used to describe a structural feature rather than a domal uplift; the 'axial intrusion' has become a diffuse area of dyke swarm. The upper mantle updoming under the rift axis is undisputed, although there is still disagreement on its detailed shape and its role in the rifting process.

The integration of data from different geophysical methods is the main characteristic of KRISP 90, and marked its successful contribution to the better understanding of the Kenya Rift. Interpretation of data from the cross-rift line of KRISP 90 is the basis of this thesis. Subsequent chapters lead through the processing and interpretation of the geophysical data to produce the resulting geological models. These in turn will be discussed in relation to the principal uncertainties/unknowns discussed here.

1.6.1 Thesis Outline

Fig. 1.26 represents the steps regarding data acquisition during KRISP 90, data processing and display following the experiment, and anticipates the main steps of the modelling sequence followed during the present study.

Chapter 2 is a description of the KRISP 90 experiment, with a summary of the results from the concomitant surveys relevant to the cross-rift seismic line. Details of the cross-rift line field work and the preliminary data processing will be presented. The Chapter is concluded with a schematic geological section along the cross-rift line.

In Chapter 3 modelling of the P-wave field arising from the cross-rift line D is described in detail; 1-D velocity-depth curves, delay-time analyses and 2-D forward modelling ray-
CHAPTER 1. THE KENYA RIFT

Figure 1.26: Flow chart showing the seismic data analysis procedure.
tracing have been undertaken as part of the kinematic modelling. The final velocity model is interpreted geologically and integrated with a gravity anomaly model constructed along the line.

Chapter 4 reports the kinematic modelling of the S-wave field for the cross-rift line D and the flank rift line E by 2-D ray-tracing of the model derived from the P-wave interpretation, with modification of the velocity field to fit the S travel times. The Poisson’s ratio distribution is also discussed.

In Chapter 5 amplitude modelling of the main P and S phases is presented. Reflectivity and finite difference modelling of the amplitudes are mainly used to test the validity of the models generated by travel time forward modelling.

Chapter 6 describes the result of a detailed study of a reflected phase from the upper mantle in relation to the diving waves under the Moho discontinuity. Petrological implications in term of upper mantle structure and temperature are discussed.

In Chapter 7 the resolution and uncertainty of the forward model is checked upon and an alternative model is created via travel time inversion (tomography) of the P-wave field. A further control is provided by the calculation of the gravity anomaly arising from the forward and inverted models.

Chapter 8 discusses the main conclusions and assesses the implications of the present study.

Rather than developing any particular geophysical technique, this work involves the use of established techniques to interpret a good quality seismic data set and to discuss its implications. The comparison and integration of different techniques allows the maximum information possible to be extracted from the available data set.

Station spacing and phase wavelength constrain the model resolution. Subjective identification of the various phases can lead to different models from the same data set. Taking these limitations into account, forward modelling ray-tracing can include a priori information in a model which is simple to construct and to modify. On the other hand, travel time fit and error are difficult to estimate, and the overall procedure is long and tedious.

Thus, forward modelling is not really satisfactory on its own, and inversion of the data should also be performed: linearized travel time inversion allows an assessment of error,
resolution and uncertainty in the model. Model parameterization is generally similar to that used in the forward modelling techniques, and together with the initial phase identification influences the quality of the final result. In the present study using the Zelt and Smith (1992) inversion scheme the ray-trace algorithm used in the starting phase of the inversion prefers smooth layer boundaries to complex structures which in fact could be resolved by the data set and phase wavelength. However, the inversion scheme quickly reaches the maximum possible convergence between observed and calculated data.

Useful information was also provided by the amplitude study, which helped to constrain model parameters such as velocity gradients and contrasts across interfaces. The use of supercomputers permitted sophisticated amplitude modelling of the whole wave field. The disadvantages in the methods used are the long computer time necessary, the impossibility of modelling lateral variations with reflectivity, and in the case of the finite difference modelling the generation of phases which are difficult to identify.

The analysis of the S-wave field was important as it increased the knowledge of physical properties and petrological composition. This study has been limited by the availability of data representing only the vertical component of the seismic wave field.

The integration of seismic and gravity data is an effective tool in helping to interpret the model geologically and isostatically, although it has been shown that the relation between seismic velocity and density is largely non-unique.

Extensive modelling has been performed on some particular phases, such as multiple reflections from the Moho and a reflected phase from within the upper mantle. In the latter case, an original combination of processing techniques to enhance the data quality and interpret the result allowed some suggestions to be outlined on the upper mantle structure which should be further investigated via deep seismic reflection profiling.
CHAPTER 1. THE KENYA RIFT

Photo 1: The Nandi escarpment; view from the SW.
Chapter 2

KRISP 90

2.1 Introduction

The 1985 Kenya Rift International Seismic Project, whose important scientific results are described in Section 1.5.3, was designed as a preliminary feasibility study for a larger experiment which followed a few years later, KRISP 90.

KRISP 90 involved the operation of a teleseismic array from October 1989 to April 1990, of a microseismic network from January to March 1990, and included three major (five segments) seismic refraction/wide angle reflection profiles recorded in January and February 1990 (Fig. 2.1). The seismic investigations were accompanied by petrological, geological, gravity and geothermal field surveys.

An overview of the experiments and a summary of the main results will be presented here. Emphasis will be given to the data which is relevant to the interpretation of the cross-rift line, the subject of the present work; the results will be discussed in depth in Chapter 8, in the light of the present study. The cross-rift line D will be described here, in more detail, from the profile deployment to the data processing.

2.2 KRISP 90 Objectives

The main purpose of KRISP 90 was to study the crustal and uppermost mantle structure, composition and physical state beneath the Kenya Rift down to depths of 100-150 km.
Figure 2.1: KRISP 90 location map showing the seismic refraction/wide angle reflection lines and the configuration of the teleseismic network (trapezoidal area) (after Prodehl et al., 1994b).
The specific scientific objectives of KRISP 90 were:
- to study details of the upward perturbation of the lithosphere-asthenosphere boundary predicted by previous geophysical surveys;
- to examine the relationship between mantle updoming and the asymmetric development of sedimentary basin structure within the Rift;
- to locate possible zones of partial melt in the crust and mantle;
- to analyse the local seismicity to provide an estimate of seismic risk, and as an aid to understanding the local tectonics,
- to study shear wave splitting as an indication of extensive dilatancy anisotropy and its associated stress regime, and
- to answer fundamental questions concerning the mode and mechanism of continental rifting (Section 1.2.1).

The axial line observations (segments A, B and C) were designed to obtain a detailed crustal model of the rift proper and structural information from as deep within the crust as possible; the northernmost segment (Line A) along Lake Turkana was deployed as a high resolution study of the upper crust.

The cross rift line D aimed to depict the west and east flank of the rift as well as the transition between rift and flanks; this profile is described further in Section 2.5.

The flank line E was planned to obtain data for the construction of a model for an area tectonically undisturbed by the Tertiary rifting process. The chosen location to the north-east of the Tertiary Rift paralleled the nearby Mesozoic Anza Rift.

The teleseismic delay-time study targeted the structure of the subcrustal lithosphere down to 150 km depth.

The microseismic network located around Lake Baringo was designed to continue to the north the seismicity study undertaken during KRISP 85 around Lake Bogoria (Young et al., 1991), to examine the 3-D structure beneath Lake Baringo where an upper crustal magmatic intrusion had been predicted (Karson and Curtis, 1989), and to study the regional stress distribution.
2.3 The Refraction Experiment, the Teleseismic Experiment, the Microseismic Network

A technical review of the KRISP 90 experiments can be found in Prodehl et al. (1994b). In the following section the most important findings from each experiment will be outlined.

The Refraction Experiment

Only the axial and flank line results are described here. The cross-rift line results are presented in the following chapters and in Maguire et al. (1994) paper enclosed as Appendix F.

The interpretation of the axial line revealed a thinning of the crust from 35 km in the south beneath Lake Naivasha to 20 km in the north beneath Lake Turkana (Fig. 2.2a after Mechie et al., 1994b).

The model resulting from the flank line shows a crust thinning from 35 km in the south-east to 28 km beneath Lake Turkana (Fig. 2.2b after Prodehl et al., 1994a). Comparison of the axial line to the flank line model led to the conclusion that the thinning along the Kenya Rift was caused by the Tertiary rifting event, and the amount of thinning increases further away from the apex of the Kenya Dome. The surficial expression of the rift shoulders is wider toward the north, ranging from 60 km at Lake Naivasha to c.180 km in the vicinity of Lake Turkana, again suggesting this splay is somehow controlled by the location of the Kenya Dome.

Mechie et al. (1994b) discussed the axial line interpretation as follows:

Upper crustal velocity variations were attributed to intrusion of igneous material increasing the velocity from 6.1-6.2 to 6.3 km/s beneath the rift itself.

Lower crustal velocities average at about 6.4-6.5 km/s, with no significant variations along the rift axis; in other words, they do not seem to have been significantly affected by the rifting process, being similar to those seen on the flank line. The decrease in velocity resulting from the increased temperature beneath the rift is counteracted by the increase due to mafic dyke intrusion under the axis.

A lower crustal layer with a velocity of 6.8 km/s thins dramatically from south to north, where the amount of crustal stretching seems to have been greater (35-40 km in
Figure 2.2: a: 2-D ray-trace model for the KRISP 90 axial line (segments A, B and C). Velocities are accurate to the nearest 0.05 km/s. The structure north of LTN has been taken from the interpretation of line A by Gajewski et al. (1994). b: 2-D ray-trace model for the KRISP 90 flank line E. Vertical exaggeration 5x. Depths are related to sea level. Dashed line is a change of velocity gradient (after Prodehl et al., 1994a).
the north versus 10 km in the south (Baker and Wohlenberg, 1971; Morley et al., 1992)). Because the upper mid-crustal layer does not thin as much as the lower crustal layer, the latter might be formed from ancient lower crust intruded by mafic mantle material and stretched in a ductile manner.

The seismic velocities appear anomalously low in the uppermost mantle along the whole axial line with values of 7.5-7.7 km/s. Combined modelling with the teleseismic data and the Bouguer anomaly data suggested the presence of a few percent of melt (5-6%) below 40-50 km (Green et al., 1991). At this level in the uppermost mantle, magma rising from below might be trapped above about 40 km (Mechie et al., 1994a). More details about the upper mantle structure along the axial line will be given in Chapter 6, in relation to the interpretation of upper mantle phases from the cross-rift line D.

The most significant features along the flank line (Prodehl et al., 1994a) in the upper crust are sill-like layers near CHF (with a velocity of 6.45 km/s compared to the average 6.0-6.2 km/s), interpreted as Precambrian ‘ophiolitic’ units and intrusions of anorthositic gabbro. The increase in mid-crustal velocity from 6.4 km/s at Lake Turkana to 6.55 km/s to the SE, could be related, as for the axial line, to the rift stretching and extensional effect being less pronounced in the SE away from the rift axis.

The complex character of the Moho reflection beneath the flank line has been interpreted as due to a laminated Moho (Prodehl et al., 1994a). It will be described in more detail in Chapter 4, which includes the interpretation of the S-wave field from line E, the flank profile.

The Teleseismic Experiment

A 3-D long-wavelength image between 10 and 38 km depth from teleseismic data shows that no significant wholesale intrusion of mafic material is present in the crust (Fig. 2.3-2.4 after Achauer et al., 1994).

Beneath the southern Kenya Rift a low velocity body exists extending from the Moho down to at least 165 km depth. In the depth range 35-65 km it seems to reflect the surface extent of the Kenya Rift, with a prolongation into the Nyanza Rift. The broadening at depth occurs below 125 km (Slack and Davis, 1994; Fig. 2.3-2.4 after Achauer et al., 1994).
Figure 2.3: a: Top view of the schematic map showing the 'rift' model used to calculate the theoretical residuals to test several structural elements. Stippled area=north-south oriented rift; TS=NW-SE-oriented narrow 'shear zone', with lower velocities relative to the lithospheric upper mantle; D=small diapirs within the low-velocity upper mantle beneath the rift. b: E-W cross-section through the model shown in (a). Numbers indicate the velocities in km/s.
Figure 2.4: Inversion result for KRISP 90 teleseismic data. Velocity perturbations per layer with respect to the starting model are shown. Layer thickness, block size and initial velocity model are similar to the real data case. Major rift faults as heavy lines; elements of the ANC-MOB suture as dashed lines.
The contact between high and low velocity portions of upper mantle appears to be surprisingly sharp at mantle depths where the heat diffusion should produce broader velocity anomalies. It is suggested that this might be related to the young age of the anomaly and will be discussed further in Section 6.6 and 6.7.

From the teleseismic results, there do not seem to be striking differences in the crust and lithosphere east and west of the Rift Valley, despite the different geological history of the two sides. In the west, Archaean lithosphere underlies the Archaean Nyanza Craton (ANC), whilst in the east Proterozoic lithosphere exists below the Mozambique Orogenic Belt (MOB).

*The Microseismic Network*

Inversion of the local P-wave arrival times around Lake Baringo down to a depth of 9 km shows a low velocity zone related to surficial geothermal activity, as previously shown for the Lake Bogoria area (Tongue et al., 1994).

On a larger scale (Ritter and Achauer, 1994) the relationship between velocity variations and geological features can still be recognised (Fig. 2.5). Data from the central Kenya Rift, from Lake Baringo to Lake Naivasha and across the whole graben shows low velocities in the graben to 5 km depth which can be related to the sedimentary-volcanic infill, and high velocities related to mafic intrusions. Down to 10 km depth low velocities are related to the extension of the Nyanza Rift; the high velocities to magma reservoirs. High velocity anomalies continue to depths of 15-35 km and have been related again to mafic intrusions connected to an axial high velocity above the Moho and interpreted as mantle material rising into the lower crust.

Part of the outcome of the seismicity study has been mentioned in Section 1.5.4 the KRISP 90 data (Lake Baringo array) being studied jointly with the KRISP 85 data (Lake Bogoria array) (Tongue, 1992).

The regional stress distribution was studied using the analysis of focal mechanisms from both surveys together with a shear wave polarisation study for the Lake Bogoria data and indicated E-W or NE-SW extension for the Lake Bogoria area and WNW-ESE extension for the Baringo region. Most of the activity recorded by local networks near Lake Bogoria and Lake Baringo lies between the surface and 16 km, with $M_I \leq 2$. It is
Figure 2.5: Contoured inversion results (after Ritter and Achauer, 1994).

NKU = Nakuru
KAB = Kabarnet
LON = Londiani
--- = main boundary fault
\( \text{---} \) = Metkei-Marmenet accommodation zone
\( \triangle \) = seismic station
deemed associated with the older faults forming the graben, and located mostly beneath the centre of the rift (Tongue, 1991, 1992; Tongue et al., 1992, 1994).

2.4 Seismic Velocities of Rock Samples

Laboratory measurements of P-wave velocity have been made on rock samples from locations in close proximity to the KRISP 90 refraction/wide angle reflection profiles (Fig. 2.6 after Mooney and Christensen, 1994). The results have been interpreted in relation to the seismic velocities recorded from the profiles to deduce the composition of the crust beneath the Kenya Rift (Fig. 2.7).

The velocity/pressure curves shown in Fig. 2.7 relate to the southern portion of the Rift, as indicated by the superimposed velocity/depth curve.

Supracrustal volcanic rocks gave laboratory-measured velocities higher than the 3.1-5.15 km/s recorded, due to extensive fracturing and jointing of the rocks in the field. Low density pyroclastic rocks interbedded with the lavas and high pore pressure could also contribute to reduce the velocity values observed in the field.

Laboratory velocities of the upper crustal rocks from the metamorphic complexes show good agreement with the velocities recorded along the profiles where the basement outcrops. Most of the samples have been collected near the cross-rift Line D, and therefore the results will be commented on in Section 3.4.2.

At depth the crustal composition is suggested to be gneissic, as the intrusive equivalent of the volcanic rocks exposed at the surface (gabbro and syenite) gave laboratory derived velocities which were too high (above 6.2 km/s) for them to be the dominant rock type in the upper crust of the rift.

The curves from the laboratory measurements reported here have been corrected for the effect of a high geotherm estimated within the rift and show a negative velocity gradient with depth. Since a positive or zero gradient is observed in the crust, Mooney and Christensen (1994) propose that the crustal composition becomes more mafic at depth (Fig. 2.8).

At mid and lower crustal depths the interpretation of composition becomes less reliable,
Figure 2.6: Map of sample locations documented in Fig. 2.7. Samples 8, 14, 15, 16: granitic gneisses; 19: dioritic gneiss; 30: tonalite gneiss. Samples 2a, 3: basalts; 7: basaltic trachyandesite; 17a, 17b: tephriphonolites; 21a, 21b: tephrite basanites (after Mooney and Christensen, 1994).
Figure 2.7: Comparison of the average velocity/depth structure beneath the southern portion of the axial rift profile with metamorphic rocks (top) and volcanic rocks (bottom).
Figure 2.8: Composition of the crust beneath the Kenya Rift at the Equator from crustal seismic velocity structure and laboratory measurements on Kenyan rock samples (after Mooney and Christensen, 1994).
as no outcrops are available. For the mid-crust under the flanking areas, gneisses of intermediate composition are favoured over an intrusive equivalent of the volcanic rocks at the surface, which would give a velocity value too high for the observed data. Beneath the rift axis at mid-crustal depths a mixture of mafic intrusives and metamorphic rocks (at a grade higher than those outcropping) is proposed. The lower crustal velocity recorded, 6.7-7.0 km/s under the flank, is higher than that expected from the intrusives corresponding to the exposed gneissic rocks and therefore amphibolite to granulite facies mafic rocks are hypothesised for the composition of this region. Beneath the southern rift, the proposed composition is of mafic residuum accreted during the upper mantle melt differentiation, after considering the high heat flow and the copious volcanics. These hypotheses will be discussed further in Section 3.4.2.

2.5 Seismic Refraction Program - The Cross-rift Profile

The definition of crustal thickness and compositional variations in the basement rocks is essential to constrain models of the rifting process. The principal questions to be resolved concern the crustal signature of the Archaean and Proterozoic terranes; the geometry of the graben and the depth of the detachment system (Bosworth et al., 1986); the crustal thickness under Kaptagat (Maguire and Long, 1976); the depth of the Kerio basin; the crustal transition from axis to shoulder of the rift and the definition of the low velocity zone in the upper mantle and its possible influence on the overlying crust.

2.5.1 Set up and Shooting

One of the most significant technical results from KRISP 85 was that the lake shot points were found to provide the best source of energy; where no underwater shots were possible, borehole shots had to be planned. Borehole shots were positioned where the ground-water surface was shallow. Moreover, the shot points and recording stations had to be accessible by road and/or by boat.

Additional requirements had to be satisfied for the deployment of line D:

- the line should be perpendicular to the strike of the rift in the central part of a
CHAPTER 2. KRISP 90

postulated half-graben, and

- the line should use the results of previous seismic surveys: the Kaptagat array and the KRISP 75 results (Maguire and Long, 1976; Swain et al., 1981) provided extra data to help to constrain the final model.

The cross-rift line D was sited nearby Lake Baringo just to the north of the apex of the Kenya Dome, from the Kavirondo Gulf in Lake Victoria (VIC shot point), to Chanlers Falls in the east (CHF shot point). The total length of 450 km was designed to enable phases from the upper mantle to be seen from the end shot points before the seismic wave energy entered the Rift proper. A shot point planned in the Ewaso Ngiro River at the east end of the line had to be abandoned for security reasons.

Two-hundred-and-six mobile seismographs were used and about 40 stations were added from the temporary earthquake networks of Leicester University (LU) and the Institut de Physique du Globe (IPG) Paris. The stations were deployed with a nominal spacing of about 1 km in the rift valley; about 2 km spacing was used on both flanks (from c.60 km to the rift margin and on the eastern shoulder to 350 km offset); around 5 km spacing was than used to both ends of the line.

The easternmost 50 km of the line was set up during the deployment of the KRISP 90 flank line, when the shot at CHF was fired.

The line was therefore positioned so that underwater shots could be used in Lake Victoria and in Lake Baringo, the latter providing good seismic wave transmission during the 1985 experiment. Both resulted in good to excellent recording. The borehole shots KAP, TAN, BAS and CHF also provided good energy propagation. Six shot points were used in total:

- VIC (underwater shot in Lake Victoria).
- KAP (borehole shot at Kaptagat, c.155 km from VIC).
- BAR4 (in Lake Baringo, c.236 km from VIC). In the same lake three small shots were separated at 250 m intervals in an E-W direction: BAX1, BAX2, BAX3. These were intended to provide some high resolution data; unless otherwise indicated, BAR will refer to the BAR4 data set.
- TAN (borehole shot near Tangelbei, c.264 km from VIC).
• BAS (borehole shot near the village of Barsalinga, c.351 km from VIC).
• CHF (Chanlers Falls, c.451 km from VIC), primarily shot for the flank line deployment.

Details on the shot point locations are presented in Table 2.1.

<table>
<thead>
<tr>
<th>Shot</th>
<th>Approx. range</th>
<th>Date(1990)</th>
<th>Local time</th>
<th>Latitude(N)</th>
<th>Longitude(E)</th>
<th>Alt.(m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>BAS</td>
<td>09-220</td>
<td>09.02</td>
<td>02:00:05.66</td>
<td>0°38.15'</td>
<td>36°04.15'</td>
<td>1000</td>
</tr>
<tr>
<td>VIC</td>
<td>00-450</td>
<td>09.02</td>
<td>02:03:00.05</td>
<td>0°12.86'</td>
<td>34°07.63'</td>
<td>1100</td>
</tr>
<tr>
<td>KAP</td>
<td>00-205</td>
<td>09.02</td>
<td>02:06:00.00</td>
<td>0°30.53'</td>
<td>34°24.03'</td>
<td>1200</td>
</tr>
<tr>
<td>BAX1</td>
<td>00-230</td>
<td>09.02</td>
<td>02:09:02.25</td>
<td>0°38.14'</td>
<td>38°04.26'</td>
<td>1000</td>
</tr>
<tr>
<td>TAN</td>
<td>00-185</td>
<td>09.02</td>
<td>02:12:00.71</td>
<td>0°48.85'</td>
<td>36°16.38'</td>
<td>1200</td>
</tr>
<tr>
<td>BAX2</td>
<td>00-230</td>
<td>09.02</td>
<td>02:15:03.32</td>
<td>0°38.17'</td>
<td>38°04.50'</td>
<td>1000</td>
</tr>
<tr>
<td>BAS</td>
<td>00-280</td>
<td>09.02</td>
<td>02:17:53.30</td>
<td>0°47.38'</td>
<td>37°06.97'</td>
<td>1100</td>
</tr>
<tr>
<td>BAX3</td>
<td>00-280</td>
<td>09.02</td>
<td>02:19:55.16</td>
<td>0°38.13'</td>
<td>38°04.18'</td>
<td>1000</td>
</tr>
<tr>
<td>CHF</td>
<td>00-255</td>
<td>10.02</td>
<td>17:12:00.00</td>
<td>0°48.96'</td>
<td>38°00.76'</td>
<td>700</td>
</tr>
</tbody>
</table>

Table 2.1: Line D shot point locations.

A study of the energy transmission during KRISP 90 (Jacob et al., 1994) proved that ‘seismic coupling at optimum depth in water is more than 100 times what it is in a borehole in dry, loosely compacted material’. Allowing for the source coupling, greater observation ranges have been found on the flanks than within the rift.

The noise at receiving sites was minimised by choosing ‘quiet’ times of the day for shooting, variable from place to place. Because ‘cultural’ noise is not a major problem in Kenya, ‘quiet’ times were mainly chosen to avoid ‘natural’ noise, principally due to the wind which varied in its strength at different times at different places: at Lake Baringo, for example, a light wind in the morning often becomes very strong in the afternoon. In fact, one of Lake Baringo shots was impeded by a storm during the shot deployment and a reserve recording window had to be used, which was still in a ‘quiet’ time.

Inflatables or local fishing boats (for the shots further away from the shore) were used to deploy underwater shots; the explosive was dispersed in a number of well separated charges which, as shown by Jacob (1975), provide linear addition in the velocity output. In the case of KRISP 90, however, the output was lower than expected, indicating that
CHAPTER 2. KRISP 90

2.5.2 Topography

The relief along the line varies from about 1000 m at Lake Victoria to 3000 m at the top of the Elgeyo escarpment, after which a drop of nearly 1500 m occurs. The Kamasia Hills, with a relief of about 2000 m, are followed by the Lake Baringo basin at about 1000 m elevation. Another rise to about 2000 m on the Laikipia plateau is gently followed by a decrease to 700 m at Chanlers Falls.

The relative position of the stations along the line used for the interpretation was calculated as an average between the distances of each station from each shot point. The relative distance between the shot points was determined by the Range utility from the USGS package written for a Macintosh IIci (Luetgert, 1992), knowing the coordinates of each of them from the KRISP 90 data report (KRISP Working Group, 1991). The distance of the station from each shot point is available as input from the SEG-Y data file using the TRAN4 program (Luetgert, 1990a). The average was calculated by the program TOPO (Appendix A).

The shot points and the stations, especially between TAN and CHF are offset with respect to a straight line, being sited on driveable roads. The shot points themselves are

<table>
<thead>
<tr>
<th>Water Shot</th>
<th>Total charge (kg)</th>
<th>Charge unit (kg)</th>
<th>No. Spacing (m)</th>
<th>Approx. geometry (m)</th>
<th>Water depth (m)</th>
<th>nature of bottom</th>
</tr>
</thead>
<tbody>
<tr>
<td>BAZ4</td>
<td>300</td>
<td>25</td>
<td>20</td>
<td>12</td>
<td>60x60</td>
<td>5</td>
</tr>
<tr>
<td>VIC</td>
<td>1000</td>
<td>25</td>
<td>40</td>
<td>15</td>
<td>180x30</td>
<td>25</td>
</tr>
<tr>
<td>BAX1</td>
<td>300</td>
<td>25</td>
<td>12</td>
<td>12</td>
<td>50x30</td>
<td>5</td>
</tr>
<tr>
<td>BAX2</td>
<td>300</td>
<td>25</td>
<td>12</td>
<td>12</td>
<td>50x30</td>
<td>5</td>
</tr>
<tr>
<td>BAX3</td>
<td>300</td>
<td>25</td>
<td>12</td>
<td>12</td>
<td>50x30</td>
<td>5</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Borehole</th>
</tr>
</thead>
<tbody>
<tr>
<td>Shot</td>
</tr>
<tr>
<td>----------</td>
</tr>
<tr>
<td>KAP</td>
</tr>
<tr>
<td>TAN</td>
</tr>
<tr>
<td>BAS</td>
</tr>
<tr>
<td>CHF</td>
</tr>
</tbody>
</table>

Table 2.2: Line D shot point characteristics.

the charges were not separated adequately (Jacob et al., 1994).

Details on the shot point characteristics are presented in Table 2.2.
offset from the recording positions, noticeable in the apparent low density of traces close
to the shot points in the displayed sections.

2.5.3 Pre-site Survey

The shot points and the recording sites were located a few months before the experiment,
during a two-month expedition in August and September 1989. Sites were marked and
coordinates determined by GPS measurements to an accuracy of ±15 m (Appendix A).

2.5.4 Communication and Timing

A radio network was set up during the experiment to ensure communication between the
field parties and the headquarters and to control the master clock of the mobile units at
shot points and observers' field camps. The master clocks were synchronised daily with a
time signal emitted by one of the headquarters and regularly compared with the universal
time broadcast by the Moscow short-wave radio station.

The network consisted of (1) short-wave radios for the communication over long dis­
tances and time synchronisation; (2) car radios for each mobile group for emergency and
car-to-car communication over distances of tens of kilometres and (3) hand-held radios for
communication over a few hundred metres, e.g., from shore to lake shot points.

2.5.5 Recording Equipment

During the experiment 1-component digital Seismic Group Recorders (SGR IIIIs) from
Stanford University (jointly maintained by the U.S. Geological Survey and the U.S. Pro­
gram for Array Seismic Studies of the Continental Lithosphere, PASSCAL) were the main
recorders used. A quartz clock was added to the SGRs to function as a timer and refer­
ence for pre-programmed recording windows. The clocks were set and synchronised to the
master clocks.

A number of 1-component analogue tape recording units were supplied by the Dublin
Institute for Advanced Studies (DIAS).

All stations were set with 2 Hz geophones, most of which were Mark II-L-4 vertical
seismometers (Prodehl et al., 1994b).
In addition LU’s and IPG’s 3-component stations were also used, as stated above. These included Geostore (analogue), Refraction Technology RefTek 72 (digital) and Teledyne Geotech PDAS100 (digital) recorders equipped with 1 Hz seismometers (Willmore Mark IIIa and Teledyne Geotech S13). The digital recording stations used the Omega world-wide radio signal as their timing system, accurate to \( \Delta t < 10 \text{ ms} \) (Tongue, 1992).

2.5.6 Data Processing

Preliminary data processing was performed at the University of Karlsruhe in Germany. The data from the IPG’s stations were unusable due to the failure to record a suitable time signal. LU’s stations recorded at 100 samples/s.

The SGR data were recorded digitally at 500 samples/sec. The data recorded by the DIAS analogue stations were digitised at 500 samples/s and processed into the same format as that from the SGRs. Digital processing of the DIAS’ data included clock drift correction, de-spiking and anti-alias frequency low pass filtering at 80 Hz, using a 4-pole Butterworth zero phase shift filter, before plotting the data in the form of reduced time-distance record sections.

These initial P-wave record sections were presented to all KRISP 90 participants decimated for plotting to 200 samples/s and low pass filtered at 12 Hz for the underwater shots and 15 or 25 Hz for the borehole shots. The reduction velocity chosen for the record sections were 6 or 8 km/s for the crustal and mantle phases respectively. The S-wave record sections were plotted at 500 samples/s and low-pass filtered at 12 Hz for the underwater shots and 12, 15, 18 or 25 Hz for the borehole shots; the reduction velocity chosen for the record section plotting was 3.46 km/s (KRISP Working Group, 1991).

At LU the USGS SEG-Y refraction wide angle reflection processing package was used for section display. RSEC90 is an interactive program written by J. Luetgert (1990b), to be run under VMS on a VAX computer.

Different band-pass filters were selected for the different shots, specified in the figure legends in Chapter 3. These were chosen after various trials giving the greatest signal to noise improvement, and being consistent with the results of the spectral analysis. The amplitudes were scaled to a uniform maximum value (trace normalised).
Spectral Analysis

Spectral analysis was carried out on selected seismograms from the dataset from all the line D shot points, after band-pass filtering from 0 to 12 Hz. The major P and S phases were investigated at Karlsruhe University using the DISCO 8.1b processing package of the Cogniseis Development Inc. Specifically, the routine used, SPECTRM, transforms traces from the amplitude-time domain to the spectral amplitude-frequency domain and plots the spectrum. The option to compute the spectral estimates using a Fast Fourier Transform was selected in the present study.

Power spectra plotted in Fig. 2.9 represent an energy-density versus frequency relationship, or the square of the amplitude-frequency response (Sheriff, 1991). A summary of all the frequency versus distance values is given in Fig. 2.10; a description of the main points arising from the diagram is presented below. For phase nomenclature see Sections 3.3 and 4.3.1.

- **VIC:**
  - near source P frequency is 12 Hz,
  - crustal phases \( a \) and \( b_1 \) have a frequency between 4 and 10 Hz; \( b_1 \) shows a frequency higher than \( a \) at short offset, and
  - P mantle phase frequency content changes from 6 Hz at short offset to 3-4 Hz in the rift; the filtering effect must be at shallow depth because beyond the rift the frequency increases again for the \( d \) and \( d_1 \) phases.

- **KAP:**
  - near source P frequency is 11 Hz,
  - crustal phases \( a \) and \( b_1 \) in both directions have a frequency between 2 and 10 Hz, and
  - P mantle phases are filtered by the rift as for VIC data: the \( c \) frequency increases from 3 to 10 Hz outside the rift.

- **BAR:**
  - the near source traces appear saturated, preventing us from analysing the frequency content,
  - crustal phases \( a \) and \( b_1 \) in both directions have a frequency between 4 and 6 Hz, and
Figure 2.9: Examples of power spectra of P and S phases from VIC shot point.
Figure 2.10: Summary of spectral analysis for P and S phases from the cross-rift line D. Continuous lines: P phases; dashed lines: S phases; thick lines: reflected phases; thin lines: diving phases. Distances are marked E and W from S-P. The rift graben is indicated by the shaded area on the distance axes.
- the rift filtering effect on the P mantle phases can be seen in the increase from 6 Hz in the rift to 10 Hz outside rift.

- TAN:
  - near source P frequency is 10 Hz,
  - crustal phases $a$ and $b_1$ in both directions have a frequency between 4 and 10 Hz, and
  - the poor quality P mantle phases have a frequency around 7 Hz.

- BAS:
  - near source P frequency is 10 Hz,
  - crustal phases $a$ and $b_1$ in both directions have a frequency between 4 and 10 Hz, and
  - mantle phases increase beyond 200 km offset to the west, i.e. on the western rift margin. Higher frequencies are observed to the east than to the west.

CHF has not been studied because of the insignificant range over which the shot has been recorded along line D.

There are some comments to be made:

In Fig. 2.10 the S phases examined are also presented. They generally show a similar trend with distance as the P phases, and also they are of lower frequency. $u$ from VIC is an exception, demonstrating that, although the crustal phases behave as expected, with S frequency lower than P (Section 4.2), deeper in the crust the P phases are attenuated more strongly than S. Although it is difficult to explain this phenomenon, it might imply that the P and S rays followed slightly different paths, sampling regions with different absorption.

Generally, diving waves $a$ and $d$ show lower frequency content than reflected phases $b_1$ and $c$ at the same offset. Again, this might be due to different ray-paths followed by the mid-crustal reflection and the upper crustal diving waves, or to the small scale structure of the reflecting interfaces.

Local high frequencies at a single station at large offset might be related to scattering from small scale fabric which generates high frequency phases; the interference of phases with a different frequency content or shifted with respect to each other could also determine
the appearance of localized high frequencies at large offset.

High frequencies at large offsets beyond a lower frequency arrival may be related to the damping effect of the shallow structure in the rift, i.e. it is possible that diving waves beneath the rift suffer less attenuation than rays passing through the sedimentary-volcanic sequence within the rift.

The position of the shots in relation to the rift seems to be most important. It would be expected that data from the borehole shots have a higher frequency content than those from underwater shots. However, the rift shot points BAR and TAN show a lower frequency content at short offset (i.e. inside the rift) for the crustal phases than the shots outside the main graben, VIC, KAP and BAS, demonstrating the importance of the shot point position.

Examples of Fourier spectra (spectral amplitude versus frequency) for all the shot points can be found in Jacob et al. (1994), which confirm that at short range a higher frequency content is found for the borehole shots on the rift flank than for the shots in the rift. The underwater shot in Lake Victoria shows a smaller decrease in spectral amplitude with distance than the lake shots in the rift valley. To explain the spectral distribution it has been suggested that attenuation within the crust beneath the rift results primarily from scattering (Jacob et al., 1994), but absorption in the sedimentary-volcanic sequence has also to be considered.

### 2.5.7 The Seismic Data

Recordings along the profile were made from all eight shots during deployment D and from CHF during deployment E.

As anticipated the lake shots provided energy to a greater distance than the borehole shots.

The observation scheme diagram (Fig. 2.11) shows that:

- VIC provided good data along nearly the complete line (420 km);
- KAP data are visible to a maximum range of 240 km to the east, and to the end of the line to the west;
- BAR first arrivals could be identified along the complete profile, in both directions;
TAN energy is distinguishable to the end of the line to the east, and to about 170 km at most to the west. It was the least effective shot;

BAS reached a distance of 200 km to the west, the maximum offset for a borehole shot along line D.

The quality of the data decreases in the Rift Valley for the shots fired on the flanks, possibly resulting from reverberations in the rift’s layered infill. This will be discussed in Section 3.4.1.

2.5.8 Profile Geology

Geological maps have been produced by the Mines and Geological Department of the Government of Kenya (1953; 1959; 1964) and recently by the East African Geological Research Unit of London and Leicester Universities (E.A.G.R.U., 1973; 1976). A schematic geological section (Fig. 2.12) is shown, approximately coincident with the cross-rift seismic line.

Note that the subdivision into five main units is related specifically to the geological interpretation of the seismic model; the locality names are in Fig. 1.7 and 1.11. In Section
CHAPTER 2. KRISP 90

3.4.2 it is possible to see how each seismic unit may comprise many geological formations:

1. "Nyanzian intrusives" includes granites, adamellites and syenite batholiths of the Nyanzian System (Photo 2).

2. "Nyanzian metavolcanics" refers to the basalts, andesites, rhyolitic tuffs and rhyolites of the Nyanzian System. From VIC to c.55 km the line crosses an Archaean Greenstone Belt.

The Nandi fault is associated with a mylonitic band of c.10 km width (Shackleton, 1986).

3. "Mozambique Orogenic Belt" includes:

(3.1) in the western sector (Kapsabet area) the biotite gneisses and banded gneisses of the basement system,

(3.2) in the central sector (Elgeyo escarpment: Photo 3) the hornblende-biotite gneisses of the same system (Photo 4), and

(3.3) in the eastern sector (Chanlers Falls) the migmatites, intrusive rocks and paragneisses of the Baragoi, Barsaloi and Loldaika tectonothermal events.

4. "Tertiary-Quaternary" volcanics involve:
(4.1) in the western sector (Kaptagat area) the lavas and tuffs of the Miocene Usain Giallu Phonolites (Photo 5) and the Elgeyo Volcanics formations (Photo 6),

(4.2) in the central sector (southern Kamasia Hills, also called the Tugen Hills Group), the Miocene Tum Phonolites, the sedimentary Ngorora Formation, the Mio-Pliocene Ewael Phonolites, Eon Basalts, Kabarnet Trachytes (Photo 7) and Kaparaina Basalts and the Pleistocene Lake Hannington Trachyphonolites, Kapturin Beds and Baringo Trachyte, and

(4.3) in the central-eastern sector (Baringo-Laikipia area) are the Miocene trachyphonolites and hawaiites of the Murgomul Volcanics, the phonolites of the Rumuruti Group formations and the Pleistocene Kaphurin Beds.

5. "Pleistocene sediments" includes:

(5.1) in the western sector (Lake Victoria) the lacustrine and fluviatile deposits,

(5.2) in the Elgeyo fault area, the alluvial Kerio Valley Beds, and

(5.3) in the central sector (Baringo area) the fluvo-lacustrine sediments of the Baringo-Hannington Lake Beds.

These formations will be referred to in Section 3.4.2 although the seismic units finally modelled will not exactly correspond to this preliminary subdivision.

2.6 Summary

The KRISP 90 project has been presented, and the results from the refraction axial and flank lines summarised. Mafic intrusions at upper crustal level, thickness variations of the lower crustal layer, crustal thinning beneath the axis compared to the flanks, anomalously low velocity in the uppermost mantle along the axis, and high-low velocity alternations in the upper mantle are some of the most significant features from the interpretation of the two lines.

From the teleseismic data recorded in the southern Kenya Rift it appears that mafic intrusions in the crust must be of limited extent, and that a low velocity body extends from the Moho to c.160 km depth (Achauer et al., 1994; Slack and Davis, 1994).

Seismic tomography in the upper crust around Lake Baringo relates low velocities to
geothermal activity or sedimentary infill and high velocities to intrusive bodies or magma reservoirs (Tongue et al., 1994; Ritter and Achauer, 1994).

The stress distribution in the Baringo region has been attributed to WNW-ESE extension, most of the seismic activity occurring above 16 km depth (Tongue et al., 1992, 1994).

Analysis of P-wave velocities from rock samples provides an important insight into the possible composition of crustal rocks when compared to the seismic velocities deduced from large scale experiments (Mooney and Christensen, 1994).

The cross-rift deployment is reported in detail and results from the data processing presented. Spectral analysis of the main P- and S-wave phases shows that the attenuation results mainly from scattering and absorption occurring beneath the crust in the rift with less damping during passage through the lower crust and upper mantle.

A schematic geological section along the cross-rift line is the starting point for the geological interpretation of the seismic model derived in the following chapters.
Photo 2: Nyanzian intrusives at c.80 km along line D.

Photo 3: Elgeyo escarpment from the Rift Valley floor.
CHAPTER 2. KRISP 90

Photo 4: Hornblende-biotite gneisses of the basement system at the Elgeyo escarpment.

Photo 5: Miocene Uasin Gishu Phonolites from the Rift Valley floor, at c.0.5° N.
Photo 6: Miocene Elgeyo Volcanics formations at the Elgeyo escarpment.

Photo 7: Pliocene Kabarnet Trachytes at the Kamasia Range.
Chapter 3

Kinematic Modelling of P-waves

3.1 Introduction

The basic objective of long range seismic (refraction) surveying is to determine the velocity depth structure of the earth by the surface recording of various artificially generated (refracted) waves. Changes in the subsurface acoustic impedance result in perturbations of the various wavefronts generated by the source and may be derived from the recorded seismic wavefield.

Interpretation of the P-wave field from the processed seismic sections involves a number of different procedures. In this chapter travel time modelling following picking of the main phases on the record sections is discussed. The procedures for 1-D and 2-D modelling with laterally varying velocities will be described. The seismic model will be used to provide starting density distribution values for the gravity modelling along the profile.

3.2 Picking

The first stage of the data analysis involves the identification of particular seismic phases on the record sections, the seismograms being displayed on time versus distance plots.

This is the most delicate stage of forward modelling because of its subjectivity, as demonstrated by workshops on crustal seismology (Finlayson and Ansorge, 1984; Mooney and Prodehl, 1984).
The main assumptions usually made in the interpretation of seismic refraction data are that:

1. the subsurface structure is subhorizontally plane layered, and
2. that the lateral velocity variations are smaller than the vertical variations, the shot point interval being comparable to the scale of the lateral variation (Mooney, 1989).

Bearing these in mind, phase correlation can be performed, usually identifying the beginning of the event, sometimes the next minimum or maximum. A correct phase correlation needs:

1. the spacing of detectors to be closer than half an apparent wavelength at surface,
2. the amplitude of the crustal phases to be above the noise level,
3. the apparent velocity to lie within a realistic range,
4. the travel time branch to be long enough to establish coherency of waveform, and
5. the crustal phase not to be confused with multiples, or phase conversions.

Many real discontinuities are formed by lamellae generating wave groups visible only over a short range; in this case phase correlation is only possible on a few seismograms. Phase correlation leads to group correlation, in which segments of travel time branches are joined to represent arrivals from idealised sharp discontinuities (Giese and Prodehl, 1976).

Assuming a simple model of a layered earth, the principal crustal and upper mantle phases to be expected are:

- diving waves in the uppermost layer, represented by a convex upward curve passing through the origin of the time versus distance plot,
- diving waves continuously refracted in the top part of each lower layer represented by convex upward curves, and
- reflected waves at the interface of impedance discontinuities represented by hyperbolas asymptotic to the curve of the refracted waves in the layer above, and tangential at the critical distance to the one in the layer below.

In order to better display the different curves on the time-distance plot a reduction velocity is applied. This is equivalent to rotating the sections with respect to the origin. The time scale is represented by reduced time, \( T_r \), with \( T_r = T \cdot \frac{X}{V_r} \), where \( T \) is the true time.
Chapter 3. Kinematic Modelling of P-waves

Figure 3.1: Basic traveltine diagram. The numbers on both axes indicate approximate values. The curves can be shifted as much as 30–50 km and up to ±2–3 s (after Prodehl, 1979).

X is distance and \( V_r \) the reduction velocity. The reduction velocity is normally chosen close to the main phase velocity one wants to investigate. For crustal phases it is conventional to choose \( V_r = 6 \) km/s, whereas for mantle phases \( V_r = 8 \) km/s.

For crustal investigations the main phases to consider are (Fig. 3.1, Prodehl, 1979):

a - \( P_g \) - diving waves through the upper crust,
b - \( P_i \) - diving waves through the middle/lower crust beneath intracrustal boundaries,
c - \( P_m P \) - reflection from the Moho discontinuity,
d - \( P_n \) - diving waves in the uppermost mantle, and
\( b_1 \) - \( P_i P \) - reflection from middle/lower crust discontinuity.

The phases’ arrival times were identified using group correlation across the record.
CHAPTER 3. KINEMATIC MODELLING OF P-WAVES

Picking of the first arrivals has been undertaken using RSEC90 (Luetgert, 1990b), which enables one to plot and pick the seismograms from a Tektronix 4014 graphic screen. The travel time branches identified on the complete record sections were picked on an expanded portion of the section for better resolution.

Due to the loss of high frequency content in the signal travelling from source to receiver, a progressive rounding of the first break is observed. The picking accuracy decreases with range, but it throughout is better than ±0.05 s.

Some of the later phases have been clearly identified on the best data sets (see observation scheme) and used as a constraint on the starting seismic model. These include reflections from intracrustal boundaries and from the Moho discontinuity for the intra-rift shots. The modelled interfaces have then been ray-traced. The calculated travel times from this model helped to identify the same phases on the data sets of poorer quality. Picking in this case resulted from an interaction with the modelling. The phases identified are therefore strictly related to the model, and have to be described in terms of the model that is finally produced. The picking accuracy is here estimated as better than ±0.2 s.

If the accuracy was estimated as less than ±0.3 s the phase would not be picked.

3.3 Record Section Description

The band-pass filtered, trace normalised, reduced time record sections are presented for the six shots; the lines represent final theoretical ray-traced travel times. A description of the phases identified is included. See Table 2.1 for shot location coordinates, and Appendix A for station location.

The following notation is used:
- $P_s$ (phase a): direct arrival from the sediments or the volcanic cover,
- $P_g$ (phase e): diving waves in the upper crust; the crystalline basement is represented in the west by the Archaean Nyanza Craton (ANC), in the central and eastern sector by the Mozambique Orogenic Belt (MOB),
- $P_t$ (phase b): diving waves from below a mid-crustal boundary,
CHAPTER 3. KINEMATIC MODELLING OF P-WAVES

- $P_{\text{II}}P$ and $P_{\text{I}}P$ (phases $b_1$ and $b_2$): intracrustal reflections,
- $P_{\text{m}}P$ (phase $c$): Moho reflection,
- $P_n$ (phase $d$): diving waves in the uppermost mantle,
- $P_{\text{um}}P$ (phase $d_1$): intramantle reflection, and
- $x$: a possible multiple from crustal boundaries.

VIC - Fig. 3.2, 3.3, 3.4

All the distances are from the shot point VIC in Lake Victoria.

$a$: is visible only on the first seismogram from the shot point, at 5 km distance with a 2.9 km/s velocity.

$a$: it is seen with an apparent velocity of 5.8 km/s to 20 km, 6.3 km/s from 20 to 50 km, 6-6.1 km/s from 50 to 160 km (Fig. 3.2).

$b_1$: the critical point is around 50 km, and energy is visible to 150 km.

$b_2$: this can be observed from 100 to 180 km.

c: it is a high amplitude phase, with its critical point at about 80 km, and clearly discernible to 200 km; beyond that distance, energy reverberating in the rift is seen to 270 km.

d: this is a first arrival from 188 km to 450 km (on Fig. 3.4 plotted only to 400 km). Over the last 30 km it has a small signal/noise ratio of less than 2.0.

$d_1$: it has been picked from 270 to 350 km (Fig. 3.4).

KAP - Fig. 3.5, 3.6

Distances are measured from the shot point KAP in the Kaptagat Forest.

$a$: this phase travels in the volcanic cover with a velocity of 4.2 km/s.

$a$: it is the first arrival in both directions from about 9 to 150 km; it has an apparent velocity of 6-6.1 km/s in both directions with large delays especially to the east towards the rift valley.

$b_1$: this is a clear large amplitude phase to the west from 60 to 150 km. To the east it is not visible.
Figure 3.2: Trace-normalized band-pass filtered (1-25 Hz) record section for shot-point VIC recorded to the east along the cross-rift line D emphasizing phase a. Note the display gain increased by four and the time scale by two with respect to Fig. 3.3. $V_r=6$ km/s (ray-traced phase notation for Fig. 3.2, 3.3, 3.4: $a=P_g$, $b_1=P_{11}P$, $b_2=P_{12}P$, $c=P_mP$, $d=P_n$, $d_1$=mantle reflection).
Figure 3.3: Trace-normalized band-pass filtered (1-25 Hz) record section for shot-point VIC recorded to the east along the cross-rift line D. $V_p=6$ km/s (ray-traced phase notation: see Fig. 3.2 caption).
Figure 3.4: Trace-normalized band-pass filtered (1-25 Hz) record section for shot-point VIC recorded to the east along the cross-rift line D emphasizing the mantle phases $d$ and $d_1$. $V_p=8$ km/s (ray-traced phase notation: see Fig. 3.2 caption).
Figure 3.5: Trace-normalized band-pass filtered (5-15 Hz) record section for shot-point KAP recorded to the west along the cross-rift line D. $V_r=6\text{km/s}$ (ray-traced phase notation for Fig. 3.5, 3.6: $a_s=P_s$, $a=P_g$, $b_1=P_{11}P$, $c=P_mP$, $d=P_n$).
Figure 3.6: Trace-normalized band-pass filtered (5-15 Hz) record section for shot-point KAP recorded to the east along the cross-rift line D. $V_r=6$ km/s (ray-traced phase notation: see Fig. 3.5 caption).
c: the reflection from the Moho to the west has a critical point at 105 km, to the east at 110 km.

d: it is observed only to the east from 170 to 260 km.

**BAR - Fig. 3.7, 3.8**

All distances are from the shot point BAR in Lake Baringo.

- **a:** arrivals from the rift infill are identified by three branches in both directions, corresponding to velocities of about 2.1, 3.8 and 5.8 km/s.
- **b:** it is a first arrival from 15 km to the west and about 20 km to the east having an apparent velocity of 6-6.1 km/s.
- **c:** this is first arrival from 160 to 180 km, (130 to 180 km to the east), but it is still visible behind the d phase towards both ends of the record section.
- **b:** it follows b from 140 to 230 km to the west and from 70 to 220 km to the east.
- **c:** this high energy phase is seen from about 70 km and picked to 140 km to the west, and from 80 km to 220 km to the east.
- **d:** this phase is a first arrival from about 190 km to the end of the section in both directions; the picking of this phase has been performed on a composite record section from the four shots in Lake Baringo.
- **a:** this phase can be clearly identified in both directions from about 130 km to 160 km.

**TAN - Fig. 3.9, 3.10**

Distances are measured from the shot point TAN in Tangulbe.

- **a:** is subdivided into three branches of 2.9, 3.6 and 5.5 km/s apparent velocity, to about 15 km in both directions.
- **b:** is the first arrival, delayed by the rift infill to the west to 135 km, and to the east to c.150 km.
- **b:** it can be picked to 160 km to the west and to the end of the section to the east.
- **b:** can be identified from 105 km to the end of the section to the west, but with poor confidence beyond 180 km. To the east it can be picked from 50 to 140 km.
Figure 3.7: Trace-normalized band-pass filtered (5-15 Hz) record section for shot-point BAR recorded to the west along the cross-rift line D. $V_s=6$ km/s (ray-traced phase notation for Fig. 3.7, 3.8 $a_s=P_s$, $a=P_g$, $b=P_i$, $b_1=P_i P$, $c=P_m P$, $d=P_n$, $x$=unidentified phase discussed in the text). Note: dashed line identifies observed but not ray-traced phase $d$ (see text).
Figure 3.8: Trace-normalized band-pass filtered (5-15 Hz) record section for shot-point BAR recorded to the east along the cross-rift line D. $V_r=6$ km/s (ray-traced phase notation: see Fig. 3.7 caption).
Figure 3.9: Trace-normalized band-pass filtered (1-15 Hz) record section for shot-point TAN recorded to the west along the cross-rift line D. \( V_r = 6 \text{ km/s} \) (ray-traced phase notation for Fig. 3.9, 3.10: \( a_s = P_s \), \( a = P_g \), \( b = P_i \), \( b_1 = P_{i1}P \), \( c = P_mP \), \( d = P_n \), \( x \) = unidentified phase discussed in the text).
Figure 3.10: Trace-normalized band-pass filtered (1-15 Hz) record section for shot-point TAN recorded to the east along the cross-rift line D. $V_p$=6 km/s (ray-traced phase notation: see Fig. 3.9 caption).
c: is identified from 70 to 100 km to the west, and 100 to 190 km to the east.

d: this is only a weak first arrival visible from 160 to 200 km to the west.

x: it has been identified from about 125 km to 165 km to the east.

BAS - Fig. 3.11, 3.12

Distances are measured from the borehole shot point BAS in Barsalinge.

a: a clear first arrival is picked to 130 km to the west; the apparent velocity being 5.9 km/s, and involving an advance between c.55 and 75 km. To the east a is clear to the end of the section with an apparent velocity of 6.1 km/s.

b: it has been picked from 130 to 180 km to the west.

b1: is identified from c.60 to 120 km to the west, from c.40 to 90 km to the east.

c: the critical point is at c.85 km to the west, and can be picked to c.270 km, although the data are noisy in the portion considered and especially beyond 200 km. Some precritical energy is seen to the east from 90 km to the end of the section.

d: this weak first arrival is visible to the west from 180 to 240 km, though less clear beyond 210 km.

CHF - Fig. 3.13

Distances are measured from the shot point CHF near Chanlers Falls.

a: this is a clear first arrival to 50 km, the end of the section, with 5.8 km/s apparent velocity.

b1: there is a second arrival visible from 10 to 30 km; the ray-traced theoretical branch arise from an interface constrained by the flank line interpretation.

The phase analysis is summarised in Table 3.1.
Figure 3.11: Trace-normalized band-pass filtered (1-20 Hz) record section for shot-point BAS recorded to the west along the cross-rift line D. $V_r=6$ km/s (ray-traced phase notation for Fig. 3.11, 3.12: $a=P_g$, $b_1=P_{11}P$, $c=P_mP$, $d=P_n$). Note: dashed line identifies observed but not ray traced phase $a$. 
Figure 3.12: Trace-normalized band-pass filtered (1-20 Hz) record section for shot-point BAS recorded to the east along the cross-rift line D. $V_r=6$ km/s (ray-traced phase notation: see Fig. 3.11 caption).
Figure 3.13: Trace-normalized band-pass filtered (1-25 Hz) record section for shot-point CHF recorded to the west along the cross-rift line D. $V_r=6$ km/s (ray-traced phase notation: $a=P_g$, $b_1=P_{11}P$).

<table>
<thead>
<tr>
<th>S-P</th>
<th>VIC</th>
<th>KAP</th>
<th>BAR</th>
<th>TAN</th>
<th>BAS</th>
<th>CHF</th>
</tr>
</thead>
<tbody>
<tr>
<td>phase</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>b</td>
<td>-</td>
<td>-/10-100/-</td>
<td>-/-</td>
<td>-/-</td>
<td>-/-</td>
<td>-/-</td>
</tr>
<tr>
<td>$b_1$</td>
<td>50-150</td>
<td>60-150/-</td>
<td>140-230/70-220</td>
<td>105-180/50-140</td>
<td>60-120/40-90</td>
<td>10-30</td>
</tr>
<tr>
<td>$b_2$</td>
<td>100-180</td>
<td>-/-</td>
<td>-/-</td>
<td>-/-</td>
<td>-/-</td>
<td>-/-</td>
</tr>
<tr>
<td>c</td>
<td>80-200</td>
<td>105-140/110-200</td>
<td>70-140/80-220</td>
<td>70-100/100-190</td>
<td>85-270/90-105</td>
<td>-</td>
</tr>
<tr>
<td>d</td>
<td>190-450</td>
<td>-/-</td>
<td>190-235/190-220</td>
<td>160-200/100-190</td>
<td>180-210/-</td>
<td>-</td>
</tr>
<tr>
<td>$d_1$</td>
<td>270-350</td>
<td>-/-</td>
<td>-/-</td>
<td>-/-</td>
<td>-/-</td>
<td>-</td>
</tr>
<tr>
<td>$x$</td>
<td>-</td>
<td>-/-</td>
<td>130-160/130-160</td>
<td>-/-</td>
<td>-/-</td>
<td>-</td>
</tr>
</tbody>
</table>

Table 3.1: Phase analysis results; the numbers in the columns are the ranges in km where each phase is observed.
3.4 Seismic Modelling

The travel time branches on the picked record sections have been used to derive 1-D velocity depth models underneath the shot points. It was evident that large lateral variations in velocity occur, especially in the area surrounding the Rift. A considerable effort was therefore put into defining a starting model before the 2-D ray-tracing was undertaken.

Constraints on the model were derived from previous seismic surveys, from the interpretation of the other KRISP 90 seismic lines and from the known surface geology.

The upper crust starting model was derived from Plus-Minus (Hagedorn, 1959) and Time-Term (Willmore and Bancroft, 1960; Berry and West, 1966; Bamford, 1972) analyses. For the crustal phases a satisfactory fit between the observed and calculated arrival times was taken to be less than 0.1 s. A top-down approach to modelling was undertaken i.e. fitting the near surface data prior to modelling the deeper phases.

3.4.1 Methods of Kinematic Modelling

Forward modelling of the travel times can be performed using various approaches, according to the type of data set and the model to be derived.

The information from a single shot into a line of receivers enables the construction of a 1-D velocity-depth function. An exact solution for the inversion of travel times into a velocity-depth function has been provided in the Wiechert-Herglotz equation, initially described by Herglotz (1907), Bateman (1910) and Wiechert (1910). An additional term taking into account the presence of low velocity zones was added by Gerver and Markushevich (1966, 1967), but it cannot overcome the indeterminacy in the solution if a low velocity zone is present.

If the function $x$, $V_p$ is incomplete, due to the presence of low velocity channels, too short profiles or uncertainties in correlation, the solution has to be approximated.

Giese (1966) proposed an approximation method enabling the calculation of the velocity depth functions from any given travel time curve system. Lateral homogeneity is assumed and finite velocity gradients are taken into account.

The method proposed by Giese (1966) has been implemented in the routines used
here, and examples of its application can be found in Prodehl (1970a, b). The solution is relatively stable when the gradient is strong, because the error in the depth determination is dependent on the gradient at the maximum penetration point of the corresponding ray (Giese, 1976). The method is suited for either positive or negative velocity gradients, and transitional velocity discontinuities rather than sharp interfaces. Isovelocity cross-sections instead of layered cross-sections result from the application of this method (Prodehl, 1979).

Combining information from more than one shot enables the application of delay-time methods (e.g., the Plus-Minus for two reversed shots and the Time-Term for a multiplicity of shots) to determine depth variations on a given refractor. These methods generally require a uniform velocity of the refractor and a topography that is not too severe (Hagedorn, 1959; Willmore and Bancroft, 1960; Berry and West, 1966; Bamford, 1972). Extension of the method can accommodate, if required, anisotropic velocity functions of the refractor (Raitt et al., 1969), lateral variations of the refractor velocity and vertical velocity gradients within the refractor (Smith et al., 1966). Another technique for interpreting in-line forward and reverse seismic refraction data is the Generalised Reciprocal Method (GRM) (Palmer, 1980, 1981). The calculation of the velocity analysis function and time-depths are similar to those for the conventional reciprocal method, mathematically analogous to the Plus-Minus method, but the migration principle is used. The refractor topography is therefore better defined combining rays leaving the refractor at the same position rather than arriving at the same detector as in the delay-time methods (Palmer, 1986).

Complex 2-D layered structures can be modelled with ray methods. Ray-tracing determines the paths along which the energy travels computing the travel time associated with those paths. The basis is Fermat's principle that a seismic raypath between point A and point B is that for which the first-order variation of travel time with respect to a neighbouring path is zero (Sheiff, 1991). In other words, seismic energy travels between two points along the path that makes the travel time stationary compared with neighbouring paths. In a homogeneous region the ray path is a straight line that bends at interfaces according to Snell's law. A series of ordinary differential equations must be satisfied by the ray path in inhomogeneous regions, according to Fermat's principle (Červený et al., 1977; Aki and Richards, 1980). In an inhomogeneous medium the path will not be straight, but
The ray-tracing method is particularly apt for the modelling of horizontal and vertical velocity gradients within layers, highly irregular or steeply dipping refractor interfaces and discontinuous layers. The ray method, also known as the asymptotic ray theory method, involves numerical modelling of high frequency seismic wave fields. The wave field is composed of the weighted sum of elementary waves (reflected, refracted, multiply reflected, converted, etc.). The elementary waves are evaluated along individual rays. It is a very powerful representative of high frequency asymptotic methods. However, it is only approximate in the sense that only the first term expansion of the wave equation is considered. It can be applied only to smooth media, and its accuracy is low in singular regions, such as critical regions and caustics. Moreover, the ray method does not include some non-ray waves; in other words, asymptotic ray theory fails when considering seismic energy travelling along non-Fermat paths, such as inhomogeneous waves, channel waves, tunnel waves, diffracted waves, etc. (Červený et al., 1977). It is nevertheless of great advantage for its universality, effectivity and conceptual clarity (Červený, 1985).

1-D Model - Results

The derived apparent velocities and intercept times were used to determine plane layer 1-D velocity-depth functions at the shot points.

The velocity-depth functions including possible low velocity zones and increasing velocity (positive gradient) layers were derived following the method described by Prodehl (1979).

The functions shown are an output of the MacR1D interactive program for calculating travel time curves from one-dimensional velocity-depth functions, written by J. Luetgert (1992b) for a Macintosh IICi (Fig. 3.14 and 3.15).

The best fitting functions are drawn assuming the simplest possible model: horizontal layering and sharp velocity contrasts (increases) with depth instead of high velocity gradients; modification of the gradients may be considered after amplitude modelling. No low velocity zones were modelled.

The functions reach mantle depth in most cases. One or more layers with velocity...
CHAPTER 3. KINEMATIC MODELLING OF P-WAVES

Figure 3.14: 1-D velocity depth functions for shot VIC, KAP and BAR. A=to the west, B=to the east.
Figure 3.15: 1-D velocity depth functions for shot TAN and BAS. A=to the west, B=to the east.
below 6 km/s at the surface are identified for all shots except BAS and CHF. One and sometimes two (VIC) intracrustal reflections are modelled between 14 and 28 km; the depth to the Moho is c.36-39 km for the shots on the rift flanks, and less than 30 km for the shots within the rift (BAR and TAN to the west). The S-wave velocities are calculated assuming a ratio \( V_p/V_s = 1.732 \).

**Plus-Minus Analysis - Results**

The basement refraction, being well recorded, from all the shot points, made it a suitable candidate for Plus-Minus analysis (Hagedorn, 1959).

Plus-Minus is considered a wavefront method, because the refractor is graphically reconstructed at the intersection of two wavefronts directed towards the shot points.

Considering two shots at either end of a recording spread, the intercept time at a receiver from each shot may be considered to be derived from two delay-times, one associated with each shot direction.

The separation of the two delay-times enables the calculation of the local depth of an irregular refractor. All detector positions where a head wave arrival from the basement was recognised from both ends of the profile line can be associated with a depth to the refractor. That portion of the profile included between the two sediment/basement arrival cross-over distances was therefore used.

Using the observed station ranges and travel times the depth of the basement beneath the surface was determined using the Plus-Minus method.

Three main sections were studied (Fig. 3.16):

1- between VIC and KAP (c.80 to 150 km along the line), the basement has a depth of 0-0.5 km in the west, to c.1.0 km at c.130 km, where volcanics cover the basement.

2- between KAP and BAR (c.185 to 230 km), the depth goes from 3 km to more than 4 km, becoming shallower near BAR. A short section between BAR and TAN provides a depth of c.2.5 km.

3- between TAN and BAS (c.275 to 340 km), the basement progressively shallows from 0.9 km to 0.2 km near BAS.

Different depths for the same portion of basement are considered to result from the
CHAPTER 3. KINEMATIC MODELLING OF P-WAVES

Figure 3.16: Plus-minus depth determined from the $a$ phase. Different lines result from the combination of different data sets.

choice of different shot combinations. These discrepancies are due to the uncertainties in velocity assumed for the top layer and in the reciprocal time. Difficulties can arise when the velocity of the sediment cover is not well defined, and when the travel time branch is irregularly shaped reducing the significance of the extrapolation to zero offset providing the reciprocal time.

Moreover, the Plus-Minus method is applicable only for dips of less than 10°, where the offset distances at the surface can be approximated to the refractor distances at depth, as required by the method. In the region surrounding the Rift the tectonic history suggests the basement might involve steeper dips than 10° or be faulted. The Plus-Minus analysis was therefore used outside the rift, or to give indications of the structural pattern inside the rift, as well as the order of magnitude of depths to be considered. Overall, the method provided a first estimate of the basement top depth variation along the profile.

Time-term Analysis - Results

The good quality of the $a$ phase and the high ratio of number of observations to numbers of unknowns (velocity and time-terms) allowed the application of the time-term method. This is described in detail in Willmore and Bancroft (1960), Berry and West (1966), Bamford (1972) and Whitcombe and Maguire (1979).

Time-term analysis is a delay-time method to evaluate seismic refraction travel time data from arbitrarily distributed shots and receivers.
From a system of \( m \) stations and \( n \) shots, \( m + n \) time-terms can be obtained.

The number of observations being greater than the number of unknowns enables a least square solution of the time-terms and velocity to be obtained. The solution of a set of normal equations relating the propagation time to the time-terms and the travel time along the refractor, can determine the time-terms.

For the time-term method at least one shot and receiver site must be coincident or one or more stations must lie on the refractor. The method gives better results if each receiving station records several shots and also if the shot points record other shots to provide reciprocal times.

The method is applicable only to true 'refracted' arrivals. No information on the internal structure above the refractor is provided (Giese, 1976).

Assuming a constant velocity in the refractor, and another constant velocity in the top layer, a first iteration was attempted.

The high values of variance obtained suggested the data should be split into smaller sections along the line, each section being of supposed less variable velocity. The observational error was estimated at \( \leq 0.1 \) s, and from an examination of the picks the observational error variance estimate (Whitcombe and Maguire, 1979) was estimated at about a tenth of the error. A value of solution variance was therefore accepted when it was \( \leq 0.01 \) s².

After various attempts with different sections and different numbers of data sets included in the inversion, the line was cut at the basement-volcanic contact (130 km from VIC), at the Elgeyo escarpment (180 km from VIC), at the sediment-volcanic contact east of Lake Baringo (240 km from VIC) and at the volcanic-basement contact near Barsalinge (350 km from VIC).

Only those stations with a minimum of two observations were selected. The results are shown in Fig. 3.17. The basement appears at a constant depth up to the Elgeyo fault, where it drops to stay at around 5 km depth before shallowing again, with a step-like structure beneath the eastern shoulder of the Rift. The velocity underneath the refractor shows strong lateral variations.

The most problematic section is below the Usain Gihu Plateau, giving relatively high
Figure 3.17: Time-term depths for the $a$ phase. Numbers give velocities in km/s. Solution variances are given in brackets.

$V_p$ values, or large depths to the top of the basement, depending on the velocity above the refractor.

This is most probably due to the abrupt lateral changes beneath the west shoulder of the rift. In Fig. 3.17 a low velocity (6.01 km/s) and large depth has been displayed. Nevertheless study of the surface geology suggests the basement is outcropping. Therefore at the eastern end of the Uasin Gishu plateau the results of the time-terms have been constrained to a zero depth to the basement, and the cover velocity chosen was 3.9 km/s.

The final depth converted time-terms are presented in Appendix B. The topography of the basement delineated by this analysis was used as a starting point for the 2-D modelling.

Two factors resulted in limited confidence in the results: (1) the uncertainty in the end-to-end time because it is extrapolated to the full range and (2) the abrupt changes in the refractor topography, resulting in complex ray paths which traversed zones of sharp lateral and vertical velocity gradient. Moreover there is uncertainty introduced by the unknown velocity and thickness of the near surface layers.
Ray-tracing - Results

The evaluation of the seismic wave field of individual elementary waves consists basically in ray-tracing. Depending on the seismological problem treated, ray-tracing algorithms can be fast but not accurate in the evaluation of ray amplitudes and ray synthetic seismograms. They can be time consuming but more accurate when completing the numerical integration of the ray-tracing system, composed of the ordinary differential equations of rays.

The evaluation of geometric spreading is the most important part of the vectorial complex-valued amplitude calculation of elementary waves; the evaluation of amplitudes will be considered in Chapter 5.

The 2-D forward modelling was performed using a standard ray-tracing routine (Červený et al., 1977), MacRay, the version for a Macintosh IIci of the USGS RAY86/R86PLT (Luetgert, 1992b). It shoots rays through a 2-D model, and can calculate the complex amplitude and geometric spreading of these rays. In the manual mode the rays are shot by boundary value ray-tracing (two-point ray-tracing). In the automatic mode, the boundary and the wave types are specified and ray-tracing is performed with an automatic search of the wave selected within the model lateral boundaries. The shooting angle increment and the time step must be defined by the user. Direct, precritical, refracted and reflected rays can be automatically searched. The program will first search for the critical angle in the ray beam defined by the user, but in certain situations (LVZ, complex structures) no critical angle can be found.

The model is defined by a number of interfaces extending from the left to the right margin; the interfaces may 'pinch out' but not cross. Any pair of successive interfaces describes a layer, within which the velocity may be defined at the top and at the bottom for different vertical speednet lines. The velocity within any layer may be inhomogeneous, but continuous. The program uses a bilinear velocity interpolation procedure. The velocity at any point is given by a combination of the four adjacent velocity nodes. First order changes in velocity and gradient across the interface provide step velocity discontinuities. Second order changes involve velocity gradient changes as gradational velocity discontinuities.

The ray-tracing algorithm (Červený et al., 1977) calculates the propagation of rays within a layer by the stepwise integration of a system of first order differential equations,
as a function of the Cartesian coordinates, the time and the angle from the vertical. They may be written (Červený et al., 1977):

\[
\begin{align*}
\frac{dx}{dx} &= \cot \theta \\
\frac{dx}{dz} &= \tan \theta \\
\frac{d\theta}{dx} &= \frac{V_s - V_g \cot \theta}{V} \\
\frac{d\theta}{dz} &= \frac{V_s \tan \theta - V_g}{V},
\end{align*}
\]  

where: (3.1) is for near-horizontal rays and (3.2) is for near-vertical rays; the initial conditions are: \(x=x_0, z=z_0\) (source location), \(\theta=\theta_0\) (ray take-off angle). \(\theta\) is the off vertical ray-tangent, \(V\) the velocity, and \(V_s\) and \(V_g\) partial velocity derivatives with respect to \(z\) and \(x\).

Equations (1) and (2) are solved with the Runge-Kutta method (Červený et al., 1977) and the ray propagates through iteration of the above. When an interface is encountered in the calculation of a ray, Snell's law is applied and the calculation is continued. The graphical output of MacRay consists of a plot of the model including the ray paths and a travel time plot with optional calculated and observed arrival times.

The starting model was constrained by surface geological observations and by the results from a previous more detailed seismic experiment (Swain et al., 1981). The latter was particularly useful in the interpretation of the rift infill structure. It involved a 50 km east-west line between Lake Baringo and the Kerio River (Fig. 1.24a), with vertical seismometers at 0.8 km intervals and with four shot points. The velocity-depth functions beneath BAR and CHF derived from the interpretation of the axial line (Mechie et al., 1994b) and flank line (Prodehl et al., 1994a) respectively were also taken into account to produce a consistent 3-D picture.

The starting model was modified to match the observed travel times of all phases from all shots as closely as possible with the calculated times of the rays shot through the model.

The record sections were plotted using the true distance of shot to receiver and the distances along the line were calculated with the averaging procedure mentioned in Section 2.5.2: this can introduce discrepancies between calculated and observed times, but the resulting errors have been considered negligible with respect to the resolution given by the
interpretation procedure. The error decreases with the shot-receiver offset, and falls off as \( \sin \alpha \), where \( \alpha \) is the angle between the seismic line and the averaged line. The resolution given by the interpretation procedure will be discussed in detail in Chapter 7.

**Phase Analysis**

Figures 3.18-3.20 show the final travel time plots following the iterative 2-D ray-trace modelling: the fit between the observed and the calculated travel time is better than 0.1 s for the crustal phases shown in the diagrams. For the mantle phases the fit is generally c.0.2-0.3 s. The detailed analysis of the mantle phases from VIC (see Chapter 6) improves the fit to 0.1 s. All the distances are given measured from the western end of the profile and the T-axis is displayed as reduced time.

**Upper crust - (Fig. 3.18)**

**Phase \( a \)**

The \( a \) phase near the VIC shot point (Fig. 3.2) requires the presence of some low velocity material (at 3.0 km/s) to explain the c.0.5 s reduced arrival time at about 5.0 km distance. There is some evidence of sedimentary cover underlying the shot point as suggested by multichannel profiling in Lake Victoria (Rach, 1985). Gravity surveys in the area have also identified low density bodies interpreted as granitic intrusions within the Nyanza Rift, where the shot was fired. Seismic velocity laboratory measurements of saturated granite provided values of about 5.0 km/s (Carmichael, 1982), higher than the velocity required at shot point VIC. 1.4 km of sediments were included in the model.

The direct arrival on the KAP record section (Fig. 3.5, 3.6) provides velocities typical of volcanic rocks (4.2 km/s). Geological observations however do not suggest the thickness obtained underneath the shot point, which is necessary to explain the delay in the following phase \( a \). About 0.5 km of volcanic rocks are observed on the Elgeyo escarpment, at 180 km along the line (Fig. 3.18) covering the basement rocks. Up to 1.7 km of volcanic cover has been modelled under KAP (Fig. 3.18), with a velocity of c.4.1 km/s.

Within the Rift Valley, modelling of the BAR shot data (Fig. 3.7, 3.8) required a layered structure in the rift infill from 180 to 330 km along the line (Fig. 3.18). An uppermost layer, of c.2.2 km/s velocity from 210 to 240 km, has been attributed to sedimentary
Figure 3.18: \(a_s\) and \(a\) phases ray-traced final model (for velocity distribution see Fig. 3.23. Number of rays for each phase limited to a maximum of 20 for display). Vertical exaggeration 10:1.
(lacustrine) deposits, with a thickness of up to 1.0 km. The deeper subdivision into two layers of 3.9 km/s and 5.5 km/s has been constrained by the Swain et al. (1981) model, although in their interpretation the deepest of the two layers is suggested as representing the basement (see Phase a Section and Fig. 1.24a).

TAN was drilled in compacted sand (Jacob et al., 1994) the resulting data providing $V_p$ velocity of 2.9 km/s and a thickness of up to 0.3 km for the deposit underlying the shot point (Fig. 3.9, 3.10). As shown in Fig. 3.18 this phase has been ray traced up to 10 km in both directions. The following phases are again attributed to two layers in the rift infill.

Phase a

On the VIC record section phase a is delayed up to 0.5 s out to 30 km from the shot point (Fig. 3.2). There is no evidence to extend the sedimentary cover to such a distance: the stations are on land, and ANC intrusives are mapped in this area; the most reasonable means of fitting the data is to decrease the basement velocity and introduce a high velocity block at depth to explain the normal (6.0 km/s) velocity beyond 30 km. At about 110 km the contact between the ANC and MOB is marked by a slight increase in $V_p$. Phase a is ray-traced to 150 km from VIC (Fig. 3.18).

The reversed branch from KAP shows the same delay in arrival times near VIC, up to where it has been ray-traced (Fig. 3.5). The calculated arrival is about 0.25 s earlier than the observed. A gradient of c.0.01 s$^{-1}$ is defined for the basement rocks between 50-150 km along the profile, somewhat less in the first 50 km.

Arrivals from KAP to the east are influenced by the strong lateral variation in structure (Fig. 3.6); nevertheless a has been modelled up to 310 km along the line (Fig. 3.18). A $V_p$ velocity of 6.3 km/s is necessary to compensate for the delay in a where the surface observation shows that MOB rocks outcrop on the Elgeyo escarpment. A delay of up to 0.7 s appears due to displacement of the basement by the Elgeyo fault, although the identification is difficult due to the low signal to noise ratio in the rift. At the base of the scarp, in the Kerio Valley, a further delay is added, providing a reduced arrival time for a of 0.9 s. A $V_p$ of 6.2 km/s with a gradient 0.02 s$^{-1}$ is modelled under the rift.

BAR to the west is delayed up to 1.5 s due to the rays travelling through the rift (Fig.
The delay progressively decreases as the rays pass beneath the rift shoulder, where they are traced to 90 km from VIC. The calculated time is 0.15 s earlier than the observed in the range 115-145 km. To the east, a similar 1.5 s delay is observed in the travel time, decreasing towards the eastern shoulder (Fig. 3.8). A $V_p$ velocity of 6.0 km/s with a gradient of 0.02 s$^{-1}$ is modelled.

Arrivals from TAN have a progressively increased delay to 1.5 s entering the rift to the west (Fig. 3.9). They have been ray traced to 130 km, successfully fitting the lateral changes across the Elgeyo Fault. The signal is clearer to the east, where a delay of up to 1.1 s appears due to the downfaulting of the rift's eastern shoulder (Fig. 3.10). It is not possible to trace phase $a$ beyond 340 km due to the presence of the high velocity upper crustal block.

A high signal to noise ratio (Fig. 3.11) allows phase $a$ to be traced to the west of BAS where the advance between c.270 and 295 km has been modelled introducing a high velocity (6.3 km/s) block at about 1.5 km depth and c.4.0 km thick. There is no constraint on its base but arrivals undershooting the block from BAR and TAN require a normal basement velocity below 5.5 km. Phase $a$ can only be traced to 260 km along the line, with a gradient of 0.02 s$^{-1}$. A $V_p$ velocity of c.5.8 km/s, with a 0.02 s$^{-1}$ gradient fits the data to the east of BAS to 430 km (Fig. 3.12), and to the west of the reversed shot CHF to 400 km from VIC (Fig. 3.13).

**Lower crust - (Fig. 3.19)**

Phase $b$

Lower crustal diving waves observed from BAR in both directions (Fig. 3.7, 3.8), from TAN to the east (Fig. 3.10) and BAS to the west (Fig. 3.11), allow a $V_p$ of 0.5 km/s to the west of the rift and a marginally lower velocity to be modelled to the east, with a gradient of 0.01 s$^{-1}$.

Rays from BAR to the west fit the data to 40 km along the line. To the east the high velocity block in the upper crust does not allow rays to be traced between 340 and 360 km. From TAN to the east no arrivals can be modelled between 340 and 400 km. Diving waves from BAS to the west pass through the high velocity block and can be ray traced to 250 km. In summary, ray-tracing constrains the mid-crustal velocity field between 110
Figure 3.19: $b$, $b_1$ and $b_2$ phases ray traced final model (for velocity distribution see Fig. 3.22. Number of rays for each phase limited to a maximum of 20 for display). Vertical exaggeration 5:1.
to 220 km, and 280 to 320 km, as shown in Fig. 3.19.

Phase \( b_1 \)

From VIC \( b_1 \) can be ray traced back to 40 km, where precritical energy is seen on the record section (Fig. 3.3), although the modelled critical point is at about 75 km.

The reversed phase from the KAP shot point (Fig. 3.5) requires a step like structure in the reflector, deepening by 3.5 km at about 95 km from VIC.

Phase \( b_1 \) on the BAR record section did not constrain the reflector, although rays traced through the trial model, fit with the high amplitude diffuse energy which follows 0.2 - 0.3 s behind phase \( a \) on the BAR record section in both directions (Fig. 3.7, 3.8).

Under the rift axis the depth of the first intracrustal reflector is constrained by the axial line interpretation (Mechie et al., 1994b), although some rays are reflected from the TAN shot to the west from this interface. To the east rays do emerge from the base of the upper crust to explain the diffuse energy following phase \( a \) thus identified as a mid-crustal reflection (Fig. 3.10).

Some energy is seen to the west of BAS in the noisy data (Fig. 3.11), and it has been ray traced. To the east a reflector at a shallower depth than to the west is necessary to explain the high amplitude phase (Fig. 3.12), consistent with the same phase identified on the record section from the shot CHF to the east (Fig. 3.13): a rising reflector is required from 390 km onwards.

Phase \( b_2 \)

This is only seen from VIC (Fig. 3.3). \( b_2 \) is less energetic then \( b_1 \), and the phase correlation is possible only over a distance range of about 90 km. It is modelled with a critical point at 135 km. The ray coverage is limited to a portion between 60 to 100 km along the line (Fig. 3.19). Under the rift axis the axial line interpretation indicates a depth of c.23 km for the lower crustal reflector, although it does not necessarily result from the same geological interface.

**Upper mantle - (Fig. 3.20)**

Phase \( c \)

On the VIC record section phase \( c \) is modelled from 95 km, where the critical point is raytraced, and can be followed to 210 km (Fig. 3.3).
Figure 3.20: \( c \) and \( d \) phases ray traced final model (for velocity distribution see Fig. 3.22. Number of rays for each phase limited to a maximum of 20 for display). Vertical exaggeration 5:1.
CHAPTER 3. KINEMATIC MODELLING OF P-WAVES

As for $b_1$, the reversed shot KAP to the west requires the Moho to deepen to the east, although the rays traced from this interface model the critical point at a distance from KAP where no energy is clearly seen on the record section (Fig. 3.5). The same happens to the east, where the critical point from the Moho should occur at about 260 km along the line. It is at this range that the rift infill introduces strong reverberations in the signal (Fig. 3.6). Beyond the rift margin the modelled travel times fit the observed to within 0.3-0.4 s.

Mantle phases from BAR have been identified/enhanced in the combined plot (Fig. 3.21) where BAX1, BAX2, BAX3, BAR4 are displayed together. Phase c to the west requires the Moho to shallow under the western margin of the rift, between 280 and 340 km, consistent with rays traced from VIC through the same region. To the east rays are reflected from outside the margin, where the Moho deepens again.

Rays modelled from TAN fit with the diffuse phase c identified in both directions (Fig. 3.9, 3.10).

Phase c from BAS to the west is ray traced with a critical point at 220 km (Fig. 3.11). To the east precritical rays fit with the energy picked at c.2.5 s reduced time (Fig. 3.12), constraining the Moho depth for few km around 400 km (Fig. 3.20).

Phase $d$

The best diving waves from the uppermost mantle, phase $d$ are picked on VIC (Fig. 3.4), and significantly constrain the upper mantle velocity distribution. A gradient of c.0.01 $s^{-1}$ is required to generate turning rays from the upper mantle into the crust beneath the rift axis. A higher gradient would be necessary in the section immediately under the Moho between 170 and 200 km, where no rays appear. A gradient of 0.01 $s^{-1}$ has been kept in this region. This determines the apparent lack of fit in the observed data between 230 and 270 km; the resulting travel time calculated branch is straight where there is an advance in the observed arrival times.

From KAP to the east phase $d$ arrivals (Fig. 3.6) are fitted by rays turning from the step-like structure under the eastern shoulder (Fig. 3.20); a slightly higher gradient of 0.015 $s^{-1}$ is modelled in this region.

It is not possible to trace rays to the surface from BAR to the west due to limitations
Figure 3.21: Merged, trace-normalized band-pass filtered (5-15 Hz) record section for shots BAR4, BAX1, BAX2 and BAX3 recorded to the west along the cross-rift line D. $V_r=6$ km/s (ray-traced phase notation: see Fig. 3.7 caption).
CHAPTER 3. KINEMATIC MODELLING OF P WAVES

in the ray-tracing procedure: the observed arrivals may be due to a refraction along the Moho or diffraction (Fig. 3.7; Maguire et al., 1994). To the east diving waves can be successfully traced to the end of the profile, as seen in Fig. 3.20.

Calculated times to the west of TAN fit with the observed times in a small region at c.100 km along the line.

From BAS to the west (Fig. 3.11) phase $d$ is fitted with rays travelling through the upper mantle emerging under the rift, reversing the rays shot from VIC between 170 and 270 km on the profile.

Phase $d_1$

The $d_1$ phase from VIC (Fig. 3.4) has been modelled by introducing a reflector at about 55 km depth. This phase is the subject of more detailed processing and interpretation in Chapter 6.

Phase $x$

As suggested by Maguire et al. (1994) this phase identified on the BAR record sections in both directions (Fig. 3.7, 3.8) and on the TAN record section to the east (Fig. 3.10) may be a precritical phase from an upper mantle reflection, or a converted or a multiple phase. It is not possible to decide between these options solely by travel time analysis. It will be seen in Section 5.3.2 how dynamic modelling aids in the identification of this phase as a multiple reflected within the rift infill.

3.4.2 Model Description

The best fit seismic model in Fig. 3.22 shows an east-west section across the Rift to a depth of 60 km, with the distribution of the $V_p$ velocities. The values of gradients used in the modelling are reported. The significance of the gradient values will be considered during dynamic modelling (Chapter 5); at this stage up to ±100% uncertainty can be supposed. Seismic amplitudes are a function of the velocity gradients and discontinuities. These latter can therefore be examined through detailed synthetic seismogram modelling. As suggested by Mooney (1989), for high quality data the accuracy would be ±50%.

The model resolution will be discussed in detail in Section 7.2. An inversion study of the seismic travel times, initially constrained by the present model enables a quantification
KRISP line D - Cross Rift profile

Figure 3.22: Final 2-D ray-trace model for KRISP 90 cross-rift line D. Dashed lines indicate transitions where there is a change in velocity gradient but no significant discontinuity. P-wave velocities in km/s, gradients in $s^{-1}$ in italic.
of the spatial resolution for the structural nodes.

In forward modelling resolution and accuracy are a function of the subjectivity of the phase correlation procedure, as mentioned above. For this reason, an iterative approach involving modelling and checking of phase correlation was necessary. Resolution of the final model is also dependent on the ray coverage through the model. To estimate the uniqueness of the model parameterization a series of perturbations of individual values of depth or velocity have been applied to the final model. The modified model must give a travel time fit within the limits specified in Section 3.4.1, must allow rays to propagate and must have realistic physical and geologic properties. Model perturbations resulted in velocity and interface depth resolution estimates of 2-3% and 7% respectively. The least well constrained zones are confined to the sides of the model, the deepest part being well resolved in the centre. Also resolution is low in zones of particular structural complexity where the ray-tracing is difficult, for example beneath the Kerio Valley, or through the upper crustal high velocity block beneath the eastern flank of the Rift.

The main features outlined by this equatorial cross-rift seismic profile are:

- an asymmetric sedimentary basin within the Rift which is thickest against the major rift boundary fault to the west,
- the transverse asymmetry of the Rift, defining a half graben shape,
- a distinct difference between the upper and lower crust velocities,
- the variation in thickness of the crust, being greater under the rift flank than under the rift axis,
- and the presence of an intra-mantle boundary.

The velocity distribution shows lateral changes, especially in the upper part of the model, and velocity gradients decreasing with depth in the crust. The uppermost mantle velocity is anomalously low beneath the rift.

**Upper crust**

From west to east, the description of the supra-crustal velocities has been geologically interpreted as follows (Fig. 3.23, to 18 km depth).

At VIC there is no evidence of thick surficial sediments. Results from an analysis of eight shot-gathers distributed over the whole of Lake Victoria showed an average sediment
Figure 3.23: Upper crustal final 2-D ray-trace model for KRISP 90 cross-rift line D. P-wave velocities in km/s, gradients in $s^{-1}$ in italic.
CHAPTER 3. KINEMATIC MODELLING OF P-WAVES

thickness of 167 m (Rach, 1985). The basin modelled with a depth of 1.4 km nearby the shot point could be related to the northern margin of the Nyanza Rift.

At the base of the Elgeyo escarpment, and in the Kerio Valley, there is asymmetric deposition associated with post-faulting erosion and the fluvial drainage. The present results suggest a recent alluvial sediment thickness of up to 0.5 km.

The sedimentary basin surrounding Lake Baringo shows a greater thickness to the west. The velocity increases to the east from 2.2 km/s to 2.4 km/s. A maximum sediment thickness of about 1.0 km is in agreement with the model of Swain et al. (1981) and with the axial line, where the velocities are slightly higher (c.2.4 km/s). The maximum exposed thickness of the Baringo Beds is about 0.2 km (Hackman, 1988). Recent magnetotelluric data (C.K. Morley, pers. comm., 1994) suggests up to 1.0 km thickness of sediments around Lake Baringo. The observed structure could be due to the asymmetric development of the Baringo-Hannington Lake Beds during the Plio-Pleistocene faulting episodes. Outcrops along the western side of the Marigat plain, surrounding Lake Baringo on the west and south, suggest a ‘pull apart’ basin internal structure, composed of small blocks tilted towards the east (Tiercelin et al., 1987).

A thin sediment outcrop is also modelled at the western scarp of the Laikipia plateau, west of TAN. It may be alluvium or the expression of the Rumuruti Phonolites Lower Tuffs, mapped with a thickness of up to 0.18 km in the Tanganbei area (Carney, 1972). The present model suggests up to 0.3 km.

Upper volcanic unit

At about 130 km along the line from VIC the first rift related volcanics outcrop. The Uasin Gishu phonolite plateau has a fairly flat topography. Geological data suggests a maximum thickness of 0.48 km, increasing towards the east where basement rocks outcrop again in the Elgeyo escarpment (Fig. 2.12). Half a kilometre of phonolites have been modelled as a result of the present interpretation. The seismic model shows a synclinal basal contact. This could be related to the Miocene erosion surface, masked by the deposition of volcanics. Further north along the Elgeyo escarpment up to 0.38 km of a sedimentary formation (Tambach Sediments) lies on the basement (Chapman et al., 1978). North-east of Baringo, on the Pliocene plateau phonolite up to 30 m of Miocene sediments
(Kirimui Formation), underlie the phonolites and fill the depressions in the sub Miocene surface (Hackman, 1988). There is no direct evidence of the seismic velocity at the base of the volcanics under KAP, therefore the presence of sediments can not be established. They may result in a velocity inversion. Nevertheless the shape of the surface could be due to erosion during the relatively wet Miocene period.

The graben infill is proposed as being formed of a layered sequence of Miocene-lower Pliocene basalts, phonolites and trachyte beds (see also Swain et al., 1981). The upper layer velocity varies from 3.8 km/s west of Kamasia, to 4.1 km/s at Kamasia, to about 3.8-3.9 km/s from Baringo onto the eastern shoulder. Laboratory measurements on seismic velocities from volcanic rocks within the Rift show values between 1.9 and 5.15 km/s. The lowest values are associated with pyroclastic products; the lavas give values generally higher than those derived from the seismic results. A basaltic trachyandesite from Baringo has $V_p$ c.5.0 km/s. Fracturing, jointing, pyroclastic intercalation or high pore pressure could all contribute to lowering the velocity of the bulk rock (Mooney and Christensen, 1994 and ref. therein).

Similar phenomena could also explain the lateral velocity variations within the upper volcanic seismic unit, composed of different lithological formations (see Section 2.5.8).

The Kamasia horst has been interpreted as due to basement block tilt during two major phases in the Late Miocene and Plio-Pleistocene (Chapman et al., 1978); MOB lithologies are exposed only 20 km north of the line, overlain by 3.0 km of lavas and sediments, forming the thickest sequence of Miocene rocks in the Kenya rift (Chapman et al., 1978). A c.3.3 km thick layer, with a velocity of 4.1 km/s, was modelled at Kamasia for the upper unit. An alternative model is possible, with a thinner (2.9 km) and lower velocity (3.8 km/s) layer. The alternative model would better agree with Swain et al. (1981) who modelled in the same area a thinner (2.5 km) and slower (3.7 km/s) seismic unit: as commented by Maguire et al. (1994) this discrepancy could be due to the complex ray path geometry in this region.

In the Baringo area velocities and thicknesses agree with the values along the axial line (Mechie et al., 1994b) to within the associated error range.

The volcanic cover is modelled as being consistent with the surface geology for the
Laikipia phonolitic plateau, indicating the presence of Miocene and Pliocene volcanics of the Rumuruti group to 330 km along the line. The thickness of this unit on Laikipia decreases to the east from 1.8 km at 265 km, to 1.2 km at 290 km to 0.2 km at 320 km. It can be correlated with part of the Rumuruti Units, mapped with a total thickness of about 0.6 km.

**Lower volcanic unit**

The lower volcanic seismic unit is modelled with a velocity above 5.0 km/s, decreasing towards the east, where it pinches out at 330 km along the line.

A very deep trough beneath the Kerio Valley is modelled to fit the seismic data and is consistent with the gravity data: its depth is poorly defined owing to the lack of rays emerging as upper crust first arrivals. It has been modelled with a maximum depth of 8.7 km.

Near vertical seismic surveys (Section 1.5.3) identified up to 3.0 km Miocene-Recent infill in the Kerio Valley, which has been interpreted as a graben. A further 2.0-3.0 km of Eocene-Oligocene deposits are inferred in the interpretation of the deep structure across the Kerio Valley (C.J. Ebinger, pers. comm., 1993), not inconsistent with the present results.

In the rift basin the lower volcanic unit has been related to the first volcanic extrusion, the Samburu Basalts, associated with the rifting process. Golden (1978) suggested the total original thickness of the Samburu Basalts to be 1.0-1.2 km north of Tangulbei (Amaya Embayment), and Griffiths (1977) considered the whole sequence to be at least 1.07 km thick.

Swain et al. (1981) interpreted the 5.7-5.8 km/s velocity underneath the rift as resulting from crystalline basement. In the present model and beneath the KRISP 90 axial line (Mechie et al., 1994b), velocities of 6.0 km/s or over are more appropriate for the basement rocks, while volcanic units in the rift infill reach 5.1 km/s.

On the Laikipia plateau the lower volcanic unit could again be correlated with Miocene volcanic products, not visible at the surface along the line but outcropping just north of Tangulbei, stratigraphically below the Rumuruti Group.
CHAPTER 3. KINEMATIC MODELLING OF P-WAVES

Basement

The P-wave velocity of 6.4 km/s to c.50 km along the line from VIC appears to be related to an Archaean Greenstone belt association. The first high amplitude mid-crustal reflection identified from VIC requires a velocity contrast of about 6.2-6.5 km/s: 'normal' upper crustal velocities must therefore underlie the high velocity zone.

To a distance of c.110 km, still in the MOB Western Sector (Mosley, 1993), lower velocities between 6.05 and 6.1 km/s were modelled, possibly due to the presence of an Archaean Granitoid complex. The Nandi Fault at c.110 km, marks the onset of MOB lithologies, or to the MOB Central Sector of Mosley (1993).

From KAP to the Elgeyo escarpment the P-wave velocities are required to be above 6.3 km/s, as discussed in the time-term interpretation results. Velocities around 6.1-6.3 km/s are typical of the basement unit across the rift.

The same values are found in the basement rocks along the flank line, where MOB gneissic lithologies outcrop. The velocity values strongly suggest that gneissic rocks are the dominant rock type also beneath the rift. Laboratory measurements of seismic velocities on the basement metamorphic rocks have been corrected for different geotherms under the flank and the axis of the rift, and show excellent agreement with the seismic velocities. The modelled positive gradient also suggests that the composition becomes more mafic at depth (Mooney and Christensen, 1994). This would agree with the previously hypothesized presence of a high density crustal zone underneath the rift axis (Swain, 1992).

On the rift's eastern margin velocities are about 6.0 km/s to the end of the line. A high velocity block has been introduced in the model and could be related to an Archaean high grade metamorphic remnant reworked in the Mozambique orogeny (Mosley, 1993); migmatitic associations outcrop south of the line (Mukogodo Migmatite). These migmatites are exposed between 37° – 37.3°E, within the Barsalolian tectonic domain. Exposures are confined to tectonic windows, as they are tectonically below Mozambiquan rocks (Hackman et al., 1989). An intrusion of anorthositic gabbro has been identified at the southern end of the flank line near CHF, with a velocity of 6.45 km/s at a depth of 2.0 km (Prodehl et al., 1994a).

At the eastern end of the line the velocity of 6.2 km/s is consistent with the KRISP 90
CHAPTER 3. KINEMATIC MODELLING OF P-WAVES

flank line interpretation (Prodehl et al., 1994a); the metamorphic grade is higher in the east as shown by the kyanite/sillimanite isograd in Fig. 1.9.

The overall picture seems to show a thinning of the upper crust beneath the axis caused by the rifting process, bordered by a thickening under the shoulders, and a progressive thinning away from the flanks. The thinning at the western end of the profile could result in part from the presence of the Nyanza Rift. The thickening at depth where ANC is overthrust by MOB could be the suture signature of the orogeny. At the eastern end of the profile the NW-SE orientated Cretaceous-Early Tertiary Anza Rift could have provided residual thinning (see location in Fig. 1.9), if, as Bosworth (1992) noted, the Anza Rift is 'one of the world’s largest single rift basin, with a width over 100 km along much of its known 500 km length'.

It is worth noting how the rift axis seems to be superimposed where the crust was thickest: this is possibly related to the Kusznir and Park’s (1987) or Dunbar and Sawyer’s (1988) models of rifting in relation to decreased lithospheric strength at zones of thickened crust. An alternative model proposes the thickening is due to lateral intrusion of mafic material at the base of the crust (Braile et al., 1994).

Lower crust

The upper boundary of the mid-crustal layer is well defined. The step of this boundary under the western rift shoulder seems to reflect the passage from Archaean Craton to Mozambique Orogenic Belt marked at the surface by the Nandi fault at about 110 km from VIC (Fig. 3.22).

The basal boundary lacks continuity as a first order interface, and marks principally a change in velocity gradient.

The lower crustal boundary is constrained by a reflection from VIC, and from the interpretation of the axial line. Where this intra-crustal boundary appears dashed it has been defined as a second order discontinuity, where only a change in velocity gradient occurs.

The mid-crustal 6.5-6.6 km/s $V_p$ velocities are consistent with a compositional model of tonalitic and dioritic gneisses in amphibolite facies. The other end-member compatible with the velocity values is the intrusive equivalent of basalts and phonolites, and a mixture
of the two - gneisses intruded by mafic dykes especially under the rift axis in the mid-crust - cannot be ruled out (Mooney and Christensen, 1994).

Along the axis of the rift the 6.8 km/s layer thins to the north, suggesting that it is composed of ancient material extended by the rifting. An alternative interpretation that it is underplated material intruded during the rifting episode is less probable because the thickness along the rift is not constant (Mechie et al., 1994b). The lower crust velocities are compatible with a mafic gneiss composition in granulite facies mixed to mafic intrusions and igneous mafic residuum from the material accreted during the upper mantle melting (Mooney and Christensen, 1994), although this residuum can not be the bulk of the lower crust for the reason stated above.

It has also been suggested that the velocity decrease across the rift in the lowermost crust, from c.6.9 km/s to 6.8-6.7 km/s towards the central and eastern part of the model could be related to the passage from ANC to MOB lithosphere.

Upper mantle

This section concentrates on the uppermost mantle structure (Fig. 3.22), while the deepest part sampled by the seismic ray paths will be discussed in more detail in Chapter 6.

The depth to the Moho is relatively well constrained, as shown by the ray coverage at uppermost mantle depth in the ray path diagram (Fig. 3.20), to an accuracy of c.1-2 km. It varies from 37 km under the western flank, to less than 30 km under the axis, and deepens again to the east reaching 38 km at 420 km along the line. The shallowing under the eastern end is mainly constrained by the flank line interpretation.

The accuracy of the velocities was estimated to be 0.1-0.3 km/s by perturbing the values and keeping the travel time fit to within the picking accuracy. The $V_p$ velocity varies from 8.2 km/s under the western flank, to a minimum of 7.6 km/s under BAR and increases again to the east to 7.8-8.0 km/s. Again the value of 8.1 km/s under CHF results from interpretation of the flank line profile (Prodehl et al., 1994a).

The thickening of the crust at the passage from the ANC to the MOB is seen at Moho depth, but the major feature at this level is the upwelling of the mantle beneath the rift, with a corresponding decrease in velocity.
A step-like structure is modelled under the eastern margin of the rift, at 350 km along the model. The ray coverage is mainly unreversed on the step itself. Modelling of the available gravity data does not require the presence of the step (see Section 3.5). However, further tests involving ray-tracing proved that in the model without the step:

- from VIC provided a worse fit: the calculated travel times are delayed by up to 0.2 s where the picking accuracy and accuracy of fit are better than ±0.2 s,
- from KAP to the east is a better fit, but the data quality is worse than from the VIC shot,
- c from BAR to the east is delayed by up to 0.2 s, while d ray paths downshoot the step, emerging beyond 350 km along the Moho,
- c from TAN to the east is again delayed more than 0.2 s, beyond the picking accuracy and accuracy of fit, and
- d from BAS is not affected by the change.

Therefore the presence of the step cannot be excluded using the available data, which are unreversed over that portion of the model. Comment on the Moho topography will be added in Chapter 7, after the presentation of inversion of the travel time data.

The low-velocity region is depicted with a V-shape. Combined interpretation of gravity and teleseismic data lead to the conclusion that a few percent of melt is required to explain the low-velocity values (Green et al., 1991). A model has been hypothesised where upper mantle material of peridotitic composition at a raised temperature and including up to 5-6% melt could have risen from below in diapiric form, reached Moho depth and stopped at the base of the crust (Mechie et al., 1994a).

### 3.5 Gravity Modelling

Gravity modelling along seismic profiles provides an important control for the seismic interpretation. It can provide additional information on the structural boundaries and suggest extrapolation between seismic boundaries. At an initial stage the seismic boundaries are assumed to be density boundaries. Refinement of the model can enable a fit of the model to the gravity data.
CHAPTER 3. KINEMATIC MODELLING OF P-WAVES

For simple modelling of crustal profiles the following boundaries are generally considered (Setto and Meissner, 1987):

- the base of the sediments, or top of the basement, where $V_p$ velocities jump to c. 6.0 km/s,
- the intra-crustal boundary (Conrad), where $V_p$ velocities jump to values greater than 6.5 km/s, and
- the crust-mantle boundary (Moho).

A Bouguer anomaly gravity profile has been derived from the data presented in the catalogue by Swain and Khan (1977).

The gravity stations are more widely spaced than the seismic stations, but generally follow the same roads. Different roads were used only west of the Elgeyo escarpment, and here the lines diverge by 5 km.

One-hundred-and-two stations were used, covering the length of the seismic profile. Details on the gravity data corrections can be found in Swain and Khan (1977).

The distance between the stations has been calculated along a straight line with N68°E azimuthal direction. The different procedure used for the seismic line (Section 2.5.2) distances lead to discrepancies considered negligible at the survey scale.

The modelling and interpretation of the gravity data has been performed by Dr. C.J. Swain and has been previously reported in Maguire et al. (1994).

The gravity anomaly was calculated for a basically 2-D model with GRAVMAG (Busby, 1987) and an arbitrary constant was subtracted for comparison with the observed data.

The procedure followed is described here due to the important implications which extend the results of the seismic modelling.

Constraining the structural boundaries from the seismic model, a gravity model has been produced using the relation between velocity and density shown in Table 3.2. The Nafe-Drake relation has been used for the crust and upper mantle rocks, except for the portion of the upper mantle where melt is hypothesised, lowering the density difference compared to the velocity difference. Densities proposed in Swain et al. (1981) and Swain (1992) have been used for the volcanics and sediments.
CHAPTER 3. KINEMATIC MODELLING OF P-WAVES

Table 3.2: Velocity versus density values used in the gravity modelling.

<table>
<thead>
<tr>
<th>Unit</th>
<th>$V_p$ (km/s)</th>
<th>$\rho_{initial}$ (g/cm$^3$)</th>
<th>$\rho_{improved}$ (g/cm$^3$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sediments</td>
<td>2.0</td>
<td>2.10</td>
<td>2.10</td>
</tr>
<tr>
<td>Sediments</td>
<td>3.0</td>
<td>2.30</td>
<td>2.30</td>
</tr>
<tr>
<td>Upper volcanics</td>
<td>3.9</td>
<td>2.66</td>
<td>2.66</td>
</tr>
<tr>
<td>Lower volcanics</td>
<td>5.2</td>
<td>2.55</td>
<td>2.55</td>
</tr>
<tr>
<td>Lower volcanics</td>
<td>5.8</td>
<td>2.63</td>
<td>2.50-2.75</td>
</tr>
<tr>
<td>Basement</td>
<td>6.1</td>
<td>2.70</td>
<td>2.70-2.75</td>
</tr>
<tr>
<td>Basement</td>
<td>6.3</td>
<td>2.76</td>
<td>2.80-2.82</td>
</tr>
<tr>
<td>Basement</td>
<td>6.4</td>
<td>2.80</td>
<td>2.80</td>
</tr>
<tr>
<td>Upper crust</td>
<td>6.6</td>
<td>2.84</td>
<td>2.86</td>
</tr>
<tr>
<td>Lower crust</td>
<td>6.85</td>
<td>2.90</td>
<td>2.92</td>
</tr>
<tr>
<td>Upper mantle LVZ</td>
<td>7.6</td>
<td>3.18</td>
<td>3.20</td>
</tr>
<tr>
<td>Upper mantle</td>
<td>8.1</td>
<td>3.25</td>
<td>3.25</td>
</tr>
</tbody>
</table>

An initial density section is shown in Fig. 3.24: there is a major difference in overall gradient along the line. Lack of compensation of the topography rising from the east to the west across the Kenya Dome could produce the necessary gradient in the observed data to be consistent with that calculated. However, it has been shown that the East African Plateau is isostatically compensated (Banks and Swain, 1978; Bechtel et al., 1987).

The source of compensation cannot lie in the crust or uppermost mantle according to the seismic model structure and velocity variations. It seems unlikely that the compensation would be provided by crustal-density variations, so it has to lie deeper in the mantle.

Similar conclusions were reached by Ebinger et al. (1989): long wavelength topography and gravity signatures over the East African Plateau region can not be explained by lithospheric effects. It has been supposed that a simple convection mechanism in the asthenosphere provides dynamic uplift of the East African Plateau. The uplift and associated heating of the thermal lithosphere above this convecting region provides the compensation whose source lies at a depth between 160 and 425 km.

A new model has been derived (Fig. 3.25) where some density values of the seismic units have also been modified. The improved model's densities are also reported in Table 3.2. The lower volcanic unit density has been increased under the rift graben. The basement density has been increased from 130 km to 375 km. The lower volcanic unit density in the Kerio trough has been decreased. The Moho step under the eastern shoulder has been reduced.
Figure 3.24: The observed and calculated gravity profiles along the cross-rift line D showing the initial density model derived directly from the final ray-trace seismic model. The gravity effect of the deep compensation as discussed in the text is also shown. (Densities in $gcm^{-3}$).
Figure 3.25: The final density model of a cross-section through the crust and upper mantle beneath the Kenya Rift along the KRISP 90 east-west line D. (Densities in $g cm^{-3}$).
CHAPTER 3. KINEMATIC MODELLING OF P-WAVES

Following the isostasy studies, the regional gradient equivalent to the gravity effect of the deep compensating masses along the line has been calculated and removed. The regional gradient was determined filtering a 2500 km long topographic profile with a function approximating the observed admittance modified from Ebinger et al. (1989). A significantly better fit is achieved between calculated and observed gravity data (Fig. 3.25).

The implications of the seismic interpretation and its relation to the gravity data interpretation will be summarised in the following section.

3.6 Summary

The main points arising from the kinematic modelling of P-waves travel times can be summarised as follows:

- the supra-crustal units show thicknesses, velocities and densities higher than expected, with the exception of the Kerio trough. It has been hypothesised that a separate episode of rifting occurred in the Kerio Valley. In the main graben relatively high velocities and densities under Baringo could be related to mafic intrusions.

- basement rocks show a clear subdivision in structure and physical properties between the ANC and MOB. Within the MOB there is further differentiation in axial and flank characteristics, suggested as relating to the rifting process.

- the mid-lower crust is similar to the upper crust in both division between major crustal units and lateral changes within these units.

- upper mantle velocity variations and Moho topography mirror the width of the surficial expression of the rift. The gravity modelling shows that an overall gradient must be removed from the Bouguer anomaly data to fit the calculated data from the seismic model. This is consistent with the idea that the East African Plateau is dynamically compensated by sources lying deep in the mantle.

The kinematic forward modelling of P-waves will be integrated in subsequent Chapters with S-wave analysis, dynamic modelling, travel time inversion plus a detailed study on the deepest part of the model.
Chapter 4

Kinematic Modelling of S-waves

4.1 Introduction

Shear wave velocities when combined with compressional wave velocities provide better constraints on estimates of crustal composition and physical state than either on their own. Compared with P-wave studies there have only been a few attempts made to interpret S-waves generated by explosion during seismic refraction surveys (most notably: Braile et al., 1974; Keller et al., 1975; Zachau and Koschyi, 1976; Assumpção and Bamford, 1978; Hall and Ali, 1985; El-Isa et al., 1987; Holbrook et al., 1988; Gajewski et al., 1990).

In general, relatively little S-wave energy is to be expected from explosions compared with P-wave energy. Its generation is discussed by Edelmann (1985) who shows that it is dependent on the local conditions at the source, which can result in the generation of SH-waves by dynamite sources. It has also been shown that S-waves can originate from shear stress at an interface of velocity change (Gamburzew, 1938) by conversion from P-waves. As well as intralayer boundaries, the interface could be the free surface or the base of the weathered layer (Fig. 4.1a). The amplitude of the S phase depends on the angle of incidence and the $V_p/V_s$ ratio; a maximum in displacement amplitude occurs for angles of incidence just beyond $45^\circ$ and for a $V_p/V_s$ ratio of $\sqrt{3}$ (Fig. 4.1b).

Three component recordings are not often available, and so it is sometimes difficult to detect S arrivals, especially refracted S phases, which appear as second arrivals in the P coda. Scattered P-wave energy can mask S arrivals. Moreover, it has been observed...
Ray paths for converted waves in a homogeneous half space: 
\[ X = h \tan i \]

Ray paths of converted waves for weathered layer on consolidated layer: 
\[ \Delta t = \Delta t_p + \Delta t_s = \frac{Z}{V_p + 1/V_s} \]

Figure 4.1: a: selected ray-paths of converted waves; b: reflection coefficient for the pS converted wave at the free surface as a function of the angle of incidence (after Fertig, 1984).
that S-waves are attenuated in complex subsurface structures and the ground motion for shallow angle arrivals is strongly dependent on the ground conditions at the station (Assumpçao and Bamford, 1978).

Recent studies have shown that the P- and S-wave propagation paths can be different. Analysis of wide angle data from south-west Germany revealed a lack of S reverberations from the lower crust in the presence of long reverberating trains in the P arrivals from the same depth (Lüschen et al., 1987; Holbrook et al., 1988). Also no refracted S-waves propagating in the uppermost mantle were observed corresponding to a strong P arrival (Gajewski et al., 1990). This particular aspect will be considered in detail in Section 4.3.2.

From the interpretation of data recorded in the Black Forest (Southwest Germany) it has been proposed that combined observation of S-waves on wide angle and near vertical incidence recordings can detect the presence of anisotropy (Lüschen et al., 1990).

Laboratory measurements on rock samples under high P and T conditions have shown that P- and S-wave velocities behave differently for different compositions (Christensen and Fountain, 1975; Fountain and Christensen, 1989). Poisson's ratio (σ) is therefore a useful parameter to use in the estimate of compositional variation, being sensitive to variables such as quartz and feldspar content; in particular, high feldspar content results in a high σ value, and high quartz content in a low σ value (Kern, 1982).

σ can be related to $V_p$ and $V_s$ (Sheriff, 1991) as:

$$\sigma = \frac{V_p^2 - 2V_s^2}{2(V_p^2 - V_s^2)}$$

and, for a homogeneous isotropic solid:

$$V_p = \sqrt{\frac{\mu + 2\lambda}{\rho}} \quad V_s = \sqrt{\frac{\mu}{\rho}},$$

where:

$\mu$ is the shear modulus (rigidity) and

$\lambda = k - 2\mu/3$ (k is the bulk modulus, or incompressibility).

$V_p$ and $V_s$ therefore respond differently to changes in the elastic properties of a rock, and when there is linearity between stress and strain, σ can be related to the physical state of the rock (Fig. 4.2a); that is, estimates of the bulk physical properties of rocks, as well as lateral and vertical variations in these bulk properties may arise from P- and
S-wave analysis (Tatham, 1985).

S-wave velocity is more sensitive to porosity than that of the P-wave, therefore \( \sigma \) can be used as a porosity indicator (Fig. 4.2b); similarly, as \( V_s \) is not affected by the type of fluid which fills the rock pores and \( V_p \) is, \( V_p/V_s \) can be used as a fluid indicator (Fig. 4.2c) (Tatham, 1985). \( \sigma \) increases at high pore pressure (Spencer and Nur, 1976, Christensen, 1984).

Picking of the S arrivals and forward modelling ray-tracing will be described for the data recorded along the cross-rift and flank line of KRISP 90. The S-wave data quality and phase correlation were not sufficiently better than the P-waves to justify changes in the structure. Therefore a model distribution of \( V_p \) and \( \sigma \) along the two seismic lines was generated using the structural boundaries as determined via the P-wave interpretation, a method as suggested by previous authors (Braile et al., 1974; Keller et al., 1975; Assumpção and Bamford, 1978; El-Isa et al., 1987; Grad and Luosto, 1987; Holbrook et al., 1987).

### 4.2 Picking

Heelan (1953) calculated the generation of S-waves from borehole explosions and White and Sengbush (1963) and Lash (1980) proved their presence experimentally. S-wave energy produced by P-wave surface conversion was demonstrated by Fertig (1984). Neither explanation predicts displacement vectors with a transverse direction component, so it should only be possible to observe such S-waves on the vertical and radial component.

The present analysis has been carried out on the vertical component record sections scaled to allow the direct comparison of P and S arrivals, thus assisting the S phase identifications. For the scaling it is assumed that \( \sigma \) is equal to 0.25, giving a \( V_p/V_s = \sqrt{\frac{3}{5}} = 1.732 \).

A preliminary qualitative analysis was undertaken transforming the record sections with a reduction velocity of 3.46 km/s (corresponding to \( 6/1.732 \)) and with a time scale compressed by 1.732. The P picks were then overlaid onto these record sections, with the zero reduced time lines coincident, to help identify the S arrivals. This procedure allows an estimate to be made of \( V_p/V_s \); for a \( \sigma \) of 0.25 the P and S arrivals should exactly
Figure 4.2: a: correlation between $V_p/V_s$, $\sigma$ and $K/\mu$. Note the relative linear relation between $V_p/V_s$ and $K/\mu$ especially at large $V_p/V_s$ (after Tatham, 1985); b: left: $V_p/V_s$ vs porosity; right: interval transit time for P- and S-wave velocity trends vs porosity in water saturated sandstones (after Tatham, 1985); c: observed $V_p/V_s$ vs porosity for gas and water saturated well consolidated sedimentary rocks and confining pressure between 0 and $6.9 \times 10^7$ Pa (after Gregory, 1976).
superimpose. An early S arrival indicates a low $V_p/V_s$ and $\sigma$ and vice versa.

P- and S-waves can be separated by an appropriate band pass filter (Garotta, 1985; Lüschies et al., 1990) as the S-wave field has a lower frequency content; for example a detailed study of S-wave fields in the former USSR estimated $f_s/f_p=0.7-0.8$, somewhat higher intensity, and greater variability of the S amplitudes (Alekseev et al., 1988).

Spectral analysis of both P and S arrivals confirmed that the S-wave fields from the cross-rift line have typically lower frequencies, and a 1 to 12 Hz band pass filter was chosen for their display compared with a maximum of 1 to 25 Hz for the P-waves.

4.3 S Model for the Cross-rift Line

The preliminary analysis on the cross-rift line revealed that about 60% of the P-wave arrivals could be identified in the S field. The data quality is variable and generally poorer than for the P-waves. The S-waves identified are sometimes visible for an offset shorter than the corresponding P-wave phase.

The shots located outside the rift (VIC, KAP, BAS) produced good S-wave data for most of the phases, travelling both in the crust and the upper mantle.

The shots located in the rift (BAR and TAN) produced S-waves for rays travelling in the lower crust and upper mantle but did not produce clear first S arrivals.

In general the clearest phases are the diving waves through the upper crust and the Moho reflection.

The picking procedure was the same as for the P-waves; the accuracy was better than 0.2 s for the crustal phases (0.1 for the clearest from VIC and BAS) and 0.3 s for the deeper phases. A more detailed analysis of the various phases will be described in the following sections.

4.3.1 Record Section Description

According to the terminology proposed by El-Isa et al. (1987), in the study of lithospheric S-waves the main phases considered are:

- $S_s$ (phase $s_s$): S direct arrival from the sediments or the volcanic cover,
- \( S_g \) (phase \( g \)): diving S-waves through the upper crust,
- \( S_i \) (phase \( i \)): diving S-waves through the middle/lower crust beneath intracrustal boundaries,
- \( S_{i1}S \) and \( S_{i2}S \) (phases \( t_1 \) and \( t_2 \)): S-wave reflections from the middle and lower crust discontinuities,
- \( S_mS \) (phase \( m \)): S-wave reflection from the Moho discontinuity,
- \( S_n \) (phase \( n \)): diving S-waves in the uppermost mantle,
- \( x_i \): as for the P-wave sections, a possible multiple.

The notation above will be used throughout this Chapter, although some of the phases mentioned have not been identified on the line D record sections.

The S-wave record sections have all been normalised, band-pass filtered and plotted with a reduction velocity of 3.46 km/s; the lines represent observed travel times. Details of the filter parameters are in the figure captions to the relative sections. All the distances in the phase descriptions below are from the corresponding shot points.

**VIC - Fig. 4.3**

\( s \): has been picked to 180 km. It is in advance of the corresponding transformed P arrival by c.0.2 s between 50 and 140 km and up to 0.5 s to 180 km.

\( t_1 \): from c.60 to 150 km a low amplitude phase can be seen corresponding to \( b_1 \) from VIC.

\( t_2 \): is visible from c.80 to 160 km, coincident with the transformed P arrival.

\( u \): is the largest amplitude phase, detectable from 90 to 200 km and nearly superimposed on the transformed P arrival.

**KAP - Fig. 4.4, 4.5**

\( s \): can be picked to the west as a first S arrival from 15 to 140 km, earlier than the analogous P arrival; to the east the data are too noisy for a clear phase identification, although some S energy can be seen in the first 30 km from the shot, and then beyond the rift from 160 to 180 km.
Figure 4.3: Trace-normalized band-pass filtered (3-10 Hz) S-wave record section for shot-point VIC recorded to the east along the cross-rift line D. $V_r=3.46 \text{ km/s}$ (observed phase notation: $s=S_g$, $t_1=S_{i1}S$, $t_2=S_{i2}S$, $u=S_mS$).
Figure 4.4: Trace-normalized band-pass filtered (4-10 Hz) S-wave record section for shot-point KAP recorded to the west along the cross-rift line D. $V_w = 3.46$ km/s (observed phase notation for Fig. 4.4, 4.5: $s = s_g$, $t_1 = s_{11} S$, $u = s_m S$).
Figure 4.5: Trace-normalized band-pass filtered (4-10 Hz) S-wave record section for shot-point KAP recorded to the east along the cross-rift line D. $V_p=3.46$ km/s (observed phase notation: see Fig. 4.4 caption).
CHAPTER 4. KINEMATIC MODELLING OF S-WAVES

\( t_1 \): as for the corresponding P phase, this is a clear large amplitude phase to the west from 60 to 150 km, but the apparent velocity, 3.9 km/s, indicates a \( c \) lower than 0.25. To the east it is not visible.

\( u \): a high amplitude reflection from the Moho to the west is seen from 105 km to the end of the section, coincident with the transformed c phase. It cannot be seen to the east.

BAR - Fig. 4.6, 4.7

\( s \): as previously mentioned this phase is visible outside the rift margins, i.e. from c.50 km in both directions and it is a first S arrival to about 150 km.

\( t \): this is a first S arrival from 150 to 200 km to the east, earlier than expected from the P transformed phase.

\( t_1 \): some energy follows \( t \) from 120 to 220 km to the east.

\( u \): this phase is picked on a few traces from about 150 km to 180 km to the west, and to the east between about 170 and 200 km.

\( x_1 \): energy is visible from about 140 km to 180 km to the west, earlier than expected from the transformed P arrivals, possibly due to the travelling in the 'fast' Archaean Nyanza Craton (ANC); the same phase to the east is less clear but corresponds to the equivalent P phase.

TAN - Fig. 4.8, 4.9

\( s \): between 25 and 90 km is the first S arrival to the east.

\( t \): can be picked on a few traces from about 130 km to the east.

\( t_1 \): between 100 and 170 km to the west a reflected phase is visible; to the east it can be picked from 50 to 150 km.

\( u \): some energy can be identified from 100 to 150 km to the west and from 100 to 190 km to the east.

\( x_1 \): from 130 to 160 km to the east few traces show a weak S arrival.

The data set is of poor quality, and the phases are identified only by the superposition of the corresponding transformed P-wave arrivals.
Figure 4.6: Trace-normalized band-pass filtered (1-8 Hz) S-wave record section for shot-point BAR recorded to the west along the cross-rift line D. $V_r=3.46$ km/s (observed phase notation for Fig. 4.6, 4.7: $s=S_g$, $t=S_i$, $t_1=S_{ii}S$, $u=S_mS$, $x_2$=unidentified phase discussed in the text).
Figure 4.7: Trace-normalized band-pass filtered (1-8 Hz) S-wave record section for shot-point BAR recorded to the east along the cross-rift line D. \( V_s = 3.46 \text{ km/s} \). (observed phase notation see Fig. 4.6 caption).
Figure 4.8: Trace-normalized band-pass filtered (3-10 Hz) S-wave record section for shot-point TAN recorded to the west along the cross-rift line D. $V_s=3.46$ km/s (observed phase notation for Fig. 4.8, 4.9: $s=S_g$, $t=S_i$, $t_1=S_{i1}S$, $u=S_mS$, $x_u$=unidentified phase discussed in the text).
Figure 4.9: Trace-normalized band-pass filtered (3-10 Hz) S-wave record section for shot-point TAN recorded to the east along the cross-rift line D. $V_s=3.46$ km/s (observed phase notation see Fig. 4.8 caption).
CHAPTER 4. KINEMATIC MODELLING OF S-WAVES

Table 4.1: Phase analysis results; the numbers in the columns are the ranges in km where each phase is observed.

<table>
<thead>
<tr>
<th>S-P</th>
<th>VSC</th>
<th>KAP</th>
<th>BAR</th>
<th>TAN</th>
<th>BAS</th>
</tr>
</thead>
<tbody>
<tr>
<td>direction</td>
<td>E</td>
<td>W/E</td>
<td>W/E</td>
<td>W/E</td>
<td>W/E</td>
</tr>
<tr>
<td>phase</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>t</td>
<td>5-180</td>
<td>15-140/10-30+160-180</td>
<td>50-150/50-100</td>
<td>70-90</td>
<td>0-100/0-100</td>
</tr>
<tr>
<td>t1</td>
<td>-</td>
<td>-/100-200</td>
<td>-/130-135</td>
<td>-/100-100</td>
<td></td>
</tr>
<tr>
<td>t2</td>
<td>60-150</td>
<td>60-150/-</td>
<td>-/130-220</td>
<td>100-170/50-150</td>
<td>-/60-90</td>
</tr>
<tr>
<td>u</td>
<td>80-160</td>
<td>-/-</td>
<td>-/-</td>
<td>-/-</td>
<td>-/-</td>
</tr>
<tr>
<td>x</td>
<td>90-300</td>
<td>100-100/-</td>
<td>180-180/170-200</td>
<td>100-150/100-190</td>
<td>180-210/-</td>
</tr>
<tr>
<td>y</td>
<td>-/-</td>
<td>-/-</td>
<td>140-180/120-160</td>
<td>-/130-160</td>
<td>-/-</td>
</tr>
</tbody>
</table>

BAS - Fig. 4.10, 4.11

s: a very clear first S arrival is picked to c.100 km in both directions.

t1: has been picked from c.40 to 90 km to the east.

w: can be identified to the west between 180 and 210 km, i.e. beyond the rift's western shoulder.

The phase analysis is summarised in Table 4.1.

4.3.2 Ray-tracing Results and Phase Analysis

S-wave travel time forward modelling was undertaken by ray-tracing.

The structural boundaries determined from the interpretation of the P-wave data were kept unaltered, the same interfaces being assumed for the S-waves. A P- and S-wave homogenous radiation pattern from the source was assumed.

For the S-wave starting model the P-wave velocities were divided by 1.732. The adjustments to the model consisted in alterations to the V_s velocity field.

As for the P-waves, the 2-D forward modelling of the S-waves was performed using the USGS MacRay (Luetgert, 1992b) ray-tracing routine.

The $\sigma$ distribution can be calculated and displayed for the whole velocity field.

P- and S-waves may follow different ray-paths if the $V_p/V_s$ ratio changes in the crust, the two wave fields having to be modelled independently and then recombined to express $\sigma$. Strictly speaking this means that the resulting $\sigma$ values are invalid because P and S ray
Figure 4.10: Trace-normalized band-pass filtered (3-12 Hz) S-wave record section for shot-point BAS recorded to the west along the cross-rift line D. $V_p=3.46$ km/s (observed phase notation for Fig. 4.10, 4.11: $s=S_g$, $t_1=S_{11}S$, $u=S_mS$).
Figure 4.11: Trace-normalized band-pass filtered (3-12 Hz) S-wave record section for shot-point BAS recorded to the east along the cross-rift line D. $V_r = 3.46$ km/s (observed phase notation: see Fig. 4.10 caption).
CHAPTER 4. KINEMATIC MODELLING OF S-WAVES

paths do not necessarily sample the same region; however, it is accepted that the velocity values are determined over layers where the suspected gradients are sufficiently low and the two ray paths are sufficiently close to enable the provision of average $\sigma$ values.

Figures 4.12-4.14 show the final travel time plots resulting from the iterative 2-D ray-trace modelling: the fit between the observed and the calculated travel time is better than 0.2 s for the crustal phases shown in the diagrams. For the mantle phases the fit is generally better than 0.3 s. All the distances in the ray-trace model description are given along the profile from the western end.

Upper crust - (Fig. 4.12)

$s$

From VIC $s$ is earlier than the transformed $a$. However it is difficult to fit such an advance beyond 150 km changing the S velocity values alone. A misfit of 0.2 s is present up to 180 km from the shot point.

$s$ from KAP to the west has the same advance compared to $a$ and fits with the $V_s$ velocity modelled from VIC in the basement.

An increase in the $V_s$ velocity of the volcanics is also necessary to fit the few arrivals to the east of KAP.

From BAR to the west the delayed arrivals outside the rift require the velocity to be lower in the sediment-volcanic cover in the rift. To the east rays can be traced only between 330 and 350 km, i.e. outside the rift.

TAN to the east also requires a lower $V_s$ velocity in the basement cover, to fit the arrivals beyond the rift’s eastern margin.

The $s$ phase from BAS suggests a normal $\sigma$ of 0.25, in both directions.

Lower crust - (Fig. 4.13)

$t$

Lower crustal diving waves are observed from BAR to the east requiring a slight increase in the mid-crust velocity gradient to fit the S-wave arrivals.

From TAN to the east it proved difficult to ray trace the emergent arrivals which appear to fit better as $t_1$ phases.
Figure 4.12: $s$ phase ray-traced final model (for velocity distribution see Fig. 4.17. Number of rays for each phase limited to a maximum of 10 for display). Vertical exaggeration 10:1.
Figure 4.13: $t$, $t_1$, and $t_2$ phases ray traced final model (for velocity distribution see Fig. 4.17. Number of rays for each phase limited to a maximum of 10 for display). Vertical exaggeration 5:1.
CHAPTER 4. KINEMATIC MODELLING OF S-WAVES

$t_1$
From VIC, rays traced to fit the reflection off the intracrustal boundary arise between 90 and 135 km. This phase traced from the KAP shot point to the west gives a 0.3 s early arrival.

The BAR phase to the east is consistent with the few rays traced in the S model between 120 and 150 km. The diffuse energy to the west requires rays emerging between 20 and 80 km.

From TAN it is possible to trace rays to fit the phase between 90 and 170 km on the western shoulder of the rift; to the east few rays fit with the high amplitude energy seen between 350 and 400 km. This arrival from BAS to the east seems to be earlier than the corresponding P phase, requiring a high $V_s$ towards the eastern end of the model at mid-crustal level.

This is only seen from VIC. It is modelled with a critical point at 135 km.

Upper mantle - (Fig. 4.14)

$u$
From VIC this phase can be traced giving a satisfactory fit from the critical point at c.90 km to the end of the visible phase at 200 km.

The reversed shot from KAP to the west does not require significant changes in the $V_s$ velocity with respect to $V_p/1.732$, and it can be traced from 20 to 50 km along the line.

$u$ can be only traced to the west of BAR between 80 and 100 km and to the east between 380 and 440 km.

Rays shot from TAN are 0.3 s later than the few $u$ arrivals picked in both directions; nevertheless, the data quality for this phase is not good enough to justify a change in the model.

From BAS to the west it is difficult to trace rays beyond the rift's western margin, although this is where the arrivals have been picked.

$v$
Diving S-waves in the uppermost mantle are never observed, despite good S-wave reflections off the Moho. The same phenomenon has been observed in France (Haggag,
Figure 4.14: $u$ phase ray traced final model (for velocity distribution see Fig. 4.17. Number of rays for each phase limited to a maximum of 10 for display). Vertical exaggeration 5:1.
1980), SW Germany (Gajewski and Prodehl, 1985), Scandinavia (FENNOLOLA: Prodehl and Kaminski, 1984) and in the north-eastern United States (Gajewski et al., 1990). It has been suggested that the S velocity in the upper mantle has a much smaller gradient than the P velocity, resulting in a $V_p/V_s$ ratio increase just below the Moho. The proposed cause for this increase is a change from mafic to ultramafic composition over 10 km below the Moho (Gajewski et al., 1990).

Moreover, as mentioned in Section 1.5.2, Gumper and Pomeroy (1970) suggested the presence of a 'gap' in the mantle portion of the lithosphere beneath the Rift. Analysing data from the WWSSN they observed that $v$ does not propagate across the Rift above the equator, whereas it does below c.10°S.

On the other hand, amplitude picking of the $d$ and $v$ phases revealed that $v$ is masked by the P-wave coda "noise". Amplitude modelling of the two phases by reflectivity for a 1-D model at VIC results in the amplitude ratio being 2.5 to 4.0 in the range 200-250 km, giving a $v$ amplitude lower than the "noise" level on the record section in the S-wave window at the same range (Fig. 4.15). The modelling was undertaken with an S velocity gradient equal to the P gradient; a lower gradient would further increase the amplitude ratio.

As for the corresponding P arrival, the $x_s$ phase detected on the BAR record section in both directions and TAN to the east, could be a precrical phase from an upper mantle reflection or a multiple phase. The hypothesis of a converted phase for the corresponding P arrivals can be now rejected, because $x_s$ is visible at the same position on the S sections. Dynamic modelling will help identify this phase as a multiple reflected within the rift infill (see Section 5.3.2).

4.3.3 Model Description

A description of the model would not be complete without a picture of the S-wave ray coverage; for comparison the P-wave coverage is also depicted (Fig. 4.16).

As suggested from the preliminary observations, the upper crust is well sampled outside the rift; the mid crustal boundary is reasonably covered but the data quality is less good.
Figure 4.15: Amplitude of observed P diving waves in the uppermost mantle and noise amplitude in the S time window. The $S_n$ theoretical amplitudes are calculated from $P_n$ using amplitude ratios derived from reflectivity modelling.
Figure 4.16: Top: S ray trace coverage. Bottom: P ray trace coverage.
Figure 4.17: Final 2-D ray-trace model for KRISP 90 cross-rift line D. Dashed lines indicate transitions where there is a change in velocity gradient but no significant discontinuity. S-wave velocities in km/s; Poisson’s ratio in italic.

The Moho depth is constrained by a high amplitude clear phase on most of the record sections, but the coverage is limited to short sections. The best fit model comprises an east-west section down to a depth of 40 km (Fig. 4.17). $V_s$ velocities and Poisson’s ratios are reported in the diagram.

Perturbations of the model showed that the resolution of the $V_s$ values is about 3%; this results in a numerical uncertainty on $\sigma$ of $\pm 0.03$. Perturbations of $\sigma$, assuming the specified picking accuracy result in about 4% resolution as a minimum. The resolution decreases in the areas where the data and ray coverage is minimum and in the mid-lower crust.

$\sigma$ is calculated from the $V_p$ and $V_s$ models using the relation (4.1).
Where no direct arrivals were identified or where no rays traversed the model the $\sigma$ was set to 0.25. This is the case for example for the sediments beneath the VIC shot point, in the Kerio Valley and in the Baringo basin.

In the sedimentary units $\sigma$ is generally equal to 0.25, with a locally high value (0.34) under the TAN shot point.

The volcanic units are associated with a value of $\sigma$ of up to 0.35 under KAP; in the rift itself the values are between 0.25 and 0.30. A low S-wave velocity has been introduced in the basement cover to fit the late $s$ arrivals outside the rift: the lack of direct arrivals in the rift itself does not allow a clear identification of the unit which is delaying the S arrivals.

The basement from VIC to 80 km along the line has $\sigma$ equal to 0.23, increasing to the more normal value of 0.25 towards KAP to TAN. The high $V_p$ velocity block under BAS requires a low $\sigma$ value of 0.23-0.24. East of BAS and to CHF values are again normal at around 0.25.

The mid-crustal and lower crustal layers have $\sigma=0.25$. In the lower crust $\sigma$ tends to decrease towards 0.23-0.24 under KAP and with depth.

4.3.4 Discussion

High $\sigma$ values are commonly associated with sedimentary rocks, and more generally exist near the surface as indicated by various field measurements (e.g., Tatham and Stoffa, 1976; Scarascia et al., 1976). Gregory (1976) measured a broad range of $\sigma$ values for dry sandstones and limestones, which were generally less than 0.25. The presence of unconsolidated and soft sedimentary deposits at the surface contribute to a high $\sigma$ value, as fluid saturation in a highly porous rock significantly increases $\sigma$ (Scarascia et al., 1976; Gregory, 1976). Values of $V_s$ in tuffs from Carmichael (1982) are in the range 1.4-1.8 km/s, resulting in $\sigma$ between 0.29 and 0.38.

In the basement rock the most significant feature is the low $\sigma$ in the Archaean Rocks and in the high $V_p$ velocity block.

A study of S-wave velocity variations in the Lewisian (Precambrian) metamorphic complex in NW Scotland indicated how the proportion of mafic versus felsic and hydrous
versus anhydrous minerals can be separately identified. Vertical variations in $\sigma$ are due to crack deformation and saturation (Hall and Ali, 1988). Kern (1982) measured a low $\sigma$ for rocks with high quartz content.

A seismic profile across the Archaean-Proterozoic terranes in central Finland revealed lower $\sigma$ for the older crust related to fracturing and saturation of cracks (Grad and Luosto, 1987). The low $\sigma$ values in the ANC could be related to higher quartz and lower feldspar content than in the Mozambique Orogenic Belt (MOB), or to anisotropy in the MOB. A higher quartz content in the ANC association could produce a high $V_s$ velocity; also the shear structures in the MOB stretching NS and NNW-SSE (Smith and Mosley, 1993; also NW-SE trends are present; Section 3.1) could result in a low velocity in the NNE-SSW direction observed. However, the ANC association also contains mafic rocks, whose low $\sigma$ could be due to a high hydration percentage (i.e. high amphibole; high retrogression of granulites). The ANC migmatitic remnant at 5.0 km depth could have a high percentage of hydrated minerals. The effect of attenuation will be considered during discussion of the S-wave quality factor ($Q_s$) values while modelling the S-waves by reflectivity (Section 5.3.3).

Taking into account the danger of extrapolating parameters which vary from place to place, even within the same tectonic environment, some notes can be added on the mid-lower crustal part of the model. At mid-lower crustal depths it is interesting to note that $\sigma$ does not differ from 0.25. However, the ray-path contribution in the upper crust accelerates the S arrivals at the surface. Removing the effect of the upper crust gives a $\sigma$ higher than 0.26 for the mid-crustal rocks, and a $\sigma$ of 0.25 for the lower crustal rocks, to achieve the travel time fit with the $t_2$ and $u$ phases. Although this increase is not largely beyond the significance of $\sigma$ (Section 4.3.3) it is interesting to note the effect of the different components of the ray path in the total travel time and the resulting $\sigma$.

High $\sigma$ (up to 0.32) in the mid-crust has been found in areas of normal faulting associated with high heat flow values, thin crust and a crustal LVZ (Basin and Range, Braile et al., 1974).
The combined effect of temperature and pressure at depth gives an increase in $\sigma$ of about 0.001 km$^{-1}$ (Simmons, 1964; Christensen, 1965, 1966; Birch, 1969). For the depth range of 13 to 34 km an increase of 0.02 is expected. The average value calculated from a data compilation by Holbrook et al. (1992) indicated $\sigma = 0.26 - 0.27$ in Precambrian Shield areas, and $\sigma = 0.29$ for the lower crust in rift areas (Fig. 4.18a). There was no data given for continent-continent collision zones.

The value observed is therefore lower than expected for a tonalitic/dioritic gneissic composition, and is consistent with a quartz-rich granitic/gneissic composition, which was not suggested in the $V_p$ model interpretation (Fig. 4.18b). Alternatively, the 'low' $\sigma$ value in the mid-crust could be due to the presence of fluids at low pore pressure and at temperatures of 200 – 300°C (Spencer and Nur, 1976).

The lower crustal values of $\sigma$ equal to 0.26 fit with a mafic granulitic composition as suggested by the $V_p$ velocity analysis.

Although adequate laboratory data for attenuation and velocity in partially molten rocks is unavailable, generally partial melt should appear as a low-velocity, low Q feature (Evans and Zucca, 1993). Temperatures near the melting point or partial melting attenuate S-waves more than P-waves, resulting in a $\sigma$ increase (Zschau and Koschycyk, 1976). Gajewski and Prodehl (1985) used the lack of shear wave energy from wide angle data to infer the presence of partial melt at the geothermal anomaly in Urach (SW Germany). The absence of S is used in attenuation tomography to locate magma chambers. No analysis of the S-waves from the KRISP 90 teleseismic data has been undertaken.

Extensive melting in the lower crust or disturbances of the Moho interface due to the presence of high percentages of melt in the upper mantle can be excluded, because $v$ is clearly observed over a significant length of the Moho.

The absence of $v$ can not be used as evidence for the presence of partial melt in the uppermost mantle, because, as previously discussed, the P coda noise level is too high to allow detection of the low amplitude refracted phase $v$. 
Figure 4.18: a: Histograms of cross-sectional area of lower crust versus $\sigma$ in rifts and shields; darker shading indicates higher quality factor; bars at the top indicate velocity ranges; b: Comparison of field- and laboratory-measured P-velocities and $\sigma$s. Field measured data from several areas (cross-hatched ovals) are plotted on fields of lab-measured data from a compilation by Holbrook et al. (1992). R=Rift Zone; Pc=Precambrian Shield; Pz=Palaeozoic crust. Numbers are laboratory measured data from lower-crustal rock types: 1=Quartzite (granulite); 2=Felsic amphib. gneiss; 3=Felsic granulite; 4=Quartz-mica schist; 5=Intermediate granulite; 6=Anorthosite; 7=Mafic granulite; 8=Amphibolite; 9=Metapelite (granulite); 10=Pyroxenite; 11=Eclogite; 12=Dunite/Peridotite.
CHAPTER 4. KINEMATIC MODELLING OF S-WAVES

4.4 S Model for the Flank Line

KRISP 90 line E lies across the north-eastern flank of the rift from the centre of Lake Turkana (LTC), intersecting line D at the CHF shot point. The P-wave model is reported in Prodehl et al. (1994a; Fig. 2.2a). Basement rocks outcrop along almost the whole onshore part of the line, producing clear S-wave arrivals. It is therefore interesting to compare the data and results with those from line D.

4.4.1 Record Section Description

The same notation as used for the line D record sections will be used here.

$s_1$ is also included representing a reflection from an upper crustal reflector.

As for line D, the record sections have all been normalised, band-pass filtered and plotted with a reduction velocity of 3.46 km/s. Details on the filtering are given in the figure captions of the relevant record sections. All the distances in the phase description are from the corresponding shot points. The LAI S-wave data set appears already in Prodehl et al. (1994a) and therefore it will not be presented here.

The lower frequency content of the underwater shots (LTC and LTS) in comparison with the borehole shots (ILA and CHF) is noticeable; the same effect can be observed on the P-wave record sections (cf. Section 2.5.6).

Generally the $s$ and $t_1$ phases are seen on all record sections as well as a high amplitude phase $u$. About 58% of the P phases are positively identified in the corresponding S field.

CHF - Fig. 4.19

$s$: is visible to the west to 50 km and at short range, to 25 km, to the east.

$s_1$: interpreted as a reflection from an intrusion of anorthositic gabbro in the P-wave model (Prodehl et al., 1994a), is still visible in both directions up to 30 km from CHF.

$t_1$: is picked to the west to 70 km.

$u$: is clear from 95 to 190 km to the west, with the same shingling structure observed on the P sections, although the delay is here between 120 and 140 km, and an advance with respect to the superimposed P arrivals is visible between 155 and 170 km. The long
Figure 4.19: Trace-normalized band-pass filtered (2-10 Hz) S-wave record section for shot-point CHF recorded along the flank-rift line E. $V_s = 3.46$ km/s (observed travel times notation: $s = S_g$, $s_1 = $reflection from an upper crustal reflector, $t_1 = S_{s1}S$, $u = S_mS$).
wave train appears more energetic than \( c \), as observed on the ILA record section to the east for the same phase. Reverberation or multiple reflection of \( u \) could explain the energy in the \( s \) coda emerging with a 1 s delay with respect to \( u \).

**ILA - Fig. 4.20, 4.21**

\( s_2 \): the direct arrival has been picked in both directions to 20 km.

\( a \): a clear arrival to c.70 km in both directions.

\( t \): can be seen from 100 to 170 km to the east but earlier than the corresponding \( P \).

\( t_1 \): has been identified from 40 to 80 km to the west and it is earlier than \( P \); it is also visible from 60 to 170 km to the east.

\( u \): to the west from c.80 to 105 km is delayed compared to the corresponding \( P \) arrival, and shows a possible multiple at about 60 km and 6 s reduced time; again on the \( P \) section a lamination in the Moho discontinuity is hypothesised from the character of \( c \), which could explain the 'multiple’ seen on the \( S \) section. To the east from 80 to 140 km \( u \) is much more energetic than the corresponding \( P \) phase.

Reverberations from the lower crust are also visible to the west.

A high amplitude phase arriving at c.2.5 s reduced time from 20 to 30 km to the west of ILA could be a converted phase from the Moho.

**LTS - Fig. 4.22**

As the shot was fired in Lake Turkana, the first 40 km of \( s \) are not recorded; it is observed from 40 km to 100 km from the shot.

\( t \): is the first arrival \( S \) phase from c.100 to 220 km.

\( u \): is of high amplitude from 65 to 106 km, maintaining the complexity shown on the \( P \)-wave section, with a delayed branch between 100 and 120 km.

**LTC - Fig. 4.23**

\( t_2 \): a reflected phase from an intra-crustal boundary is seen from 180 to 280 km and is earlier than the corresponding \( P \) phase.

\( u \): is visible only from c.150 to 200 km.
Figure 4.20: Trace-normalized band-pass filtered (2-10 Hz) S-wave record section for shot-point ILA recorded to the west along the flank-rift line E. $V_r=3.46$ km/s (observed travel times notation for Fig. 4.20, 4.21: $s=S_g$, $s_1$=reflection from an upper crustal reflector, $t_1=S_{11}S$, $u=S_mS$).
Figure 4.21: Trace-normalized band-pass filtered (2-10 Hz) S-wave record section for shot-point ILA recorded to the east along the flank-rift line E. $V_r=3.46$ km/s (observed travel times notation: see Fig. 4.20 caption).
CHAPTER 4. KINEMATIC MODELLING OF S-WAVES

Figure 4.22: Trace-normalized band-pass filtered (1-10 Hz) S-wave record section for shot-point LTS recorded along the flank-rift line E. $V_r=3.46$ km/s (observed travel times notation: $t=S_t$, $u=S_m S$).
Figure 4.23: Trace-normalized band-pass filtered (3-12 Hz) S-wave record section for shot-point LTC recorded along the flank-rift line E. $V_s=4.62$ km/s (observed travel times notation: $t_2=S_{12}S$, $u=S_mS$).
4.4.2 Ray-tracing Results and Phase Analysis

The ray-tracing diagrams from MacRay (Luetgert, 1992b) are displayed in Fig. 4.24 and 4.25; the accuracy of the fit being the same as for the cross-rift line (Section 4.3.2). Rays were traced more easily than through the cross-rift line because of less structural complexity in the flank line. The first 100 km from LTC were mainly constrained by the P-wave model, based on the results from the axial profile (Gajewski et al., 1994; Mechie et al., 1994b).

Upper crust - (Fig. 4.24)

The relative simplicity of the model allowed the direct and diving crustal phases to be traced over the whole length of the profile, apart from $s$ to the SE from LTS which was only traced from 50 to 75 km from the shot.

$t$ from ILA to the east is seen up to 130 km from the shot point.

$t_1$ to the west of CHF is consistent with the calculated arrival to 60 km; from ILA to the west, rays are shot to 70 km, and to the east only to 90 km, due to the presence of the high velocity intracrustal reflector south-east of ILA at c.4.0 km depth.

Lower crust and Upper mantle - (Fig. 4.25)

$t_2$ originates from rays reflected off a lower crustal interface at about 20 km depth, to an offset of 270 km from LTC to the east.

$s$ from CHF to the west is modelled as a reflection from a single interface for the range where the phase is observed, but the shingling character of the observed P and S phases suggests a laminated structure of the Moho, not inconsistent with the intensive volcanic activity in the area (Prodehl et al., 1994a). The energy seen one second later in the $u$ coda could therefore be an effect of the lamination but it also is consistent with a multiple reflection from the crust-mantle boundary in the sedimentary layer.

The same observation can be made for $u$ from ILA to the west, and again the late arrival could be a multiple within the sediments. From ILA to the east the rays are reversed with those from CHF to the west, and show the same high amplitude and long reverberative wave trains, more energetic than the corresponding P phase.

Rays can be shot from LTS to the east to be consistent with the phase $u$ to 100 km from the shot, although again a single crust-mantle interface model does not describe the
Figure 4.24: $s$, $t$, and $t_1$ phase ray-traced final model (for velocity distribution see Fig. 4.26. Number of rays for each phase limited to a maximum of 10 for display). Vertical exaggeration 10:1.
Figure 4.25: \( t_2 \) and \( u \) phase ray traced final model (for velocity distribution see Fig. 4.26. Number of rays for each phase limited to a maximum of 10 for display). Vertical exaggeration 5:1.
CHAPTER 4. KINEMATIC MODELLING OF S-WAVES

Figure 4.26: Final 2-D ray-trace model for the KRISP 90 flank line E. Dashed lines indicate transitions where there is a change in velocity gradient but no significant discontinuity. S-wave velocities in km/s; Poisson’s ratio in italic.

complex shape of the reflection.

From LTC to the east rays are traced to 190 km from the shot; an increase in $V_s$ compared to $V_p/1.732$ is necessary to fit the arrival time.

4.4.3 Model Description

Fig 4.26 represents the best fit seismic model for the S phases along the flank line. $\sigma$ values are also reported. There are a few observations to be made:

- there is no significant overall difference from $\sigma$ equal to 0.25 throughout the model.

There is some indication of a lower $\sigma$ under CHF, suggested by the $t$ phase from ILA arriving earlier than the corresponding P. The same may exist under Turkana, as indicated
by the $t_2$ and $u$ phases from LTC advanced compared to the transformed P. The latter observation could be confirmed by the data recorded along the axial line, but no S model along it is available;

- Prodehl et al. (1994a) modelling the S-waves indicated a value of 0.26 for $\alpha$ under the eastern part of the model at lower crustal depths. This value has been included in the present model. They suggest that south-east of LAI the lowermost crust is transitional to the Moho. At shallower depths the basement shows an anomalously high P velocity;

- the most striking feature is not the variation in $\alpha$, but the different P and S reflectivity especially in the $u$ phase from the Moho interface between ILA and CHF. The S arrivals show long wave trains and high amplitude content. This may be due to a high effective $Q_s$; the lack of radial and horizontal components prevents any further analysis. Modelling by reflectivity for the vertical component, however, suggests that S-waves must be converted from P to have the high amplitude observed. Mode conversion from P-waves could occur at the free surface; the travel times of the observed S phases indicate this is the most probable explanation as the waves have travelled for most of their path as S;

The presence of the $u$ arrivals confirms the suggestion of Prodehl et al. (1994a), that the laminated Moho is not due to recent accretion from the mantle; in a hot regime $u$ would be strongly attenuated, and the $V_p$ velocity would be lower than the observed 8.1 km/s.

### 4.5 Summary

Some of the remarks made in the analysis and travel times modelling of the S-waves can be synthesised here. From observations of the S-wave field recorded along both the cross-rift and the flank line, the travel times and amplitude content suggest that the S-phases observed are generated by conversion from P at the surface.

Comparison of the two data sets from line D and E indicate that S-wave propagation outside the rift is better than inside. The study of mantle phases cannot be used as confirmation that a hot regime exists beneath the axis of the rift because $u$ is observed and also if $v$ had been recorded it would be masked by the P-coda.
More specifically from the cross-rift line kinematic modelling, a lower $\sigma$ is observed in the ANC (c.0.23) than in the MOB (c.0.25), which could be explained by compositional differences. The value of $\sigma$ equal to 0.26 in the mid-crust may be due to the presence of fluids at low-pore pressure.

The $x$ phase recorded in the P-wave domain is not a converted arrival and thus the most likely explanation for this phase is as a multiple reflected within the rift infill (Section 5.3.2).

From observation of the flank-line S-wave field, a high effective $Q_s$ in the uppermost mantle could be present. Lamination on the Moho boundary seen in the P-wave model (Prodehl et al., 1994a) is suggested also by the S-waves arrivals.
Chapter 5

Dynamic Modelling

5.1 Introduction

Important analyses can be undertaken using synthetic seismograms:

1. dynamic modelling allows a quantitative comparison between observed and calculated data, using the amplitude and frequency information of recorded phases;

2. the importance of converted waves or non-ray waves can be identified via comparison of an observed section with that produced via synthetic seismogram generation;

3. the effect of variation in quality factor (Q, specific attenuation) due to intrinsic attenuation and scattering can also be examined;

4. seismic amplitudes are sensitive to velocity gradients and discontinuities in the earth, which can thus be refined in the velocity model through amplitude modelling.

Generally, commercial seismic reflection prospecting involves the use of reflected phases recorded at near vertical incidence. Their modelling includes the use of theoretical seismograms constructed at zero offset. In refraction seismology wave methods are used, which are applicable to vertically and/or laterally inhomogeneous media. The most frequently used methods can be divided into:

1. those which are based on the decomposition of the wave field into elementary waves, termed ray methods (Červený et al., 1977), and

2. those which compute the whole wave field, based on the exact solution of the equation of motion of an elastic medium (Fuchs and Müller, 1971; Kelly et al., 1976).
1. Ray methods can be used for sub-horizontally layered media with curved interfaces and recordings obtained along a linear profile; 3-D calculations are also possible. The three principal high frequency asymptotic methods are the standard ray method (SRM), the paraxial ray method (PRM) and the Gaussian beam method (GBM).

2a. For a vertically inhomogeneous layered medium the reflectivity method can be employed to construct synthetic seismograms (Fuchs, 1970). The name originates from the function which is integrated, namely the reflection coefficient or reflectivity of a layered medium (Fuchs and Müller, 1971).

2b. Finite-difference and finite-element methods compute the whole wave field and can be applied to laterally inhomogeneous media with strongly curved interfaces and including vertical discontinuities. They can be successfully used in the method developed by Alekseev and Mikhailenko (1980) which operates in the wavenumber-frequency domain and allows the computation of seismograms for a complex vertically and laterally inhomogeneous half-space.

The ray methods (Červený et al., 1977) and the generation of full-wave synthetic seismograms (Fuchs and Müller, 1971; Kelly et al., 1976) are discussed here.

Synthetic seismograms have finally been computed using three methods: the standard ray method (Červený and Pšenčík, 1981), the reflectivity method (Fuchs and Müller, 1971) and finite-difference modelling (Sandmeier, 1990).

5.2 Ray Method

In the SRM the following steps are applied in the calculation of synthetic seismograms (Fertig and Pšenčík, 1985):

1. boundary value ray-tracing (two-point ray-tracing) needs to be performed to determine the ray paths and travel times; the position of both source and receivers have to be specified. The search for the ray path may be solved by the bending or the more reliable shooting method. The bending method guesses an initial path and perturbs it until the appropriate differential equations of the rays are satisfied. The shooting method iteratively alters the initial conditions (shooting angle) to find the rays that pass through
the endpoint;

2- dynamic ray-tracing involves evaluating the matrix of geometrical spreading from source to receivers, the matrix of the second derivatives of the travel time field and the orientation of the polarisation vectors. The reflection/transmission coefficients are also evaluated. The above quantities, except the matrix of the second derivatives of the travel time field, are resolved at the receivers and stored for the next stage. Ray paths are not needed and can be discarded;

3- the determination of complex geometrical amplitudes can be completed knowing the wave amplitude at the source, the geometrical spreading and the reflection/transmission coefficients along every ray;

4- with the knowledge of the high-frequency (HF) field characteristics from (3) above and the travel times, the elementary synthetic seismogram is evaluated for a known HF source-time function.

The PRM uses initial value ray-tracing and is computationally simpler: only the initial angle of incidence is required and the receiver positions are used only in step (3). It also evaluates the wave field primarily in the vicinity of the ray, with decreasing accuracy away from the ray.

As mentioned in Section 3.4.1 the ray method is inapplicable and gives low accuracy or even invalid ray amplitudes in singular regions of the ray method. This means that both the SRM and PRM become unreliable in the following situations:

- caustic regions ("a surface to which rays emanating from a single source and reflected by a curved reflector are tangent" (Sheriff, 1991)),
- critical points,
- transition zones between shadowed and illuminated regions,
- areas of sudden change in vertical gradients,
- corners in the interfaces, and
- oscillations of the velocity distribution and interfaces.

In these cases anomalous amplitudes occur and incorrect synthetics are calculated (Červený, 1985).

The GBM uses a weighted superposition of the beams emerging around the receivers
in the evaluation of the synthetics, introducing a frequency-dependent smoothing effect. It is valid in singular regions of the ray method, and does not require two-point ray-tracing.

Limited use has been made of the ray method for calculating synthetic seismograms. The routine available (SEIS81, Červený and Pšenčík, 1981) employs an algorithm based on the SRM with computational accuracy limited in certain regions. The wave field is computed as a superposition of elementary waves, corresponding to the zero order approximation (the first term expansion of the wave equation) of the ray method. For each elementary wave, two-point ray-tracing is performed by the shooting method from the source to all the specified receivers where the travel times, complex amplitudes, phase shifts and synthetic seismograms are to be determined (Červený and Pšenčík, 1984). Ray codes may be defined to generate arbitrarily refracted and multiply reflected waves.

The 2-D model may include lateral inhomogeneities, curved interfaces, block structures, isolated bodies, fractures and vanishing interfaces; the velocity field may vary both laterally and vertically. The source may be anywhere in the model, and the receivers may be located regularly or irregularly at the surface. The P- and S-wave source radiation pattern can be independently defined.

The source time function has a Gaussian envelope with zero time corresponding to the envelope maximum:

$$f(t) = e^{-\frac{(t - \tau)^2}{2 \tau^2}} e^{i(\omega t + \phi)}$$

where: $t$ is the time, $\omega = 2 \pi F_M$, ($F_M$ is the dominant frequency), $\gamma$ is the number of extrema in the wavelet, and $\phi$ is the phase shift.

$\psi$ may be altered in order to enable alignment between the onset of the observed wavelet and the calculated one. The dominant frequency for which the seismograms are calculated may also be chosen. It is possible to simulate a wide variety of wavelets by choosing appropriate $F_M$, $\gamma$ and $\psi$ (Červený et al., 1977) (Fig. 5.1).

The theoretical seismograms can be plotted using normalized amplitude scaling or true amplitude display.

The geometrical spreading of the rays is determined by solving the dynamic ray-tracing system (two linear ordinary differential equations) using a modified Euler-Cauchy's
Figure 5.1: Function \( f(t) \) versus \( 2\pi f_m t \), for various \( \gamma \) and \( \psi \). For \( \psi = 0 \), the impulse is symmetrical; for \( \psi = \pm \pi/2 \) the impulse is antisymmetrical (after Červený et al., 1977).
Figure 5.2: Synthetic seismograms calculated with SEIS81. $f_M = 5$ Hz, dominant frequency; $\gamma = 4$, no of extrema in the source wavelet; $t_0 = 0$, time; $\psi = \pi / 2$, phase shift. $V_r$ is 8.0 km/s, the time shift of the envelope maximum is 0.3 sec.

method (Sheriff and Geldart, 1983).

An example of a synthetic record section computed using SEIS81 is shown in Fig. 5.2. Comments on this section can be found in Section 6.5.2.

Červený et al. (1977) discussed the validity conditions of the ray method. The ray method gives reliable results only for ‘thick’ layer velocity-depth models with large spatial changes of density, impedance and radii of curvature, due to the HF character of the wave field.

The ray field must be regular, i.e. in the region considered: 1) one and only one ray comes through every point and 2) there is a system of surfaces transverse to the rays in the considered region (existence of wave fronts). The geometrical spreading is related to the cross-sectional area of the elementary ray-tube. When the area approaches zero, as at a caustic, the amplitude of the ray field becomes infinite, and the ray field behaves
irregularly.

The source-receiver distance must not be too great. Červený et al. (1977) suggest that:

\[ L < \frac{l_o}{\lambda}, \]

where \( L \) is the length of the ray path and \( l_o \) is a measure of the inhomogeneity in the wave propagation direction.

Overall, 'the ray method, of course, cannot compete in the accuracy of computations with the reflectivity method' (Červený, 1985).

5.3 Reflectivity

Reflectivity computes the wave field as a whole, including multiple reflections, diffracted waves, etc. via a numerical integration of the plane wave reflection coefficients of the medium in the horizontal wavenumber or angle of incidence domain.

Multiplication of the assumed source spectrum and inverse Fourier transformation yields the synthetic seismograms (Fuchs and Müller, 1971).

The method is based on two concepts (Fuchs, 1980):

1. the response of a stack of horizontal, elastic layers to an incident harmonic wave of frequency \( (\omega) \) and angle of incidence \( (\gamma) \) which can be computed analytically.

2. the Sommerfeld integral (Treitel et al., 1982), which allows a spherical wave radiating from a point source to be decomposed into, or synthesised from, homogeneous plane waves radiating in all directions and a number of inhomogeneous waves travelling horizontally. A recomposition of the individual waves will produce the reflected and transmitted wave field generated by a point source.

The following steps are applied in the calculation of synthetic seismograms (Fuchs, 1980):

a. In the wavenumber-frequency domain the plane wave is propagated through the medium, undergoing changes in amplitude and phase shift, expressed by the complex reflectivity function (Fig. 5.3a):

\[ R_{pp}(\omega, \gamma) \]

which is due to the interference of all possible converted and multiply reflected waves, and is a function of the frequency \( (\omega) \) and the angle of incidence \( (\gamma) \).
CHAPTER 5. DYNAMIC MODELLING

Figure 5.3: a: Point source response over a layered medium (after Fuchs, 1980). b: Point source Q and observation point P over a layered medium. The compressional reflection from the reflecting zone suffers elastic transmission losses and time shifts in the layers 1 to m (after Fuchs and Müller, 1971).
The response at the surface is obtained by multiplying the plane waves by the corresponding reflectivity of the reflecting zone. The effect of the free surface and the two-way transmission in the overburden must also be taken into account.

b - The variable of integration is changed from horizontal wavenumber $k$ to angle of incidence $\gamma$ at the top of the reflecting layer:

\[ k = \frac{\omega}{c} = \frac{\omega}{\alpha_m} \sin \gamma = k_{0m} \sin \gamma \]

where $\alpha_m$ = P velocity of the $m$-th layer.

In this way the products of the transmission and the reflection coefficients at the free surface depend only on the angle of incidence. It is assumed that the angle of incidence, $\gamma$, is real.

c - The vertical displacement of a reflected P-wave from a point source recorded at the free surface is finally (for the derivation and details of the expression see Fuchs and Müller, 1971):

\[ v(r, 0, \omega) = \hat{F}(\omega)k_{0m}^2 \int_0^{\gamma_m} \sin \gamma \cos \gamma J_0(k_{0m}r \sin \gamma)R_{pp}(\omega, \gamma)g(\gamma)P(\gamma)e^{-2i\omega_{ms} \sum_{i=1}^{m} h_i} d\gamma \]

where (Fig. 5.3b):

- $r =$ offset
- $\omega =$ frequency
- $\hat{F}(\omega) =$ Fourier transform of the source wavelet
- $m =$ layers in the overburden
- $\alpha_i =$ P velocity of the $i$-th layer
- $J_0 =$ Bessel function of the first kind and zero order
- $h_i =$ thickness of the $i$-th layer
- $g(\gamma) =$ effects of the free surface
- $P(\gamma) =$ transmission coefficient for source and receiver regions
- $\exp( ) =$ phase effects due to source and receiver regions
- $j =$ imaginary unit
- $\eta_i = [\frac{\alpha_m}{\alpha_i} - \sin^2 \gamma]^{1/2}$

i.e. to integrate numerically the complex reflectivity function has to be computed (Fuchs, 1980):
over a range of angle of incidence \( \gamma_1, \gamma_2 \);

for body waves the range can be restricted between zero (vertical incidence) and \( \pi/2 \) (grazing incidence at the top of the reflection zone), or less (Fig. 5.4).

- at discrete angles with an increment \( d\gamma \),

- over a range of frequencies \( \omega_1, \omega_2 \), and

- at discrete frequencies with an increment \( d\omega \).

It follows that, if:

\[
N_a = \frac{\gamma_2 - \gamma_1}{d\gamma} \\
N_f = \frac{f_2 - f_1}{df}
\]  

(5.1)
where \( N_a \) = number of angles and \( N_f \) = number of frequencies for which the reflectivity has to be computed, the total number of reflection coefficients, \( n \), is given by:

\[
n = N_a N_f
\]

The computation of the reflectivity values requires considerable computing time, but once the values have been calculated they may be used for an arbitrary number of distances.

In this study the method proved useful for vertically inhomogeneous structures and large velocity contrasts.

For a solid model the computations with reflectivity 'can be assumed to be exact' (Fuchs and Müller, 1971). A direct comparison with the ray method, to check its approximations, showed that the evaluation of the amplitude near the critical point is the major difference between the two methods (Fuchs and Müller, 1971). For example, using a two-layer model in which a 0.6 km thick layer with a P-wave velocity equal to 1.8 km/s lies above a 2.5 km/s half-space, the ray method shows an incorrect abrupt increase in amplitude at about 1.2 km offset (Fig. 5.5).

The main limitation of the reflectivity method is that the earth is always assumed to be horizontally layered and laterally homogeneous, i.e. layering is allowed beneath source, receiver and reflection zone. It is therefore recommended that the reflectivity method be used to calculate the response over limited offset and for detailed studies and special features.

The reflectivity method has been used to model the amplitude of the \( a \) phase, the \( x \) phase and the S-wave field.

The program REFGMV from Karlsruhe University (Fuchs and Müller, 1971, further developed by Kennett, 1975; Kind, 1976; 1978) has been vectorised by Sandmeier (1984) to reduce the many hours of computation time required on a sequential computer for a complicated model with many layers, large source-receiver offsets and high-frequency sources.

The program has been successfully installed on the Fujitsu VPX240 Vector Processor of the University of Manchester Regional Computing Centre and the output files imported
Figure 5.5: Comparison of synthetic seismograms computed with the reflectivity method and the SRM. The model is two layered: $V_{p1}=1.8$ km/s, $V_{s1}=1.2$ km/s, $h_1=0.6$ km, $\rho_1=1.8$ g/cm$^3$, $V_{p2}=2.5$ km/s, $V_{s2}=1.25$ km/s, $h_2=\infty$, $\rho_2=2.0$ g/cm$^3$. a: reflectivity section, b: SRM section. Both sections are NMO corrected; the data are corrected for spherical divergence. c: amplitude versus distance plot.
via 'ftp' (file transfer protocol) to Leicester where the plot files, generated on the Geology Department MicroVAX, were produced.

The velocity model is defined by a stack of uniform layers of isotropic material. The effect of attenuation may be introduced in the program, with different values of Q for P- and S-waves being possible. The source can be defined as a point, line source or double couple. The source signal may be digitised, e.g., from the real data, or may be a spike, or more commonly a Fuchs-Müller signal (Müller, 1968).

The program requires the definition of:

1. parameters related to the frequency, duration and angle of incidence:
   - $\Delta t$, time increment in the source and in the seismogram, which determines the Nyquist frequency $f_{Ny}$.
   - $T$, duration of the response time series, large enough to include all the phases of interest. It is equal to the number of points (NPTS) multiplied by the time increment $\Delta t$.
   - Frequency and phase velocity window (see Fig. 5.4), determined by the selection of $\Delta t$ and $T$. i.e. if:
     \[
     f_{Ny} = \frac{1}{2\Delta t}, \quad \Delta f = \frac{1}{T} = \frac{1}{\Delta t \times NPTS} \quad \rightarrow \quad N_{FNY} = \frac{f_{Ny}}{\Delta f} = 0.5 \times NPTS
     \]
     where $N_{FNY}$ is the number of frequency points up to the Nyquist frequency.

     The phase velocity window is chosen to exclude the velocities of the overburden and to include the whole velocity range of the chosen reflecting layers, down to the desired depth. This avoids the generation of high-amplitude/low-velocity surface waves, or direct waves in the weathered layer.

   - $d\gamma$, angle of incidence increment. The condition to avoid numerical noise is (J. Mechie, pers. comm., 1995):
     \[
     d\gamma \leq \frac{1}{4f_{S}\sigma_{max}}
     \]
     where:
     - $\sigma_{max}$ = maximum distance
     - $f_{S} = \frac{1}{T}$, where $T$ = time duration of source signal.

     The most common value is 0.25°, with a possible range between 0.05° and 0.5°; larger values generate numerical noise (Fuchs, 1980).
Figure 5.6: Reduced travel time diagram for the reflecting zone of an arbitrary model. The theoretical seismograms have to be computed inside the rectangle (after Fuchs and Müller, 1971).

2 -reflecting zone, i.e. the time-distance range over which to compute the seismograms (Fig. 5.6). This requires the definition of overburden layers and reflecting layers, the modelling of transition layers and the flat earth approximation (for depths greater than 30 km, Fuchs, 1980).

Transition layers (positive or negative velocity gradient) are modelled as a stack of thin layers where the velocity jumps should generally not be larger than 0.2 km/s due to limitation in the reflectivity algorithm (Fuchs, 1980); however, the dominant frequency of the signal has to be taken into account. If the frequency is lower, thicker layers may be used.

The sphericity of the earth can not be neglected for wide-angle crustal rays sampling to a depth greater than c.30 km. The assumption upon which the reflectivity method is based, of flat homogenous layers, requires an increase in the velocity at a particular depth to compensate for the curvature of the earth. This transformation can be performed by REFGMV.
Amplitude studies are useful to reduce the range of possible models fitting the travel time data. Systematic studies (Banda et al., 1982) have shown that in particular is sensitive to different models of the continental upper crust, displaying distinctive patterns.

The amplitude of the phase has been studied as follows. Firstly, the true amplitude record sections have been plotted applying the conversion scale factors for the different instruments. However the amplitude range for the different instruments proved to be too large to be resolved with the RSEC90 program conversion factors. A homogenous data set was preferred. Only stations with USGS SGR instruments, which formed the biggest group, were therefore used, still providing a sufficient number of data points.

Secondly, interactive picking of the amplitudes using RSEC90 was performed. The faster automatic picking of amplitudes was preferred to the manual mode; analysis of the amplitude values from the two modes showed that the difference between them was small in relation to the scatter within either mode. The amplitude values were picked in counts and represent true maximum peak to peak values.

Thirdly, the true amplitude data files were transferred to a PC and plotted against distance on a logarithmic scale chart (Fig. 5.7-5.12). Conversion from counts to cm/s has been calculated using a factor equal to 2.5 x 10^8 determined following calibration of the USGS SGRs in the Pacific Northwest seismic experiment in September 1991 (W. Kohler, pers. comm., 1992).

There is a generally smooth variation with distance especially for the refracted rays within the basement. The largest scatter occurs within the Rift Valley, where the rays travel in sedimentary or volcanic layers.

The instrument response variation effect on the amplitude has been eliminated using one type of recorder only. Site response and topography of the sedimentary and basement layer together with the focusing effects from lateral heterogeneities can also affect the amplitude. Braile et al. (1994) estimate that the combination of these effects can result in variations of up to a factor of four, restricting the resolving power of the modelling; the variation is indicated by the vertical bar in the left corner in Fig. 5.7-5.12.

Amplitudes were modelled using REFGMV. Velocity gradients were approximated by
Figure 5.7: Observed and calculated amplitude distribution from shot point VIC. The vertical bar indicates a factor of four variation. The amplitudes in the current figure are corrected to an absolute particle velocity with units of cm/s.
Figure 5.8: Observed and calculated $a$ amplitude distribution from shot point KAP. The vertical bar indicates a factor of four variation. The amplitudes in the current figure are corrected to an absolute particle velocity with units of cm/s.
Figure 5.9: Observed and calculated amplitude distribution from shot point BAR. The vertical bar indicates a factor of four variation. The amplitudes in the current figure are corrected to an absolute particle velocity with units of cm/s.
Figure 5.10: Observed and calculated $a$ amplitude distribution from shot point TAN. The vertical bar indicates a factor of four variation. The amplitudes in the current figure are corrected to an absolute particle velocity with units of cm/s.
Figure 5.11: Observed and calculated $a$ amplitude distribution from shot point BAS. The vertical bar indicates a factor of four variation. The amplitudes in the current figure are corrected to an absolute particle velocity with units of cm/s.
Figure 5.12: Observed and calculated $a$ amplitude distribution from shot point CHF. The vertical bar indicates a factor of four variation. The amplitudes in the current figure are corrected to an absolute particle velocity with units of cm/s.
CHAPTER 5. DYNAMIC MODELLING

A stack of thin layers with a velocity increment of 0.01 km/s. The thicknesses of the layers were less than one wavelength of the maximum frequency considered in the source. Only vertical components have been modelled, using a point source at a depth of 0.05 km, radiating P, SV, and SH energy and a waveform equal to a Fuchs-Müller source wavelet.

The phase velocity interval used in the integration performed by the reflectivity program included all the compressional waves of interest propagating in the upper crust. High amplitude surface waves were excluded.

A combination of geometric spreading and absorption determines the amplitude decay with distance. The geometric spreading simulating amplitude decay for a point source is automatically taken into account by REFGMV. The resulting amplitudes are proportional to $r^n$ ($r=$distance source-receiver). For a reflected phase $n \approx -1$ and for a head wave $n \approx -2$.

For a diving wave in a gradient layer (e.g., phase $a$) probably $-1 < n < -2$; an average value of 1.5 will be assumed here.

At crustal ranges attenuation has little impact on the shape of the amplitude versus distance curve (Jacob et al., 1994). Observed amplitude studies have provided values for $n$ in the Kenyan crust: Braile in an unpublished report on the KRISP 85 data, measured $n \approx -1.6$, and Henry (1987), also studying KRISP 85 data, found $n \approx -1.7$. Amplitude variations in $a$ along the axial line from the LTC (Lake Turkana) shot to the south show a value of $n=-2.4$ between 10 and 90 km; analysis of the flank line data indicates that transmission is more effective outside the rift than inside, implying a smaller value of $n$ (Jacob et al., 1994).

Banda et al. (1982) have shown that the shape of the $a$ amplitude-distance curve (Fig. 5.13) calculated by reflectivity modelling is sensitive to the presence of sediments, which, in the model shown, introduce a maximum in the amplitude curve at a distance. An increase in the depth of the sediments increases this maximum amplitude near the shot. Decreasing the velocity contrast between cover and basement decreases the amplitude. Lower amplitudes are observed if a velocity gradient is introduced in the sediments. The curve is steeper if a higher gradient is present in the basement.

The frequency content of the source was derived from spectral analysis of the near shot traces. Source spectra with a dominant frequency of 10 Hz have been used for all shots.
Figure 5.13: Amplitude distance curve for model shown in the inset upper right with 0.0, 0.2, 2.0 and 5.0 km of low-velocity material overlying the basement (after Banda et al., 1982).
expt VIC, which required a value around 12 Hz.

Values of $Q_p$ equal to 100 for the sediments and 500 for the basement rocks, were chosen so as not to significantly affect the amplitude values.

An attempt was made to preserve the velocities of phase $a$ and the upper crustal thickness derived from the ray-tracing model. Changes were permitted where control was poor on the thickness and velocity of the sedimentary-volcanic cover. The 1-D velocity depth functions used for the modelling are reported in Fig. 5.14. The fit was considered satisfactory after the scatter resulting from the local site response and topography had been taken into account. Also theoretical amplitudes are corrected to an absolute particle velocity with units of cm/s.

VIC (Fig. 5.7): the amplitude best fitting model includes a basement layer, with a gradient of 0.03 $s^{-1}$ and $V_p$ equal to 5.9 to 6.2 km/s covered by 0.2 km of sediment (Fig. 5.14a). The high amplitude near the source proved difficult to model. Even reducing the thickness of the sedimentary basin, which lowers the amplitude value, does not provide a good near source fit. The data scatter allows a range between a $1/a^{1.5}$ and $1/a^2$ scaling with distance.

KAP (Fig. 5.8) shows a good fit with a 1.2 km thickness cover with a velocity of 4.40 km/s, over a basement with a gradient of 0.01 $s^{-1}$, and $V_p$ of 6.20 to 6.35 km/s (Fig. 5.14b). The data falls between the curves $1/a^{1.5}$ and $1/a^2$. The steepest decrease in observed amplitude toward the east, away from the rift, should be noted. This could be due to the absence of cover which would give a smoother decrease and produce a secondary maximum in the curve.

There is no suitable measure of the BAR source spectrum, since the near source seismograms are saturated. A source frequency of 10 Hz has been assumed equivalent to that of TAN, i.e. the other shot within the rift, which, despite being a borehole shot, gave a low-frequency source similar to an underwater shot (Jacob et al., 1994). A good fit with the general trend of the very scattered observed data is achieved (Fig. 5.9). The model is composed of three layers covering the basement which has a gradient of 0.02 $s^{-1}$ and $V_p$ velocity of 6.2 to 6.4 km/s (Fig. 5.14c). The observed data fall between the curves $1/a^{1.5}$ and $1/a^2$; they seem to show the same amount of scatter in both directions, as expected
Figure 5.14: 1-D velocity depth functions used for the $a$ amplitude modelling.
CHAPTER 5. DYNAMIC MODELLING

considering that BAR is surrounded by sedimentary/volcanic deposits.

The TAN amplitudes (Fig. 5.10) are consistent with a model including three layers in the sedimentary-volcanic cover, consistent with the 2-D model, and a gradient of 0.01 \( s^{-1} \) in the crystalline basement, which has a \( V_p \) of 6.4 to 6.5 km/s at 16 km depth (Fig. 5.14d). Again an area between the curves \( 1/x^{1.5} \) and \( 1/x^2 \) seems to cover the scatter in the observed data. The scattering observed at far offsets could be due to the volcanic cover present in both directions.

The BAS data (Fig. 5.11) have been modelled using a one layer model with a surface velocity of 5.85 km/s and a 0.01 \( s^{-1} \) gradient; the best fit involves a distance scaling of \( 1/x^2 \) (Fig. 5.14e). As for the KAP shot point, the observed data appears to give higher values at near offset (to 50 km) where no basement cover is present; in the case of BAS this direction is to the east.

For CHF (Fig. 5.12) a \( V_p \) velocity of 5.85 km/s has been used, without any vertical gradient (Fig. 5.14f); the data is best fit with a theoretical distance scaling of \( 1/x^2 \).

There are a number of comments:

Caution must be used in the interpretation of the amplitude modelling; as expressed by Banda et al. (1982) the conclusions are valid as long as the assumption of a laterally homogeneous model is valid. Focusing effects might completely mask the amplitude-distance behaviour. Bearing this in mind, reflectivity modelling can still provide a powerful tool to test the validity of the basement forward model.

Velocity values and structure of the basement cover used in the modelling seem to provide adequate fit between calculated and observed amplitudes. The velocity gradients presented in the final forward model can now be constrained to an accuracy of ±0.005 \( s^{-1} \). A value of 0.01 \( s^{-1} \) fits the calculated KAP, TAN and BAS amplitude decay curves with those observed; VIC requires a higher gradient, possibly indicating the presence of a mafic composition in the Archaean Nyamz Craton (ANC) Greenstone Belt. The \( a \) amplitude from BAR also is consistent with a slightly higher gradient of 0.02 \( s^{-1} \); as previously suggested (Section 3.4.2), the Mozambique Orogenic Belt (MOB) composition becomes more mafic at depth under the rift axis.

The rapid amplitude decay within the basement cover could be due to the thick ac-
cumulation of volcanic pyroclastic deposits. The contrast for the shots where the cover is absent in one direction (KAP to the west and BAS to the east) is striking.

The most important result arising from the $a$ amplitude modelling is the value of $n$, between $-1.5$ and $-2$. These values are higher than the ones measured along the axis south of Turkana ($n=-2.4$) (Jacob et al., 1994), and more consistent with the values measured by Henry (1987) in the central Kenya Rift. These measurements might indicate that in the central Kenya Rift energy transmission is better than in the northern Kenya Rift.

Commenting on the maximum distance at which energy is observed on the flank line and within the rift valley, Jacob et al. (1994), conclude that noise levels are generally higher in the rift than on the flank.

Studies of the $a$ amplitude elsewhere concentrate more on the effect of variations in gradient than on the possible variations in $n$. Examples can be found in Müller and Mueller (1979) on data from south-western Utah and Banda et al. (1982) on data from the Black Forest and the Basin and Range. Although a small sample, it is interesting to note that amongst these, the data set which provides the biggest scatter is that from the Basin and Range province, suggesting that tectonically active areas may produce the largest variability in the $a$ amplitude with distance.

5.3.2 Modelling the $x$ Phase

The $x$ phase has been detected on the P-wave record sections from BAR in both directions (Fig. 3.7, 3.8) and from TAN to the east (Fig. 3.10); also, on the S-wave data sets on the BAR record section in both directions (Fig. 4.6, 4.7) and TAN to the east (Fig. 4.9).

It has been previously suggested that it could be a precritical phase from an upper mantle reflection or a multiple phase.

Both options fit the travel time equally well ray-tracing through the 2-D model (for example Fig. 5.15a).

Finite-difference (FD) modelling using the final 2-D best fit model, generated a phase almost coincident with $x$, in time and amplitude distribution, especially on the TAN record section (Fig. 5.16a), which therefore will be described in detail.

To discriminate between the various possibilities, the data were modelled using re-
Figure 5.15: a: Ray-trace diagram for the $z$ phase showing the possible fit with a multiple in the mid-crust (1) and with a reflection from the uppermost mantle (2). b: Ray-trace diagram for the $x$ phase showing the fit with a multiple from the Moho interface in the basement cover.
Figure 5.16: a: Space-time window from the finite-difference record section calculated from TAN showing the theoretical $x$ phase. b: Space-time window from the reflectivity record section calculated from TAN showing the theoretical $x$ phase.
CHAPTER 5. DYNAMIC MODELLING

reflectivity introducing sequentially the response of each boundary until the phase under consideration appeared on the synthetic section (Fig 5.16b).

The synthetic sections will be presented for TAN; the same considerations appear valid for BAR. The model included a velocity-depth function extracted from the 2-D model in the vicinity of the shot point.

Full wave field seismograms, including conversions, were calculated.

A point explosive source at 30 m depth, dominant frequency 8 Hz, Fuchs-Müller signal was chosen.

Damping given by a $Q_p$ as low as 50 for the sediments and the upper volcanic unit, $Q_p=100$ for the lower volcanic unit and the crust and a representative value of $Q_p=1000$ for the upper mantle was required to enhance the $x$ phase. Normal values in absence of attenuation are considered $Q_p=2000$ and 500 with attenuation in the upper mantle (e.g., Kennett, 1977 interpreting data from a profile across France); $Q_p=450$ in the upper, 1000 in the middle and 2000 in the lower crust (e.g., Sandmeier and Wenzel, 1986 on data from the Black Forest in Southwest Germany).

The transition layers were constructed using thin layers with a 0.05 km/s increase in velocity across each layer boundary.

A precursor of $c$ at a lower apparent velocity than $c$ in fig. 5.16b may be a converted phase at the interface upper-lower crust.

The use of a 1-D method, reflectivity, obviously has to be taken into consideration in relation to fully 2-D method such as ray-tracing and FD. The differences between these methodologies will be greater for structures with complex lateral inhomogeneity. The difference in travel time for the $x$ phase between the FD section and the reflectivity section is therefore not significant. The presence of the $x$ phase on both is incontrovertible.

The $x$ phase from TAN could be identified as arising from the Moho, and multiply reflected within the rift infill (Fig. 5.15b). The calculated travel time for a two-way path in the basement cover, at the shot point and at the receiver, gives the necessary delay to the multiple phase with respect to the primary $c$.

Modelling by reflectivity showed that the $x$ phase could be explained using the structure determined by ray-tracing, and provided the additional information that the sed-
iments and volcanic units in the rift infill are associated with high attenuation. The complexity of the 2-D structure does not preclude the $x$ phase being due to a combined effect of diffraction, conversion, and multiple reflection.

5.3.3 Modelling the S-waves

An attempt has been made to model the vertical component S-wave field.

Numerical problems arising during the modelling of the S-waves with FD (Section 5.4) resulted in the S field being modelled using the reflectivity method.

During the modelling, the assumption was made that the S-waves seen on the vertical component were converted, therefore a buried explosive source was used. Conversion necessary to generate S-waves could occur at the lake bed (in the case of the underwater shots), at the surface, or at the sediment-basement, or volcanic-basement interface i.e. where a strong velocity contrast occurs.

However, the borehole shots generated primary S-waves. The underwater shots probably had the S-waves generated at the lake bed, and also these S-waves can, for practical purposes, be considered primary.

All the layers were included in the calculations, but only PP and SS reflections were considered; PP phases were included in the modelling to decrease the S amplitude on the normalised record section, in order to compare the theoretical to the observed seismograms.

Source dominant frequencies were derived from the power spectra of the near source seismograms as for the $a$ amplitude modelling (Section 5.3.1), the effective frequency band being between 0.0 and 12 Hz. The number of frequencies used was 1250; this value allowed a time span that included the S arrivals in the resulting seismogram.

Source depths used in the field (Table 2.2) were included in the models.

In the starting models $Q_p=200$ for the upper crust, and 500 in the lower crust and uppermost mantle. $Q_s$ was equal to 100 and 200 for the upper crust and deeper layers respectively. These values are comparable to the values used in the Basin and Range Province, where similar geophysical characteristics were found (Benz et al., 1990; Holbrook, 1990). However, $Q_p$ and $Q_s$ in the deepest part of the models are higher in their studies, equal to 1000 and 500 respectively.
The curvature of the earth was taken into account using the approximation by Müller (1977).

The time increment in the seismograms was between 0.015 and 0.02 s. $f_{M}$ being between 25 and 33 Hz allowed the low frequency S-wave field to be reproduced.

Examples of the modelling of $v$ for the cross-rift line from VIC (Section 4.3.2) and $u$ for the flank line from CHF (Section 4.4.3) have been previously discussed.

Modelling of the S-wave field for the cross-rift line for the five main shot points will be described in the following sections.

The starting models used involved velocity depth functions extracted from the 2-D best fit model near each shot point, derived from the finite-difference modelling of the P-wave field.

A maximum gradient of 0.02 $s^{-1}$ within each thin layer in the transition layers was found necessary to avoid reverberations in the S field.

The phase velocity range in the calculation was 2.0-15.0 km/s. These values include all the body wave velocities of interest in the model.

The main objective was to fit the amplitude ratios between the different S phases.

The travel time fit was considered of secondary importance because 1-D modelling in a structure with strong lateral variations cannot produce a satisfactory fit using a velocity-depth function derived from a 2-D model.

In the theoretical and observed plots (Fig. 5.17-5.21) the gain for S-waves is 1.4 times that for the P-waves, and the sections are all trace normalised. The P velocity-depth function is also plotted in each figure.

The seismogram sections illustrate one of the problems that may arise with the reflectivity modelling: the appearance of numerical phases with velocity equal to the maximum allowable in the window (indicated by an arrow in Fig. 5.19 for example).

Time-domain aliasing at far offsets has generally been suppressed by choosing an appropriate frequency increment $1/T_{T}$ to give a 'Nyquist time' $T_{T}$ larger than the response time.

The synthetic sections were generated with a trace spacing of 5 km. To allow a qualitative comparison of the recorded and theoretical sections, new SEG-Y files were created,
selecting stations with a c.5 km spacing. The same spatial and temporal window was plotted for each observed and synthetic section to ensure that the amplitude normalisation occurred on the same phases with a similar effect.

A more detailed description will be given for the individual shots.

**VIC**

$s$, $t_1$, $t_2$ and $u$ phases are detectable on the synthetic section (Fig. 5.17). The characteristics of the two clearest phases $s$ and $u$, i.e. the amplitude distribution of $s$ and the position of the $u$ critical point are adequately modelled. The amplitude of $t_1$ needed to be decreased from that produced in the starting model.

To decrease the $t_1$ amplitude the velocity contrast at 13.6 km depth has been decreased, resulting in an increased gradient in the upper crustal layer (the same has been done for the $P$ phase $a$). The thickness of sediments needed to be smaller than in the 2-D model, to fit the high $s$ amplitude observed at short range.

$Q_s$ in the lower crust and upper mantle was set equal to 500, to increase the amplitude of $t_2$ and $u$.

Excluding superficial layers from the model resulted in no S-wave generation, indicating that S-waves are generated by conversion from the water/sediments interface.

In the final S-wave reflectivity section the $v$ phase can be clearly seen, whereas, as discussed in Section 4.3.2 the noise level in the observed data is too high to detect any $v$ arrivals.

**KAP**

The most prominent phases, visible from KAP to the west on the S record section, are reproduced on the S synthetic section (Fig. 5.18).

The basement gradient of 0.01 $s^{-1}$ has been preserved from both the 2-D and the $a$ amplitude modelling.

In the lower crust and the upper mantle a value of $Q_s=500$ is required to increase the amplitude of $t_1$ and $u$.

The source has been positioned at 40 m depth; top layers were excluded from the
CHAPTER 5. DYNAMIC MODELLING

Figure 5.17: Top: synthetic seismograms calculated with the reflectivity method for the 1-D VIC model in the upper right. Bottom: trace-normalized band-pass filtered (5-10 Hz) S-wave record section for shot-point VIC recorded to the east along the cross-rift line D. $V_r=3.46$ km/s.
Figure 5.18: Top: synthetic seismograms calculated with the reflectivity method for the 1-D KAP model in the upper right. Bottom: trace-normalized band-pass filtered (4-10 Hz) S-wave record section for shot-point KAP recorded to the west along the cross-rift line D. $V_p=3.46$ km/s.
computation to avoid the generation of non required phases. To generate S-waves the volcanic-basement interface has to be included, and it is therefore the most likely surface where conversion from P- to S-waves occurs.

High amplitude Rayleigh waves and phase v are visible on the calculated record section but not on the observed.

**BAR**

The 1-D model composed of three layers covering the basement generated a complex pattern of arrivals and reverberations in the S-wave field, possibly due to generation of converted waves at both sediment-volcanic and volcanic-basement interfaces.

The exclusion of sediments from the model and restriction of the velocity window produced a simpler synthetic (Fig. 5.19); the s phase contains less energy than the original data at short offsets. t1 is possibly too energetic on the theoretical section, indicating that S-waves are attenuated beneath the rift.

A value of $Q_s$ as low as 200 was required in the lower crust and the upper mantle to model the low u amplitude, and can be explained by the attenuation effect of the rays travelling through the rift.

Generation of phase u requires the sediment-volcanic interface to be included in the model; the resulting section is not presented, because many other additional phases generated with x and not identifiable on the original data due to a low signal to noise ratio confuse the final picture.

**TAN**

The S phases identified on the record section to the west, t1 and u are reproduced on the synthetic section (Fig. 5.20).

As for BAR, the other rift shot, $Q_s=200$ in the lower crust and the upper mantle.

The gradient in the lower crust has been decreased with respect to the 2-D model.

The energy following the first arrival with 0.1 s delay between 30 and 90 km is also detectable on the recorded data.
Figure 5.19: Top: synthetic seismograms calculated with the reflectivity method for the 1-D BAR model in the upper left. The arrow indicates an unwanted numerical phase generated by reflectivity. Bottom: trace-normalized band-pass filtered (1-8 Hz) S-wave record section for shot-point BAR recorded to the west along the cross-rift line D. \( V_r \approx 3.46 \text{ km/s}. \)
Figure 5.20: Top: synthetic seismograms calculated with the reflectivity method for the 1-D TAN model in the upper left. Bottom: trace-normalized band-pass filtered (3-10 Hz) S-wave record section for shot-point TAN recorded to the west along the cross-rift line D. $V_p=3.46$ km/s.
CHAPTER 5. DYNAMIC MODELLING

BAS

A source at 50 m depth allowed the generation of S-waves via conversion at the free surface in a model where no strong velocity contrast is present near the surface, i.e. there is no basement cover.

In Fig. 5.21 s and u are of high amplitude, possibly due to the lack of sediments which could have had an attenuating/absorbing effect in the model. The amplitude ratios for the main phases s and u are well modelled.

Although no phases have been picked, energy is visible on the record section between 50 and 100 km and from 0 to 2 s reduced time and can be correlated with that seen on the theoretical section.

The main difference with the starting model is in the velocity-depth function which does not include the high velocity block.

No Rayleigh waves are generated due to the choice of a restricted velocity window.

There are a number of comments:

- The amplitude ratios between different S phases have been qualitatively modelled achieving generally a satisfactory fit with the observed data. 1-D velocity depth functions extracted from the 2-D ray traced S-wave model produced an adequate model for most of the shots;

- From VIC the large s amplitude at near offset suggests that the sedimentary basin modelled by ray-tracing should be shallower, but no systematic study has been undertaken for the s amplitude distribution.

The greatest misfit between theoretical and observed data was found in the amplitude values from the mid-crustal reflections, which in general were too large. An attempt was made to modify only the S velocity values, keeping the interface position fixed. $V_s$ gradients in the upper crustal layers were set higher than in the ray traced model in order to decrease the velocity contrast at the mid-crustal boundary;

- Lower $Q_s$ was used in the rift than outside it, although it is difficult to establish where the strongest attenuation occurs due to the lack of near offset S events for the shots within...
Figure 5.21: Top: synthetic seismograms calculated with the reflectivity method for the 1-D BAS model in the upper left. Bottom: trace-normalized band-pass filtered (3-12 Hz) S-wave record section for shot-point BAS recorded to the east along the cross-rift line D. $V_p=3.46$ km/s.
the rift;
- there has been little reflectivity modelling of S-waves in previous studies:

Sandmeier and Wenzel (1986) generated synthetic seismograms by the reflectivity method to model refraction data from Southwest Germany. S-waves are generated by conversion at the free surface. They used quality factors almost twice as high as in the present study. In order to model the strong difference in lower crustal observed reflectivity between P- and S-wave fields, a laminated lower crust was introduced for the P field and a homogeneous lower crust for the S field. They explained the difference by alternating high and low P velocity and high and low Poisson's ratio in the lower crust which result in strong P reflectivity and lack of S reflectivity.

A study involving the modelling of S-waves in Jordan used ray-theoretical seismogram calculations (El-Isa et al., 1987); it is interesting to note that they needed to decrease the velocity contrast at the mid-crustal level to diminish the theoretical $t_2$ amplitude. They suggested this might indicate a fine structure in the lower crust below 20 km, without need to change the lower crustal average velocity model.

Reflectivity seismograms presented by Holbrook et al. (1988), reproduce the absence of lower crustal S-wave reverberations from Southwest Germany for a realistic petrological model where the laminae are thicker than in the Sandmeier and Wenzel (1986) model. Alternating quartz-rich rocks and more mafic rocks account for a greater change in $V_p$ than $V_s$. Possible lithologies are gabbroic granulites and rocks of intermediate composition.

Although in the present case no reverberations in the lower crustal P phase are observed, the S mid-crustal phase is seen to have low amplitude. The presence of fluids at low pore pressure in the mid-crust could lower $\sigma$ and reduce the S amplitudes.

This could affect the amplitudes of the $u$ phase during their passage through the mid-crust. The lower crust shows a 'normal' reflectivity as demonstrated by the $t_2$ phase being clearer than $t_1$ on the only section, VIC, where both have been identified. Furthermore the $u$ phase amplitudes are only slightly lower than expected from the kinematic model, not justifying the presence of large scale modification (partial melt, fluids at high pore pressure, laminations, ductile shear zones) of the lower crust.
CHAPTER 5. DYNAMIC MODELLING

5.4 Finite-difference

The finite-difference technique for 2-D heterogeneous media enables a complete description of the seismic wave field. It accounts for the proper relative amplitudes of the arrivals including diving waves, converted waves, Rayleigh waves, diffractions from fault zones, head waves and waves from theoretical shadow zones. It can therefore enable the matching of calculated to observed times in regions where 2-D kinematic ray-tracing breaks down (Kelly et al., 1976).

The two dimensional partial differential wave equations are replaced by a suitable finite-difference representation which can be solved on a discrete spatial grid in a strictly numerical way, without physical approximation. The half-space is considered to be composed of a set of locally homogeneous lithologic regions, identified by constant density and elastic parameters (Kelly et al., 1976).

Travel times and amplitudes of waves in a medium (Kelly et al., 1976) may be described by an approximate finite-difference expansion of the elastic equations for each region. In the first case the equation of motion will model the complete waveform and create synthetic seismograms at any point in the model.

A regular or irregular grid of velocity values is therefore superimposed on the model. Boundaries are represented by velocity discontinuities within the grid.

Two second-order partial differential equations can be used to describe the motion of P-waves and SV-waves in a medium.

\( x \) and \( z \) are the horizontal and vertical rectangular coordinates in a 2-D medium, with \( z \) positive downward.

Then:

\[
\frac{\partial^2 u}{\partial t^2} = (\lambda + 2\mu) \left( \frac{\partial^2 u}{\partial x^2} + \frac{\partial^2 w}{\partial x \partial z} \right) + \mu \left( \frac{\partial^2 u}{\partial z^2} - \frac{\partial^2 w}{\partial x \partial z} \right) \tag{5.2}
\]

\[
\frac{\partial^2 w}{\partial t^2} = (\lambda + 2\mu) \left( \frac{\partial^2 w}{\partial x \partial z} + \frac{\partial^2 u}{\partial z^2} \right) + \mu \left( \frac{\partial^2 w}{\partial x^2} - \frac{\partial^2 u}{\partial x \partial z} \right) \tag{5.3}
\]

where:

\( u \) and \( w \) are the horizontal and vertical displacements,
\( \rho \) is the bulk density

\( t \) is the time

\( \lambda \) and \( \mu \) are the Lamé parameters for the medium. In particular, \( \mu \) is the shear modulus or rigidity, and \( \lambda = k - 2\mu/3 \) where \( k \) is the bulk modulus, or incompressibility.

Explicit finite-difference equations corresponding to (5.2) and (5.3) have been derived by Ottaviani (1971):

\[
\begin{align*}
\text{Equations (5.4) and (5.5) allow the } u \text{ and } w \text{ displacements at each spatial grid point at the time } (l+1) \text{ to be calculated, exclusively, from the particle motion in the two previous steps } l \text{ and } (l-1) \text{ (Kelly et al., 1976).}
\end{align*}
\]

In a medium where the interfaces are inclined or geometrically complex it becomes
difficult to use the homogeneous formulation. In particular, problems arise in attempting to satisfy the boundary conditions across interfaces where the Lamé parameters change.

In the heterogeneous formulation the model elastic parameters are allowed to change across the boundaries. The only limitation in the complexity of the model is the choice of the grid interval $\Delta x$ and $\Delta z$, which limit the resolution of the model.

A new set of second-order differential equations to describe $P-SV$ propagation in a 2-D medium can be written (Kolsky, 1963):

$$\rho \frac{\partial^2 u}{\partial t^2} = \frac{\partial}{\partial x} \left[ \lambda \left( \frac{\partial u}{\partial x} + \frac{\partial w}{\partial z} \right) + 2\mu \frac{\partial u}{\partial x} \right] + \frac{\partial}{\partial z} \left[ \mu \left( \frac{\partial w}{\partial x} + \frac{\partial u}{\partial z} \right) \right]$$

(5.6) and (5.7) reduce to the homogeneous case equations (5.2) and (5.3) when $\rho, \lambda, \mu$ are constant.

From (5.6) and (5.7) the approximation by finite-difference terms for a heterogeneous medium can be derived (Kelly et al., 1976).

A limitation of the method is the production of unwanted reflections from the edge of the model, whereas in the physical processes considered, waves pass through these boundaries. Boundary conditions which reduce edge reflections have been developed by Reynolds (1978) and implemented in the program used here. These transparent, non-reflecting boundary conditions are particularly stable in the acoustic case but less stable in the elastic case (Reynolds, 1978).

A disadvantage of finite-difference modelling compared to the synthetics derived from the ray method is that it is sometimes difficult to identify the phases generated, since the complete wave field is modelled. This problem can be overcome by stopping the procedure at different times, generating 'snapshots' to separate complex wave interactions and help to detect 'mystery events' in the observed seismograms. The effect of non-ray waves can also be demonstrated and explained (Fertig and Pšenicík, 1985).

The finite-difference technique was applied using the FDELAREF program, implemented by Sandmeier (1990), using the algorithm for 2-D heterogeneous media from Kelly et al. (1976) and transparent boundary conditions from Reynolds (1978). FDELAREF
was run on the Siemens S600 supercomputer in the computer centre at Karlsruhe University. The output files were transferred by ‘ftp’ and plotted at the Geophysical Institute at Karlsruhe University.

The finite-difference algorithm needs to be stable, i.e. the difference between the numerical and the exact solution of the finite-difference equation must remain limited with increase in the time index \( i \), with a constant time increment \( \Delta t \). This stability condition is:

\[
\Delta t \leq \frac{\Delta h}{\sqrt{V_p^2 + V_s^2}}
\]  

(5.8)

\( \Delta t \) is therefore defined by the choice of the grid interval \( \Delta h \) and by the maximum \( V_p \) and \( V_s \) in a particular homogeneous layer (Kelly et al., 1976).

To avoid grid dispersion, the number of grid-points can be calculated knowing the maximum offset and a grid spacing given by (J. Mechie, pers. comm., 1993):

\[
\Delta h \leq \frac{V_{min}}{10f_o}
\]  

(5.9)

where:

\( \Delta h \) is the grid spacing, and

\( f_o \) is the dominant frequency of the signal.

The 2-D velocity structure from the ray-tracing modelling is digitised with a grid spacing given by the relation (5.9). In the present calculations \( \Delta h \) was set equal to 40 m.

A point source radiating P, SV and SH energy was selected, and a Fuchs-Müller wavelet chosen. The source was positioned at the surface.

The dominant frequency for which the seismograms were calculated was 4 Hz, according to the stability conditions of the finite-difference modelling (5.9) and the analysis of the power spectra.

The free surface response was included, and Reynolds’ boundary conditions applied.

The seismograms were calculated with a time increment given by (5.8), and a duration necessary to include all the predicted P arrivals. \( \Delta t \) was 0.0039 for VIC and 0.0037 for all the other shots.
CHAPTER 5. DYNAMIC MODELLING

The seismogram spacing was 2 km and these were plotted with a reduced travel time scale and normalised amplitudes. Only the vertical component is presented.

There is general consistency between the observed and theoretical record sections; for the crustal phases a 0.1 s fit and for the mantle phases a 0.2 s fit are considered satisfactory. It is important to point out that FD may create energy/phases with travel times in advance of those produced by ray-tracing. This is due to the previously mentioned capacity of the method to generate non-ray waves (e.g., diffractions) which travel along different paths and interfere creating phases not seen on a ray traced section.

The P-wave field has been modelled. Time restrictions prevented modelling the S-waves at the Geophysical Institute in Karlsruhe. S-wave modelling with FD presents problems of instability on the grid after the long calculation times required for the longer S-wave arrival times. The calculation time challenges the CPU of the Siemens computer and requires particular tricks to fit with the maximum value of CPU time. Moreover, Rayleigh waves are also a bigger problem with S-waves than with P-waves because the Rayleigh wave velocity is close to the S-wave velocity. With FD it is not possible to exclude the response of the surface as it is when reflectivity modelling is used.

Fertig and Pšencík (1985) suggest the use of a time increment:

$$\Delta t = \frac{\Delta h}{V_{max} \sqrt{2}}$$  \hspace{1cm} (5.10)

to keep the computational scheme stable. They also propose a reduction in grid dispersion by choosing $\Delta h = 1/12$ of the dominant wavelength.

VIC - Fig. 5.22, 5.23

The model was ‘padded’ with an extra 3 km of extra-length near the source at the western end of the model, to avoid artificial reflections from the model edge.

Finite-difference record sections from VIC have been presented and commented on in Maguire et al. (1994). The dominant source frequency was slightly lower than the one used here (2.8 Hz for the crustal and 2.25 Hz for the mantle phases). Additional reverberations are visible in Fig. 5.22 (see e.g., a coda). The normalisation procedure only enhances the a phase over a short range, but the advance at about 180 km is clearly modelled and
Figure 5.22: Finite-difference synthetic seismogram section for shot point VIC recorded to the east along the cross-rift line D. Grid spacing 40 m. Dominant frequency 4.0 Hz. $V_p=6$ km/s. Correlated phases from ray-tracing as for Fig. 3.2 caption.
shows up better than with the ray-tracing method.

$b_i$ is visible, possibly with a higher amplitude and to a larger offset than on the observed data.

c is the dominant phase beyond c.100 km as in the observed data and transmits energy to at least 340 km on the FD section (Fig. 5.23), where no rays could be traced (Fig. 3.4: maximum ray traced distance is 220 km). On the observed data it can be followed to 260 km, where it becomes indistinguishable from the other crustal phases.

The $d$ advance between 240 and 280 km (Fig. 5.23) is well modelled by the FD calculation, whereas it was not visible in the ray-traced calculated branches.

The mantle reflector $d_i$ moves to a later travel time and the increased velocity contrast increases the phase amplitude compared to the section in Maguire et al. (1994).

**KAP - Fig. 5.24, 5.25**

The phase $a$ advance at 120 km to the west of KAP (Fig. 5.24) is modelled and followed by the delay introduced by the basin beneath VIC. To the east (Fig. 5.25) the first arrival is slightly early at about 25 km from the shot. The delays due to the Elgeyo escarpment and Kerio Valley and at 60 km due to the Baringo Basin are clearly reproduced.

$b_i$ arrival times and amplitudes seem consistent with the observed data between 70 and 150 km to the west.

c fits with the recorded phase between 120 and 150 km, but fits less well near the western end of the line. To the east $c$ is modelled 0.2 s earlier than the observed data.

$d$ is well modelled to the east to 220 km, and in advance beyond that distance.

A $d_i$ phase can be seen on Fig. 5.25 beyond 250 km, where no observed data have been picked (see Fig. 2.11).

**BAR - Fig. 5.26, 5.27**

To the west of BAR (Fig. 5.26) the delay over the Kerio Valley and the advance over the Elgeyo escarpment for phase $a$ are well modelled. Between 50 and 200 km $a$, and then $b$ (beyond 150 km) is 0.2 s early. This may be due to non-ray waves passing through a laterally complex structure, as for example to the west of BAR. To the east $a$ matches
Figure 5.23: Finite-difference synthetic seismogram section for shot point VIC recorded to the east along the cross-rift line D. Grid spacing 40 m. Dominant frequency 4.0 Hz. $V_p=8$ km/s. Correlated phases from ray-tracing as for Fig. 3.4 caption.
CHAPTER 5. DYNAMIC MODELLING

Figure 5.24: Finite-difference synthetic seismogram section for shot point KAP recorded to the west along the cross-rift line D. Grid spacing 40 m. Dominant frequency 4.0 Hz. $V_r=6$ km/s. Correlated phases from ray-tracing as for Fig. 3.5 caption.
Figure 5.25: Finite-difference synthetic seismogram section for shot point KAP recorded to the east along the cross-rift line D. Grid spacing 40 m. Dominant frequency 4.0 Hz. $V_r=6$ km/s. Correlated phases from ray-tracing as for Fig. 3.6 caption.
with the observed data, and is overtaken by \( b \) at 130 km (Fig. 5.27).

\( b_1 \) to the west appears only 0.1 s early relative to the real data and is consistent with these data to the east.

Phase \( c \)'s critical point to the west is correctly positioned but to the east it is 0.2 s early at about 150 km. The phase amplitude appears larger than in the real data, suggesting a more complex interface than the one modelled (Maguire et al., 1994).

The phase \( d \) cross-over is correctly positioned in both directions compared to the real data. To the east the phase is 0.2 s earlier than in the observed data.

\( x \): energy in the \( c \) coda is visible in both directions, although it is not possible to identify a particular phase coinciding with the one picked in the observed data. The identification has already been made with the aid of reflectivity modelling (Section 5.3.2).

No surface wave phase is as clearly visible on the BAR synthetic section as on the VIC synthetic section, being similar to the corresponding VIC and BAR observed sections.

**TAN - Fig. 5.28, 5.29**

\( a \) to the west reproduces the shape of the correlated recorded phase out to 130 km from the shot, including the advance at the Kamasia Hills, the delay due to the Kerio Valley, and the advance at the Elgeyo Escarpment (Fig. 5.28). To the east the phase is 0.2 s early from 50 km onwards, possibly due to the effect of the high velocity block through which no rays could be traced. The advance is still visible after the cross-over at 150 km in the \( b \) phase, to the east. This could imply that the lateral extension of the block to the east is smaller than on the ray traced model presented (Fig. 5.29).

\( b_1 \) to the west is visible from the critical point to around 90 km but it becomes clearer beyond 200 km, matching the observed data (Fig. 5.28). To the east it matches with the picked phase from 60 km onwards (Fig. 5.29).

The \( c \) phase critical point and travel time fit the observed data from 100 km to the west, becoming 0.2 s early at 150 km (Fig. 5.28). To the east the match is satisfactory (Fig. 5.29). The phase is somewhat stronger than on the record section; as for BAR, this might be due to the Moho interface being more complex than a single boundary with a large velocity increase.
Figure 5.26: Finite-difference synthetic seismogram section for shot point BAR recorded to the west along the cross-rift line D. Grid spacing 40 m. Dominant frequency 4.0 Hz. $V_r = 6$ km/s. Correlated phases from ray-tracing as for Fig. 3.7 caption.
Figure 5.27: Finite-difference synthetic seismogram section for shot point BAR recorded to the east along the cross-rift line D. Grid spacing 40 m. Dominant frequency 4.0 Hz. $V_r=6$ km/s. Correlated phases from ray-tracing as for Fig. 3.8 caption.
Figure 5.28: Finite-difference synthetic seismogram section for shot point TAN recorded to the west along the cross-rift line D. Grid spacing 40 m. Dominant frequency 4.0 Hz. $V_p=6$ km/s. Correlated phases from ray-tracing as for Fig. 3.9 caption.
Figure 5.29: Finite-difference synthetic seismogram section for shot point TAN recorded to the east along the cross-rift line D. Grid spacing 40 m. Dominant frequency 4.0 Hz. $V_r=6$ km/s. Correlated phases from ray-tracing as for Fig. 3.10 caption.
CHAPTER 5. DYNAMIC MODELLING

\[ d \] fits the picked phase to the west out to c.200 km, although beyond 180 km the data quality is poor; no rays were traced beyond 175 km (Fig. 5.28).

\[ z \] has been described and shown in Section 5.3.2. It is visible on the FD sections in both directions (to the east between 100 and 150 km, 2-3 s reduced time), but on the observed data it is clearer to the east while on the FD section it is clearer to the west (Fig. 5.28).

**BAS - Fig. 5.30, 5.31**

\[ a \] to the west (Fig. 5.30) reproduces the observed phase better than the ray-tracing: no rays could be traced beyond 100 km, whereas the FD first arrival fits with the observed data to 160 km.

To the east the amplitude beyond 60 km is higher than in the observed data, suggesting real attenuation is stronger than the theoretical (Fig. 5.31).

\[ b_1 \] to the west is well modelled between 50 and 100 km (Fig. 5.30). To the east two branches appear, between 40 and 60 km, and 70 to 100 km; this might be due to the presence of the eastern edge of the high velocity block, dipping eastward and flattening the reflected branch from its top (Fig. 5.31). Although it has not been ray-traced, \( b_1 \) does not show lateral continuity as on the observed data.

\[ c \] is adequately represented in both directions. To the east the precritical energy is modelled at shorter offsets than observed; however, only a few traces are available from BAS to the east on the data set.

\[ d \] matches the real data set up to 230 km to the west (Fig. 5.30), beyond which distance the observed data have not been picked.

High amplitude surface waves are visible on both the calculated and observed sections, as they were on the VIC sections.

There are again a number of comments to be made:

Finite-difference modelling of the P-wave field from the ray-traced model generally reproduces the observed features. This modelling technique is so time consuming that it
Figure 5.30: Finite-difference synthetic seismogram section for shot point BAS recorded to the west along the cross-rift line D. Grid spacing 40 m. Dominant frequency 4.0 Hz. $V_r=6$ km/s. Correlated phases from ray-tracing as for Fig. 3.11 caption.
Figure 5.31: Finite-difference synthetic seismogram section for shot point BAS recorded to the east along the cross-rift line D. Grid spacing 40 m. Dominant frequency 4.0 Hz. $V_p=6$ km/s. Correlated phases from ray-tracing as for Fig. 3.12 caption.
was not possible to run multiple iterations of the model.

Nevertheless, important observations can be made on the presented FD sections.

Travel times calculated from the FD sections are in agreement with the observed data. Some of the phases are earlier than the corresponding ray-traced phases as expected from the FD technique. This fact suggests that some of the interfaces should be deepened or the velocity decreased, as FD can be trusted to be a more rigorous modelling technique than ray-tracing. A number of important features of the model have been confirmed via these sections:

- the $a$ advance at 180 km from VIC is visible in the FD section,
- $a$ from BAS to the west is identified here to a longer offset than when ray-traced, consistent with the observed data,
- the $d$ advance in VIC around 260 km was not ray-traced but it is modelled here,
- $d$ from BAR to the west is generated here but it was not ray-traced,
- $d$ from TAN to the west is visible to a longer offset than when ray-traced, and
- $z$ is seen on the FD BAR and TAN sections and then successfully reproduced and interpreted with reflectivity.

However, on the other hand,

- $b_1$ from BAS to the east shows a different character on the FD section than on the observed one, and
- mid-crustal and Moho reflections seem to have an amplitude content too high compared to the observed data.

These combined observations give more confidence in the reliability of the model in terms of structural boundaries and layer thicknesses, but suggest that some of the reflecting boundaries might not be discrete discontinuities. They might be better modelled as the top or the bottom of zones of many individual discontinuous reflectors.

An alternative hypothesis could involve, as suggested for the S-wave reflectivity modelling, the presence of an attenuating mid-crust, affecting S-waves more than P-waves. This could be combined to a velocity gradient in the upper crust higher than in the ray-traced model; gradients larger than in the ray-traced model are also required by the a amplitude modelling.
The second interpretation is the preferred, because no reverberative character is seen in the reflected phases in question. Therefore it is difficult to sustain the first hypothesis, although it is always possible that only one of the reflectors at the supposed depth generates a detectable phase.

5.5 Summary

Dynamic modelling of both the P- and S-wave fields, mainly by reflectivity and finite-difference, has developed the model of physical properties beneath the cross-rift line.

Modelling the $a$ phase revealed that the amplitude decay for the cross-rift line falls off as $z^{-1.5}$ to $z^{-2}$. The smallest values occur outside the rift, near VIC and BAS to the east indicating a better energy propagation outside the rift. Also there is possibly better propagation in the central Kenya Rift than in the northern Kenya Rift. At VIC the higher $V_p$ gradient in the basement than for the other shots may be related to the presence of an Archaean Greenstone Belt. From BAR a smaller gradient than from VIC but higher than from the other shots might indicate the presence of mafic intrusions in the crust under the rift.

Moreover at VIC the value of $Q_p$ in the basement seems significantly higher than for the other data sets; amplitudes of the $a$ phase also appear higher from the western end of the line. Dynamic modelling by reflectivity for $a$ and S-waves suggests that the sedimentary basin modelled by ray-tracing at VIC should be shallower than in the ray-traced model presented in Fig. 3.23. Unfortunately surface waves near the source on the FD section prevent a good observation of the theoretical $a$ amplitude variation with offset in the presence of a sedimentary basin as modelled in Chapter 3. On the whole, however, it has to be accepted that the 1-D curve extracted from the 2-D model cannot easily predict sediment thickness, $Q_p$, and $V_p$ gradient in the basement in such a complex structure as that near VIC. The increase in velocity at c.3 km depth at about 10 km from VIC is not included in the 1-D function, and might account for the high amplitude seen on the observed data.

Modelling of the $x$ phase revealed that it most probably resulted, from the Moho
reflection being multiply reflected at the shallow sediment-volcanic and/or intra-volcanic boundary in the upper crust. The amplitude and frequency content of this phase suggest that strong attenuation occurs during the ray's transmission through the crust, or in the basement cover.

S-wave modelling demonstrated a lower $Q_s$ within the rift than outside it, although it is difficult to establish where the strongest attenuation occurs due to the lack of near offset S events for the shots within the rift.

Finite-difference modelling confirmed that some of the phases not ray-traced can be modelled through the 2-D structure produced, their travel times being consistent with the observed times.

Generally, mid-crustal and crust-mantle velocity contrasts from ray-tracing generates amplitudes which are too large, both for P- and S-wave synthetic sections but in particular for the S-waves. To account for the amplitudes observed, it has been suggested that fluids at low pore pressure in the mid-crust and a gradient in the upper crust higher than in the ray-traced model should be envisaged. The gradient is particularly high in the Archaean Nyanza Craton indicating a more mafic composition than in the Mozambique Orogenic Belt with the exception of the rift axis where a slightly higher gradient than outside the rift has been modelled. Again this may be explained in terms of compositional variations inside the MOB as suggested in Chapter 3.
Chapter 6

Analysis of the Mantle Phases

6.1 Introduction

One of the objectives of KRISP 90 was to detail the upper mantle structure beneath the Kenya Rift Valley, where the rifting process is still active.

It is generally accepted that rifting episodes are associated with magmatic activity whose origin resides in the lithospheric upper mantle or in the asthenosphere. Previous geophysical studies in Kenya have demonstrated large gravity and heat flow anomalies indicating the presence of anomalous mantle material or partial melt beneath the rift axis (Sections 1.5 and 2.3). A structural high has been modelled in the upper mantle from the interpretation of the cross-rift line D (Section 3.4.2). A close relationship has been suggested between asthenospheric upwelling and penetrative magmatism as driving forces for lithospheric rifting (Bhattacharji et al., 1987).

This Chapter describes and interprets phase 'd1', reflected from a discontinuity below the Moho recorded on the section from the Lake Victoria shot point (VIC). A paper extracted from this Chapter is enclosed as Appendix E. Discontinuities within the upper mantle in rift zones have also been identified elsewhere, for example in the Afar depression and the Red Sea (Mooney et al., 1985; Prodehl, 1985; Mochle et al., 1986; Makris and Ginzburg, 1987; Rihm et al., 1991).

The d1 phase has been the subject of further processing and analysis to define its polarity, in order to determine whether a low velocity zone (LVZ) exists in the upper mantle.
beneath the reflector. Study of lithosphere structure elsewhere (Hirn et al., 1971; Kind, 1974; Cassell and Fuchs, 1979; Burmakov et al., 1987; Ansoerge et al., 1992) has revealed thin layers of both high and low velocity explained either via changes in composition (Blundell, 1992) or physical state. The interpretation of the KRISP 90 axial line has revealed a low velocity layer between two high velocity layers at a depth between 50 to 65 km beneath the northern-central part of the profile (Fig. 2.2a after Keller et al., 1994a) and for this reason it was important to study the polarity of phase \( d_1 \).

If it is recognised that this phase has the same polarity as the first 'd' arrival, being the diving wave beneath the Moho, such polarity can be interpreted as resulting from an increase in velocity across the reflector. The results may be compared with those from a number of long profiles within rifts which have provided information from reflectors below the Moho (Keller et al., 1994a).

The seismic velocities derived from the present analysis have been compared with values determined from petrological modelling. The elastic properties of individual minerals can be used to define the seismic velocities at a certain pressure and temperature for a given assemblage of these minerals. Mineralogical compositions are derived from xenolith studies (Henjes-Kunst and Altherr, 1992; Fig. 1.12), and a possible geotherm for the continental crust is based on a conductive model after Pollack and Chapman (1977).

### 6.2 The Seismic Data

The upper mantle reflection \( d_1 \) can be seen in Fig. 3.4 and it is enhanced here in a 'window' extracted from the VIC record section (Fig. 6.1). Assuming an average velocity of 6.9 km/s for the shallow part of the model, a Normal Move Out (NMO) corrected record section has been plotted (Fig. 6.2) to show the character of the studied reflection. Using the assumption that the earth has a constant mean velocity above the reflector, the numerical procedure involved in hyperbolic moveout correction is presented. The travel time equation as a function of offset is:

\[
t^2(x) = t^2(0) + x^2/v^2
\]

where:
Figure 6.1: Record section for the Lake Victoria (VIC) shot point. $V_p=8.0$ km/s. Band pass filter 1-20 Hz. The traces in the shaded areas are stacked in Fig. 6.5, 6.6.
Figure 6.2: NMO corrected record section for the Lake Victoria (VIC) shot point. NMO constant velocity correction = 6.9 km/s.
\[ \Delta t_{NMO} = t(x) - t(0) = t(0) \left\{ \sqrt{1 + \left( \frac{x}{v_{NMO}(0)} \right)^2} - 1 \right\} \] (6.1)

The sections in Fig. 6.1, 6.2 are trace normalised, and band-pass frequency filtered using RSEC90 (Luetgert, 1990b), which also performed the NMO correction.

Seismic sections and slant-stacked traces showing the polarity relations between \( d \) and reflected phases from the uppermost mantle recorded along the axial line are presented in Fig. 6.3 from Keller et al. (1994a) for comparison with the data from the cross-rift line.

### 6.3 Methods

The investigated phase is observed at a distance of 270 to 350 km from Lake Victoria; the data quality required detailed processing to be undertaken before enabling comparison of the polarity of the reflection to the sub-Moho diving wave at the same offset. Across the reflector a decrease in velocity would produce a reflection of opposite polarity to the first arrival. With a velocity increase then the polarity should be the same as that of the first arrival as long as the reflection is still subcritical. Beyond the critical point the phase of the reflection starts to change and eventually at large distances it will again have opposite polarity to the first arrival. The distance over which the phase change occurs depends on the velocity contrast. For large velocity contrasts it changes over a fairly short range, whereas for mantle reflected events involving small velocity contrasts it usually occurs over a large distance.
CHAPTER 6. ANALYSIS OF THE MANTLE PHASES

Figure 6.3: Seismic records and stacked traces from the axial line. a: $d$ refraction and $d_1$ reflection; b: $d_1$ and $d_2$ reflections (after Keller et al., 1994a).
6.4 Procedure

6.4.1 Processing

RSEC90 (Luetgert, 1990b) enables band-pass frequency filtering to improve the quality of the recorded seismograms. This was designed to enhance the crustal phases. The $d_1$ phase was first detected on those trace normalised band-pass filtered sections and together with the preceding first arrival was extracted for further processing, in particular Correlated Adaptive Noise Cancellation (CANC) filtering (Hattingh, 1988) and subsequent stacking.

6.4.2 CANC Filtering

CANC was first developed by Hattingh in 1986 (Hattingh, 1988) as an extension of Adaptive Noise Cancelling (Widrow and Hoff, 1960; Widrow et al., 1975, 1976). CANC was initially designed to remove interference noise, random noise, and linear or non-linear drift from a time series using adaptive filters instead of fixed filters.

Fig. 6.4 (modified from Hattingh, 1988) illustrates the CANC procedure. The primary input $D$ contains the signal $S_o$ and the noise $n_o$. The reference input $X$ contains the same signal $S_i$ and noise $n_i$, which is assumed uncorrelated with $n_o$. The reference input is adaptively filtered to produce the filter output $Y$, which is a least-squares estimate of $D = S_o + n_o$. The filter output is subtracted from the primary input to give the system output or error $E$. The error is then fed back into the adaptive filter and used to adjust the filter weights. Since $n_o$ is uncorrelated with the filter input $X = S_i + n_i$, the filter output is a least-squares estimate of the signal $S$ (Kirk, 1989).

Following Hattingh (1988) and Kirk (1989), the CANC algorithm can be described by:

The two input signals given by the vectors:

$$Primary \quad D = [d_1, d_2, ..., d_n]$$

$$Reference \quad X = [x_1, x_2, ..., x_n]$$

The filter input is a tapped delay-line version of the reference input, the delay being equal to the filter length $L$. At the $i$-th instant the reference signal is given by the vector:
$X_x(i) = [x_i, x_{i-1}, \ldots, x_{i-L+1}]$

where $L \leq i \leq n$.

The filter coefficients at the same $i$-th instant are:

$W(i) = [w_{1i}, w_{2i}, \ldots, w_{Li}]$

The filter is applied to the tapped delay line vector $X_x$, the filter input, to give the filter output:

$Y(i) = W^T(i)X_x(i)$

As the filter uses past data to estimate a centre value, the filter output $Y$ is subtracted from a delayed version of the primary to give the system output or error $E$. The delay is half the filter length, $L$. Therefore, $L/2$ points at the beginning and at the end of the series are unusable and are set to zero in this algorithm:
E(t) = d(t - L/2) - y(t)

As indicated above, the error signal is then fed back into the adaptive filter and used to calculate the next set of filter coefficients by the Widrow-Hoff Least Mean Squares algorithm (LMS). This is performed by minimising the mean square error between the output Y(i) and the primary signal. The process is repeated down the whole series until convergence is reached. The last filter weights are taken and used as starting values for the next iteration. The filter output is then an approximation of the underlying signal in the primary input. The error is the noise that has been removed from the primary signal.

The filter requires various input parameters.

L, the filter length. Tests on synthetic data (Kirk, 1989) showed that the best signal recovery is provided by filter length between 20 and 40 samples and small convergence rate. Values from 25 to 38 samples were used here.

\( \mu \), the convergence rate. Long filters with small convergence rate provide the greatest signal resolution (Kirk, 1989). Values were set between 0.001 and 0.0001.

I, the number of iterations. The CPU time required increases with increase in I. Between 20 and 40 iterations were sufficient.

CANC filtering of wide angle seismic data is ideal for enhancing separate signals from one shot point recorded at one recording site. In the present case with signals recorded at a large distance from a single shot, correlation is assumed between the recorded phase at two adjacent recorders (spacing 0.3 - 3.8 km) but with uncorrelated noise.

CANC filtering can perform satisfactorily if there is a small phase shift between reference and input signal (Kirk, 1989). As Kirk (1989) suggests from synthetic data tests, the noisier signal should be on the primary input. Referring to Fig. 6.4 it is possible to see that the reference input is adaptively filtered to become similar to the primary input. The system output power will be high if the noise power on the primary input signal is high.

Where possible, the signal closest to the shot was used as the reference signal. For very noisy signals, the reference and input signal were reversed to improve the output quality.
The $d$ phase was analysed within a 1 second (200 samples) time window starting at 6 sec reduced time (for a reducing velocity of 8.0 km/s). The $d_1$ phase was analysed within a window of 2 seconds length.

The results are presented for selected traces for ranges between 270 and 350 km from the VIC shot point (Fig. 6.5 and 6.6).

The derived cross-correlation function between the time windowed CANC filtered traces provides an estimate of the necessary time shift to achieve the greatest similarity. The accepted time shift was always less then a wavelet cycle. Each CANC filtered pair of traces has been cross-correlated with the input traces and only those with a cross-correlation $> 60\%$ were accepted.

Following CANC filtering and the subsequent time shifting, the best data sets were stacked. Two examples of the filtered and resultant stacked seismograms are shown for location numbers 68 to 74 (269-274 km) and 31 to 42 (320-345 km) for the $d_1$ phase in Fig. 6.5a and 6.6a and for the same range of location numbers for the $d$ phase in Fig. 6.5b and 6.6b. The chosen locations are expected to be in the critical and precritical range, the reflected phase having opposite polarity to the first arrival in the distant overcritical range. The stacked seismogram for each group of traces is shown at the right hand side of each figure.

For each location number group the final output shows a $d_1$ phase with the same polarity as the first arrival $d$. In order to quantify the similarity in each group between the $d$ and the $d_1$ phases, cross-correlation has been undertaken between $d$ and $d_1$ as well as between $d$ and an inverted $d_1$ phase for each group of location numbers. An example (location numbers 27 to 42 for $d$ and 27 to 45 for $d_1$) is shown in Fig. 6.7.

The measures of similarity suggest that the optimum result is achieved when cross-correlating the $d$ phase with the $d_1$ phase rather than the inverted $d_1$ phase i.e. the polarity of the two phases may be taken to be the same, and the $d_1$ phase results from an increase in acoustic impedance in the uppermost mantle.
Figure 6.5: a: CANC output seismograms for location numbers 31 to 42 and their stack for the $d_1$ phase. b: CANC output seismograms for location numbers 31 to 42 and their stack for the $d$ phase.
Figure 6.6: a: CANC output seismograms for location numbers 68 to 74 and their stack for the $d_1$ phase. b: CANC output seismograms for location numbers 68 to 74 and their stack for the $d$ phase. It should be noted that the filter length may be different and in some cases the traces used are not the same for $d$ and $d_1$. 
Figure 6.7: Example of CANC output stacked seismograms $d$ cross-correlated with a time shifted $d_1$ (location number 27 to 45) top: $d$ with a normal polarity $d_1$; bottom: $d$ with an inverted polarity $d_1$. The maximum amplitude of the cross-correlation function is indicated.
CHAPTER 6. ANALYSIS OF THE MANTLE PHASES

6.5 Interpretation

6.5.1 Ray-Tracing

Prior to the polarity analysis \( d_1 \) was modelled via ray-tracing using a reflector at a depth of 55 km beneath the western margin of the rift (between about 140 to 160 km from VIC) to generate the high amplitude phase \( d_1 \) immediately after phase \( d \) on the VIC record section. A reflector at a depth of about 50 km is also identified beneath the flank line (Prodehl et al., 1994a).

Maguire et al. (1994) arbitrarily suggested the reflection resulted from a positive impedance contrast, consistent with the present flank profile model. Beneath the axis of the rift the anomalously low velocity material immediately beneath the Moho was modelled extending to a depth of at least 60 km. This is consistent with the axial profile model which identified a boundary with a small impedance contrast at this depth south of Lake Baringo (Keller et al., 1994a). It is also consistent with the recent model derived from teleseismic data (Green et al., 1991). On further detailed examination of the VIC data set it is possible that the reflector identified beneath the flank of the rift may continue beneath the rift. This reflector would be equivalent to the one modelled at about 60 km depth along the axial line from the \( d_2 \) phase in Fig. 6.3 (Keller et al., 1994a).

The present new evidence increases confidence in the picking of both \( d \) and \( d_1 \). \( d_1 \) appears to occur at about 7.5 sec reduced travel time, \((V_s=8.0 \text{ km/s})\) between 280 and at least 350 km from VIC (Fig. 6.1). The \( d_1 \) picks obtained in the present study are somewhat later than those shown in Fig. 3.4 which has necessitated the \( d_1 \) reflector being placed at greater depth; in Fig. 6.1 the revised picks appear.

The revised picked phases \( d \) and \( d_1 \) were again modelled using the MacRay package (Lustgert, 1992b). The velocity structure above 55 km depth has not been changed from the model in Fig. 3.22. Perturbation of the models has shown that once the phase correlation has been fixed, resolution of the relevant velocity and depth to interface may be accepted as better than 1% and 5% respectively.

Two different models are kinematically acceptable; one with and one without the presence of a LVZ (Fig. 6.8, 6.9). In the former case the reflection arises from the base of
At about 350 km distance the travel time difference between the \( d \) and \( d_1 \) phases is about 0.5 sec and the two phases have very similar apparent velocities. For this reason and in order to move the theoretical critical point to around 280 km distance, a LVZ was introduced. In this case the observed travel times and amplitudes can be matched using a velocity of 8.4 km/s under the reflector together with a 4 km thick LVZ between 55 and 59 km depth and an average velocity in the LVZ of 7.7 km/s. The velocity in the LVZ is taken from the axial line model.

However, the observed travel times and amplitudes can also be fitted within the error limits without the necessity of a LVZ. In this case a velocity of 8.65 km/s beneath the reflector which should now be at 63 km depth, is required. The resultant theoretical critical point is at 304 km distance from VIC. Correcting for the sphericity of the earth (Mereu, 1967) the velocity below the reflector decreases to 8.56 km/s.

Both models fit the travel times equally well. In both cases the ray coverage on the reflector is limited to the distance range 100-180 km from VIC i.e. to the west of the surface expression of the rift. In the next section the amplitude ratios between the \( d \) and \( d_1 \) phases will be used to try to discriminate between the two models.

### 6.5.2 Synthetic Seismogram Modelling

Synthetic seismograms for the two seismic models were computed using two methods: the ray-method (Červený et al., 1977) and finite-difference modelling (Kelly et al., 1970; Sandmeier, 1990).

The synthetic record section in Fig. 5.2 was computed using SEIS81, for the model without the LVZ. The wave field was generated by a point source at zero depth situated in the 2-D velocity model resulting from MacRay ray-tracing. The velocity field approximation in any layer for MacRay is derived from a linear interpolation between values defined at nodes of a net whose upper and lower boundaries are given by the structural boundaries. The SEIS81 model velocity distribution is given by bicubic spline interpolation of values in a net whose upper and lower boundaries are horizontal and including the relevant layer. Nevertheless the estimated difference in travel times resulting from the two
Figure 6.8: Ray-trace diagram for the intra-mantle reflection $d_1$ for a model with a LVZ (bottom) and corresponding travel time diagram (top).
Figure 6.9: Ray-trace diagram for the intra-mantle reflection $d_1$ for a model without a LVZ (bottom) and corresponding travel time diagram (top).
CHAPTER 6. ANALYSIS OF THE MANTLE PHASES

parameterizations is less than 0.1 sec.

The theoretical seismograms were computed using a normalized amplitude scaling. For completeness the section obtained with SEIS81 has been included, but as pointed out above the calculation by finite differences gives more reliable values.

The synthetic record sections in Fig. 6.10 were computed by a finite-difference technique for 2-D heterogeneous elastic media.

The ray-traced model was digitized with a grid spacing of 40 m for the calculations. The dominant frequency for which the seismograms were calculated was 4 Hz, according to the stability conditions of the finite difference modelling (Kelly et al., 1976) and the analysis of the power spectra for the mantle phases. This analysis showed that relatively high frequencies (up to 10 Hz) are preserved beyond the anomalous upper mantle, which has no strong damping effect on the P-waves.

The top section arises from the model with the LVZ, the bottom section from the model without the LVZ. Both adequately model the arrival time and polarity of \( d \) and \( d_1 \) in the range 270-330 km. Beyond that distance \( d_1 \) merges with \( d \) in the model without the LVZ indicating that the velocity underneath the reflector is possibly too high. An additional phase with inverse polarity appears as a precursor of \( d_1 \) from the model with LVZ, which is not seen on the recorded section; therefore, a model with the LVZ should include a negative gradient instead of a sharp discontinuity on top of the LVZ. Time constraints did not allow many iterations using the FD technique, but the sections suggest confirmation of the validity of the two models.

6.5.3 Amplitude Ratio versus Offset Calculation

The maximum and minimum values of the observed amplitudes were automatically measured for the chosen phase within a selected time window. Due to considerable scatter of the observed amplitude data, probably caused by varying site response at the recording stations, the use of amplitude ratios of the two phases instead of 'true' amplitudes, was preferred (\( d_1/d \)).

A comparison of the amplitude ratios of the synthetic seismograms using the two velocity functions (with and without the LVZ) shows that the model with the LVZ produces
Figure 6.10: Synthetic seismograms for $d$ and $d_1$ phases for the model with the LVZ (top) and without the LVZ (bottom) calculated by using a finite-difference technique (Sandmeier, 1990).
amplitude ratios slightly higher than the observed ones while the model without the LVZ produces values somewhat too low before 280 km distance. The $d$ and $d_1$ phases apparently merge beyond 330 km from VIC on the synthetic section. It should be noted that these amplitude ratios have been derived using isotropic layered models. If, as is proposed below, the high velocity layer at about 60 km depth is anisotropic with preferred orientation of olivine crystals then the amplitude ratios shown here may be up to 20% in error in the undercritical range and up to 5% in error in the overcritical range.

Thus the two models fit the observed amplitude ratios with more or less equal quality (Fig. 6.11). Considering the difficulty in defining the exact position of the critical point for the $d_1$ phase the calculated amplitude ratios lie within the scatter of the observed ratios. For the finite-difference model without the LVZ the critical point is at about 300 km from VIC, for the one with the LVZ it is at about 280 km from VIC.

An eastward dip on the reflector of 5° would move the critical point by no more than 30 km, still achieving a satisfactory travel time fit. It is apparent that a range of models between the two end members presented can be considered to fit the travel time and amplitude data. Appropriate variations in the thickness of the LVZ, depth of the interfaces, and seismic velocities may achieve the desired fit.

Although it is therefore not possible to exclude the presence of a LVZ, the seismic modelling does constrain the velocity underneath the reflector at about $8.4 \pm 0.1 \text{ km/s}$, at a depth of $59.0 \pm 2.0 \text{ km}$ in the model with the LVZ, and at about $8.6 \pm 0.1 \text{ km/s}$, at a depth of $63.0 \pm 2.0 \text{ km}$ in the model without the LVZ. As mentioned in Section 6.5.1 it is possible that the reflector continues across the rift but the velocity underneath it along the axis (7.8-7.9 km/s) is less than the one found here. In either case the relatively high velocity beneath the flank of the rift adjacent to lower velocities underneath the rift axis needs to be explained. The possible presence of a LVZ has also to be discussed.

**6.5.4 Petrological Modelling**

A technique that allows the evaluation of the seismic velocity of a rock at a known pressure and temperature from estimated elastic properties of individual minerals is described in detail in Fuchs (1983) and Mechie et al. (1994a).
Figure 6.11: Comparison of observed and theoretical amplitude ratios for the model with and without the LVZ.
The velocity ($V$) of an isotropic rock at a given pressure ($P$) and temperature ($T$) based on a net travel time argument can be derived from (Fuchs, 1983):

$$\frac{1}{V} = \sum_{i=1}^{n} a_i/v_i$$

where:

- $n$ = number of minerals in the rock
- $a_i$ = volume percent of the $i$th mineral
- $v_i$ = P- or S-wave velocity of the $i$th mineral at pressure $P$ and temperature $T$.

The basic travel time argument for isotropic rocks considered here gives velocities within the Voigt-Reuss-Hill (VRH) average, an arithmetic mean of the Voigt and Reuss averages (Watt et al., 1976). If some olivine is preferentially oriented and the rest of the rock is isotropic, for the compositions considered here up to 0.04 km/s difference might exist between the net travel time argument velocities and the VRH average (Mechie et al., 1994a).

Knowing the mineral composition of a rock and in situ temperature and pressure, then P- and S-wave velocities can be calculated. This method has been applied here to model the seismic velocity in the uppermost mantle underneath the western flank of the rift valley.

The surface temperature and pressure (STP) elastic properties and their partial derivatives with respect to pressure and temperature for the minerals used here can be found in Mechie et al. (1994a). They estimate the error in the theoretical velocities to be within c.0.05 km/s of the correct value from the error analyses they carried out.

The temperature-depth values are estimated from Pollack and Chapman (1977) for a continental geotherm. Normal heat flow values, c.50 mWm$^{-2}$, are found on the rift flanks in Kenya with high but very variable values, up to c.100 mWm$^{-2}$, being limited to the rift valley region (Morgan, 1982; Nyblade et al., 1990; Wheildon et al., 1994). An average value of 40 mWm$^{-2}$ was found on the western shoulder of the rift valley, north of the Nyanza Rift (Wheildon et al., 1994).

The pressure $P$ can be estimated (Mechie et al., 1994a) at a certain depth $z$ integrating the density-depth profile ($\rho(z)$):
$P = g \int \rho(z)dz$

where $g = 9.81 \text{ m/s}^2$. The density profile is calculated from a velocity depth function derived from the seismic/gravity model beneath the western rift flank as 17.1 kbar at 59 km depth or 18.4 kbar at 63 km depth. Errors caused by uncertainty in the pressure are thought unlikely to be significant as a change of 2 kbar (equivalent to about 6 km change in depth) produces only a small change of about 0.02 km/s in P-wave velocity. In contrast a 100°C change in temperature produces a change of about 0.05 km/s in P-wave velocity.

The average mantle composition, P1 (Fig. 6.12) has been derived from xenoliths brought up by Quaternary volcanoes (Henjes-Kunst and Altherr, 1992) from the eastern shoulder, because mantle xenoliths from the rift itself are unknown; this composition has been assumed to model the uppermost mantle under the rift (Mechie et al., 1994a), and it seems reasonable to extend the assumption under the western flank. P1 corresponds to an undepleted spinel peridotite at minimum olivine content with 50% olivine, 30% orthopyroxene, 18% clinopyroxene and 2% spinel. At 63 km depth some garnet may be present at the expense of spinel, and taking into account that the spinel-garnet transition is likely to occur over a wide interval, a garnet peridotite, P3 (Fig. 6.13) including 55% olivine, 23% orthopyroxene, 14% clinopyroxene and 8% garnet has also been modelled.

Combining a surface heat flow of 40-50 mWm$^{-2}$ with the Pollack and Chapman (1977) geotherms leads to temperatures of around 600°C at about 60 km depth. Even for such a low temperature, isotropic pure olivine only has a velocity of 8.05 and 8.15 km/s for P1 and P3 respectively (Fig. 6.12a and 6.13a). However the velocity of either 8.4 km/s at 59 km depth or 8.6 km/s at 63 km depth beneath the western rift flank can be explained if preferred mineral orientation (anisotropy) is invoked.

Mechanisms like shearing possibly accompanied by recrystallization tend to orient the 'b' axis of olivine perpendicular to the plane of the flow. Flow and recrystallization may randomly orientate 'a' and 'c' axes in the plane of the flow (Nicolas et al., 1971).

Involving anisotropy, either the transverse isotropic olivine model or the orthorhombic olivine model is possible. In the case of the transverse isotropic model the faster 'a' and
CHAPTER 6. ANALYSIS OF THE MANTLE PHASES

Figure 6.12: a: Ternary diagram showing the relationship between P-wave velocity and mineralogical composition PI for: a: a pure isotropic model at 63 km depth. b: a transverse isotropic model at 63 km depth. Note that although P1 does not have velocities exceeding 8.35 km/s compositions with 75-85% olivine have velocities of about 8.6 km/s. c: a transverse isotropic model at 59 km depth. Note the similarity in the velocity for P1 between Fig. 6.12b and 6.12c.
Figure 6.13: Ternary diagrams showing the relationship between P-wave velocity and mineralogical composition P3 for: a: a pure isotropic model at 63 km depth. b: a transverse isotropic model at 63 km depth. Note that although P3 does not have velocities exceeding 8.45 km/s compositions with 70-80% olivine have velocities of around 8.6 km/s. c: a transverse isotropic model at 59 km depth. Note the similarity in the velocity for P3 between Fig. 6.13b and 6.13c.
'c' axes of the olivine should be randomly orientated in the horizontal plane. Although at temperatures as low as 600° C at around 60 km depth both P1 and P3 have velocities of 8.3-8.4 km/s, compositions richer in olivine can have velocities up to 8.6 km/s even at temperatures up to 800° C (Fig. 6.12b-c and 6.13b-c).

In this respect it is interesting to note that kimberlites of uncertain age near the western end of the cross-rift line have brought up ultramafic xenoliths including olivine-rich harzburgites and dunites (Ito, 1986).

In the case of the orthorhombic model, the fast 'a' axis of the olivine should be orientated along the direction of the cross-rift line.

In the case of the model without a LVZ for both P1 and P3 at temperatures between 600° and 800° C at 63 km depth, more than 75% of the olivine should be preferentially orientated in order to attain a velocity of 8.6 km/s. For the model with a LVZ more than 50% of the olivine should be preferentially orientated in order to attain a velocity of 8.4 km/s for both P1 and P3 at temperatures between 600° and 800° C at 59 km depth. For compositions richer in olivine a smaller amount of the mineral is required to be preferentially orientated.

Keller et al. (1994a) and Mechie et al. (1994a) have also invoked anisotropy to explain high velocity layers at 40-45 km and 60-65 km depth in the mantle below the axis of the northern part of the Kenya Rift. Their high velocities could be explained either by a transverse isotropic model similar to that utilized here or by an orthorhombic model in which the fast 'a' axis of the olivine crystals is orientated along the rift axis. If the layer identified at 60-65 km depth beneath the rift axis and the layer identified here have a common origin, then either a transverse isotropic model or an orthorhombic model with the fast 'a' axis of olivine orientated radially away from the apex of the Kenya Dome could account for the observed seismic velocities. In either case the cause of the anisotropy could be shear stress along more or less horizontal planes perpendicular to the slow 'b' axis of olivine (Nicolas et al., 1971) and possibly directed radially away from the apex of the Kenya Dome.
6.6 Discussion

The upper mantle seismic velocity distribution beneath the western flank of the rift derived here is supported by both kinematic and dynamic modelling of the seismic wavefield and petrological modelling.

The proposed best-fitting model is shown in Fig. 6.14a. An alternative model can be envisaged with a LVZ between 55 and 59 km, and 8.4 km/s velocity underneath the LVZ (Fig. 6.8). The implications are similar. For comparison the axial line model is also shown in Fig. 6.14b. The reflector modelled at 63 km depth marks an increase in velocity to 8.6 km/s. This velocity is consistent with temperatures of 600 – 800°C, a pressure of 18.4 kbar and a petrological model involving the presence of garnet and preferred orientation (anisotropy) of the olivine crystals either in a transverse isotropic or orthorhombic structure.

There are four points of note:

(A) An upper mantle Low Velocity Zone.

The presence of a Low Velocity Zone in the uppermost mantle underneath the western flank of the rift can not be excluded. Low velocity layers have been identified in the upper mantle beneath the axial line. The velocity inside the LVZ beneath the cross-rift line can not be constrained because there is no reflection (with inverted polarity) seen from the top of such a zone.

The finite-difference synthetic seismogram modelling provides similar amplitude ratios for the models with and without a LVZ.

Petrological modelling requires relatively low temperatures that do not suggest the presence of melt at depth, but cannot exclude its presence, laterally penetrating from the axial zone of the rift.

Alternating high and low velocity layers have been found beneath the axis of the rift, where other geophysical results suggest the presence of anomalous mantle and partial melt. The low velocity layers are present below the northern part of the rift and do not extend as far south as the cross-rift line.

Elsewhere beneath rifts or tectonically active areas alternating high and low velocity
Figure 6.14: Velocity model of the cross-rift (top) and axial (bottom - distance scale halved) profiles derived from ray-trace and synthetic seismograms modelling. P-wave velocities in km/s. The reflector at a depth of c.60 km beneath the axis of the rift on the cross-rift line is obtained from the axial line model.
layers have also been found (Prodehl, 1981; Prodehl et al., 1992; Nagumo et al., 1981).

The following considerations are valid assuming that the reflector seen under the western flank of the rift does not extend under the rift axis, and therefore the high velocity is limited to occur under the western flank of the rift:

- if the model with the LVZ is accepted the anomalous mantle material occurring beneath the rift may extend laterally into this LVZ under the rift shoulder.
- if the model without the LVZ is accepted, the anomalous mantle material can be laterally constrained to exist beneath the surface expression of the rift.

The second possibility is preferred because there is direct evidence for the presence of the high velocity below the reflector, whereas there is no positive evidence for the LVZ above it. Also there is scant evidence for the LVZ beneath Lake Baringo from the axial line data.

(B) Teleseismic results.

The results from the KRISP 90 teleseismic survey show (Achauer et al., 1994), in an east-west section through Lake Baringo, a narrow wedge of low-velocity material underneath the rift axis broadening at about 100 km depth. The width of this zone at Moho depth is about 20 km, extending to 30-40 km at 68 km depth, and it has velocities of between 7.5 to 7.8 km/s in the uppermost mantle. Such a pattern would fit the width of the anomalous mantle found here. It does not show: 1) a thin layer of LVZ beneath the margin of the rift, and 2) high velocity values under the rift flank. However, firstly the depth resolution of the teleseismic results is not capable of resolving the thickness of the LVZ layer modelled here, and secondly, the proposed anisotropy giving the high velocity values is in the horizontal plane, so it could not be seen by the rays inverted in the teleseismic analysis, expected to travel along near vertical paths.

The velocity anomaly from the teleseismic results is in abrupt contact with its surrounds. It has been suggested (Keller et al., 1994b) that this is indicative of the young age of the intrusion or diffusion would have smeared it out laterally.

Young rifts are often characterized by high heat flow along their axis, suggesting that a significant amount of heat is transferred from the upper mantle into the crust raising the geotherms (e.g., Morgan et al., 1986). However, it has been shown that in the Kenya
Rift the high heat flow values on the rift floor can be attributed to local magmatic and hydrothermal activity therefore Welldon et al. (1994) suggest that rifting and uplift may precede the conduction of a significant thermal anomaly to the surface.

Geological evidence indicates that the first volcanic activity associated with the rift in the vicinity of the E-W seismic line occurred about 16 Ma ago (Baker et al., 1971).

(C) High velocity value / Anisotropy.

Examples of high velocities in upper mantle material have been identified in regions surrounding other rifting areas.

In the Jordan-Dead Sea Rift (Ginzburg et al., 1979) a discontinuity at a depth of 55 km within the upper mantle marks an increase in velocity to 8.6 km/s, a result obtained from seismic data recorded along the rift axis.

In the Afar region anisotropy in the mantle under the highlands surrounding the depression has been suggested to explain velocities higher than those beneath the depression itself (Gehlen et al., 1975).

In the Baikal region (Burmakov et al., 1987) high velocities in the mantle have been explained by anisotropy in garnet pyrolite, from the interpretation of a seismic line across the Baikal rift axis.

Anisotropy in the mantle has also been suggested to exist in other geodynamic regions, where alternating high and low velocity layers have been found. In Fennoscandia (Caswell and Fuchs, 1979) it has been suggested that shear stresses may create preferred orientation of olivine crystals which in turn would cause orientation of the maximum velocity direction.

Inversion of teleseismic S waveforms from the Grafenberg array (Southern Germany) detected the presence of two anisotropic layers in the upper mantle, explained by crystal orientation related to shear flow. The anisotropy in the layer with a relatively high velocity is considered to be 'frozen', i.e. due to fossil shear flow; the boundary between the two layers, at 54 km depth, is interpreted as the spinel to garnet phase transformation in the mantle material (Farra et al., 1991).

From ScS arrivals the upper mantle layering has been modelled between the Moho and 410 km depth in the western Pacific, Japan, the Philippine Sea and Australasia. An H (Hales) reflector at a mean depth of 60 km shows a positive impedance contrast and is
explained as the seismic expression of the spinel/garnet phase boundary (Revenaugh and Jordan, 1991).

Anisotropy and a spinel/garnet phase change are possible explanations for the velocity observed under the western flank of the Kenya Rift.

(D) Rifting or orogenic signature?

There is another intriguing possibility. The seismic line crosses the contact between the Archaean Nyanza Craton (ANC) and the Proterozoic Mozambique Orogenic Belt (MOB) approximately 70 km to the west of the rift’s western bounding fault, in the vicinity of the Nandi fault. It is considered that the contact results from a Tibetan style collision orogen (Shackleton and Ries, 1984; Berhe, 1990).

It is possible the upper mantle discontinuity identified here is a pre-rift structure, resulting from that orogen.

One of the clearest images of a sub-horizontal reflector within the mantle at depths equivalent to that discussed here is seen beneath the DRUM deep reflection line across the Caledonides offshore NW Scotland. This line identified the ‘W’ reflector at a depth of 45-50 km (McGeary and Warner, 1985). The reflector was interpreted as a major fault or shear zone within the mantle resulting from the Caledonian orogeny possibly reactivated during Mesozoic extension (Warner and McGeary, 1987).

It is not inconceivable that the reflector identified here beneath the western flank of the Kenya Rift has an equivalent origin and is a result of the Proterozoic Mozambique orogeny.

6.7 Summary

In conclusion, an interface within the upper mantle beneath the western flank of the rift has been identified. Detailed analyses of the kinematic and dynamic characteristics of the seismic wavefield confirm the presence of a velocity of at least 8.4 km/s underneath the western flank of the rift. The petrological modelling suggests:

- relatively cool temperatures adjacent to the rift, beneath which partial melt material has been predicted;
- the possibility of anisotropy in which the olivine crystal orientation is suggested as an explanation for the observed velocity structure, as the change in composition from spinel to garnet peridotite is not enough to account for the positive impedance jump across the reflector;

- a further possibility is that the reflector is a pre-rift structure, resulting from the ANC-Kibaran craton collision.

The temperature distribution modelled from the velocity values could be used to give an indication of the age of anomalously hot mantle material beneath the rift axis. Turcotte and Schubert (1982) model the temperature effect of an intrusion on the surrounding material at different times after the initial intrusion to provide a way to constrain its age. However, even assuming a large initial temperature difference of 500°C between the thermal anomaly and the surrounding lithosphere at about 35-45 km depth, then even 20 Ma after the first appearance of the thermal anomaly, the rise in temperature 20-30 km away from the anomaly would be only 200 – 300°C. Such a temperature rise would cause a P-wave velocity decrease of only about 0.1-0.15 km/s. A better constrained model incorporating results for example from possible future deep seismic reflection profiling could allow this type of modelling to be applied to the upper mantle beneath the Kenya Rift.

All the evidence taken together suggests that the upper mantle region of anomalously low velocity material is confined below the surface expression of the rift itself, to a depth below 60 km. There is the possibility of a thin lateral intrusion immediately above the high velocity layer which has been shown to exist beneath the western flank at about this depth.
Chapter 7

Inverse Modelling

7.1 Introduction

Inverse modelling aims to minimize the discrepancy between the observed travel times and those calculated through the seismic velocity model.

The advantage of inversion is that it provides an estimate of the model parameters' resolution and uncertainty, whereas this is problematic with forward modelling. Moreover, the interpretation time is greatly reduced compared to the time-consuming forward step; the data from multiple phases and source locations are used simultaneously. The result is relatively more objective than from forward modelling, but it still depends on the subjective phase picking and model parameterization.

Least squares inversion

Travel time inversion is a non-linear problem. The dependent travel time variable $t$ is a non-linear function of the independent velocity variable $v$.

The travel time along a path $L$ for a continuous velocity field can be written (Zelt and Smith, 1992):

$$ t = \int_{L} \frac{1}{v(x, z)} dl $$

(7.1)

and in a discrete form, used in practice:
\[ t = \sum_{i=1}^{n} \frac{l_i}{v_i} \quad (7.2) \]

where:

- \( l_i \) = length of the \( i \)th ray segment
- \( v_i \) = velocity of the \( i \)th ray segment

One can see that \( t \) is linearly dependent on the inverse of velocity, but the ray path is non-linearly velocity dependent, i.e. the problem is non-linear.

The linearisation of the velocity function requires a Taylor series expansion. For a generic function \( t = f(V) \), this is:

\[ t = f(V_0) + f'(V_0)dv + \frac{1}{2} f''(V_0)dv^2 + \text{higher order terms} \quad (7.3) \]

where \( V_0 \) is the starting velocity and \( f'(V_0) = \partial V_0 / \partial t \)

Second and higher terms can be neglected because the velocity is optimised by an iterative approach, resulting in:

\[ t \approx f(V_0) + f'(V_0)dv \quad (7.4) \]

Substituting \( t_o = f(V_o) \), where \( t_o \) is the theoretical travel time for \( V_o \),

\[ t - t_o \approx f'(V_o)dv \quad (7.5) \]

Defining the travel time residual \( r = t - t_o \), i.e. the difference between the observed and calculated travel time gives the least squares equation:

\[ r \approx f'(V_o)dv \quad (7.6) \]

The optimisation of the velocity model is achieved by minimising iteratively \( r \) with respect to the starting model. The iterative procedure is illustrated in Fig. 7.1. The partial derivatives are calculated while forward modelling ray-tracing in the starting model; the parameter correction value \( dv \) is solved using the least squares equation and used to produce an updated model. This is applied until an adequate fit between calculated and

\[ \]
observed travel times is achieved or a prescribed stopping criterion is fulfilled (Spence et al., 1985).

The partial derivatives of travel time with respect to velocity are analytically calculated using Fermat's principle by velocity and boundary node perturbation simultaneously. In the case of velocity perturbation, from (7.1):

\[
\frac{\partial t}{\partial v_j} = \int_L -\frac{1}{v^2} \frac{\partial v}{\partial v_j} dl
\]  

(7.7)

where:

\( v \) = velocity of the unperturbed model

\( v_j \) = \( j \)th velocity value selected for the inversion

\( \frac{\partial v}{\partial v_j} \) = velocity perturbation

_Damped least squares inversion_

In vector notation (7.6) can be expressed:

\[
A \Delta m = \Delta t
\]

(7.8)

where:

\( A \) = partial derivative matrix; it contains \( \partial t_i/\partial m_j \) elements, where \( t_i \) is the \( i \)th observed travel time, and \( m_j \) is the velocity value or coordinates of a boundary node selected for the inversion,

\( m \) = model parameter correction vector,

\( t \) = travel time residual vector.

To account for the error on the travel time data and the limited knowledge of the velocity function an error term is introduced in (7.8). The inverse problem being over-determined, a damping parameter \( D \) is also inserted to increase the stability of the inversion (Lutter et al., 1990). \( D \) must be chosen to minimise the trade-off curve between spread of model resolution and size of model covariance (Menke, 1984) (Fig. 7.2).

The damped least-squares (DLS) equation can be derived from (7.8) as (Lutter et al., 1990; Zelt and Smith, 1992):

\[
\Delta m = (A^T C_i^{-1} + D C_m^{-1})^{-1} A^T C_i^{-1} \Delta t
\]

(7.9)
a) Starting Model. $V_0$

Calculated travel time, $t_0$

Apply the least squares equation, $t - t_0 = f(V_0)dv$

b) Updated Model. $V_1 = V_0 + dv$

Calculated travel time, $t_1$

Reapply the least squares equation, $t - t_1 = f(V_1)dv$

c) Minimize the travel time residual, $r$

$r_0 = t - t_0$

$r_1 = t - t_1$

Figure 7.1: A simplified example of a linear travel time inversion scheme, using a ray-trace forward step and a least squares minimization technique.
Trade-off between resolution and damping

Figure 7.2: Trade-off curve resolution and variance for a given discretization of a continuous function (modified after Menke, 1984). The damping factor should be chosen to optimize the trade-off between parameter resolution and model stability.

where:

\[ C_t = \text{diag} \sigma_i^2 = \text{estimated data covariance matrix} \]
\[ C_m = \text{diag} \sigma_j^2 = \text{estimated model covariance matrix} \]
\[ D = \text{overall damping factor, generally equal to one} \]
\[ \sigma_i = \text{standard deviation, is the estimated uncertainty of the } i\text{th travel time} \]
\[ \sigma_j = \text{is an a priori estimate of uncertainty of the } j\text{th model parameter}. \]

From Fig. 7.2 it can also be seen that if the model parameterization is continuous the curve is monotonic and asymptotic to the two axes; an adequately fine discretization of the model maintains the trade-off curve characteristics (Menke, 1984). The trade-off between the size of the velocity and boundary corrections is determined by the relative sizes of the data and model covariance. The D parameter controls also the size of parameter corrections (Lutter et al., 1990; Zelt and Smith, 1992).

The travel time residual vector \( \Delta t \) and the partial derivative matrix are analytically calculated during ray-tracing.

The matrix to be inverted in (7.9) is limited in size since the model parameters are
usually less than the observations. The inversion is done with a lower-upper decomposition to produce a standard parameter resolution matrix ($prm$). The parameter resolution is related to the number of data points, of rays, and of nodes used in the inversion.

$prm$ diagonal elements are between 0 and 1, indicating the degree of linear dependence of the model parameters. They physically represent the relative number of rays that cover each sample parameter. Values greater than 0.5 generally indicate well resolved and reliable model parameters (Zelt and Smith, 1992).

Standard errors and uncertainty are calculated by the inversion scheme as a function of the uncertainty on the travel time picks only. Other possible sources of errors are: phase misidentification; modelling 3-D structure as 2-D; assuming a straight line for the receiver positions; and using an inadequate model parameterization to represent the complex real earth. Therefore, the calculated errors can be used only in a relative sense. Perturbing a parameter at a time, gradually, until the fit is lost, gives an absolute estimate of the uncertainty on the parameter itself.

Model uniqueness is estimated via the use of different starting models that might achieve a similar fit. The range of models that fit the data equally well can provide a measure of model non-uniqueness. Moreover, to study model uniqueness, it is possible to vary the damping parameter, to use different estimates of model uncertainties, or to hold some parameters fixed for a few iterations (Zelt and Smith, 1992).

Inversion of travel times from crustal refraction/wide angle reflection data in 2-D media generally employs ray-tracing algorithms designed for forward modelling (Huang et al., 1986; Firbas, 1987; Lutter et al., 1990). The efficiency of the inversion is therefore limited: a minimum number of model parameters and a computationally robust ray-tracing algorithm are specifically required by the inversion procedure.

To respond to these needs Zelt and Smith (1992) developed a technique where model parameterization and ray-tracing are designed for the forward step of the inversion.

The simultaneous travel time inversion for velocity and interface position of Zelt and Smith (1992) has been used. The program is called RAYINVR1.3 and it has been installed on a Sun/Sparc workstation of the Geology Department at Leicester University; x-windows graphic libraries are required by the inversion package; outputs can be in
CHAPTER 7. INVERSE MODELLING

The problem of travel time fit is non-linear, therefore the inversion scheme is iterative (Fig. 7.3). The forward step of the inverse approach includes ray-tracing through the starting, parameterized, seismic model.

Model parameterization is undertaken using a layered, variable block-size representation of a 2-D velocity structure. The medium is supposed isotropic. The model definition is similar to the one used for the forward modelling ray-tracing with MacRay (Section 3.4.1). Layers are defined by interfaces crossing the model from left to right, composed of linked linear segments. The velocity nodes are placed at arbitrary distances along the top and the bottom interfaces of each layer. Layers can 'pinch out' but not cross-over. The model can be checked using VMODEL (Fig. 7.3); format errors, large vertical and lateral velocity gradients or negative gradients, very small or large velocity values, low velocity zones, crossing interfaces and high slope interfaces are detected, flagged, and may be, in part, corrected. VMODEL can smooth the layer boundaries to increase the stability of the inversion results in the case of models with large lateral variations; it removes shadow
Chapter 7. Inverse Modelling

287

zones and extends travel time branches.

During ray-tracing the model is automatically divided into trapezoids in which the velocity is linearly interpolated.

The first step of the inversion is the generation of the partial derivatives by ray-tracing; the ray-tracing equations (equations 3.1, 3.2 in Section 3.4.1) are solved with the Range-Kutta method by the RAYINVR program (Fig. 7.3) (Červený et al., 1977; Zelt and Ellis, 1988). Two-point ray-tracing is avoided by linear interpolation of travel times and partial derivatives across ray endpoints to the locations chosen for the calculated travel times.

To test the travel time fit $\chi^2$, the normalised form of the misfit parameter (Bevington, 1969), is used. $\chi^2 = 1$ signifies a good fit without overmodelling. $\chi^2 > 1$ means it is not possible to find a model which resolves short-wavelength variation sampled by the travel time data. $\chi^2 < 1$ results from overmodelling the data, i.e. the fit is better than the travel times uncertainty (Zelt and Forsyth, 1994). See section 7.2.4 for further comments on $\chi^2$.

During ray-tracing the travel time residuals are also calculated and, together with the partial derivatives are output from RAYINVR and input into DMPLSTSQRF (Fig. 7.3).

DMPLSTSQRF updates the model parameters applying the DLS method to the linearized inverse problem. The modified model is then studied, edited, plotted or smoothed if necessary with VMODEL, and input again into RAYINVR and DMPLSTSQRF; the procedure is repeated until the stopping criteria are satisfied.

A good solution should give the best trade-off between rms (root-mean-square) travel time residual and parameter resolution, and trace rays to the maximum number of observed data with a minimum $\chi^2$. Stopping criteria and number of model parameters are therefore related, because the travel time residual can be reduced by increasing the number of model parameters; this generally reduces the model resolution.

7.2 Inverse Modelling of Line D

Inversion of travel times was applied to the data recorded along the cross-rift line D.

About 1200 observed data were used for the six shot points, including crustal and mantle phases.
The picks' uncertainty is dependent on the quality of the seismograms but it has been considered reasonable to estimate the following values over a group of traces, as done for the forward modelling (Section 3.2).

The picking uncertainty of \( a \) was estimated as 0.05 s to 100 km offset, 0.08 s between 100 and 150 km, and 0.10 s beyond 150 km for the shots outside the rift. For shots within the rift it increased at shorter offsets.

For \( b_1 \) the uncertainty was set to 0.15 s, with the exception of the near offset from VIC; the same value of 0.15 s was used for \( b_2 \), \( c \) and \( d_1 \).

For \( d \) the uncertainty was fixed at 0.20 s; between 250 and 350 km from VIC the uncertainty was decreased to 0.10 s, because detailed processing allowed more precise picking (Section 6.4.2). Beyond 350 km it was increased to 0.20 s and again to 0.30 s from 380 km to the end of the section.

Braile et al. (1994) have previously applied inversion to the cross-rift line; however, the approach used here is slightly different. They did not introduce interfaces in the shallow part of the model, while in the present inversion the same number of interfaces as in the forward model (Fig. 3.22) were used in the rift graben. Both approaches are justified because the velocity structure in the first few kilometres is very complex for the ray-tracing algorithm. For the same reason here the arrivals from the basement cover were not ray-traced and average velocities in the same cover were derived from the forward model. In other words, the basement cover was not inverted and the velocity structure inside it was kept as simple as possible to invert the deeper layers.

For brevity the forward model described in Fig. 3.22 will be called hereafter model F, and Braile et al.'s (1994) will be model G.

Braile et al. (1994) used a smooth velocity model uniformly gridded in the upper crust, but with closer grid spacing in the areas where the ray coverage was greater. Wang (1993) showed that with an uniform grid spacing DLS tends to change the velocity in areas of good ray coverage and leaves areas with poor ray coverage unchanged. Only a starting model very close to the real earth would avoid solution divergence and problems during ray-tracing.

In the inversion presented here the velocity nodes were placed in the same position as
in model F.

For the deeper part of the model a layered velocity structure similar to model F was used.

The topography was defined as the first interface and held fixed during all iterations.

Some of the array sizes within RAYINVR had to be altered to suit the present problems. The maximum number of points defining a single model layer had to be increased from 50 to 80; the maximum number of trapezoids in a layer from 100 to 150. 10 model layers were used.

Step lengths of 0.025 km were used during the inversion to minimise the computation time and also to detail local structures.

Ray take-off angles were automatically calculated for all shots into the line once the required phases were specified. A maximum of 20 rays per phase have been traced.

A ‘layer stripping’ technique was applied. Starting from the shallow part of the model, deeper layers were progressively inverted and then fixed during the inversion of still deeper layers.

During each iteration the model was run through VMODEL to be smoothed in order to increase the stability of the results.

After considerable testing, satisfactory stopping criteria resulted in $\chi^2 \leq 3$, an rms travel time $\leq 0.2$ and successful rays using $\geq 80\%$ of the data points.

The damping parameter was normally set to 1.0. For some iterations it was necessary to increase it up to 20 to avoid the updated model having a poor travel time fit. As suggested by Zelt and Smith (1992) the final parameter resolution estimates were calculated with $D=1$ for consistency during the inversion procedure and as the lowest possible value which improved resolution of the model parameterization while damping undesired oscillations of the model parameters (Lutter et al., 1990).

During the DMPLSTSQR inversion the velocity and boundary accuracy were set respectively to 0.1 km/s and 1.0 km.

prm values were checked after each iteration and either the areas where poor resolution occurred were constrained to the value from the forward modelling, or the whole model was re-inverted with a higher $D$, or the number of model parameters was decreased. prm values
CHAPTER 7. INVERSE MODELLING

$> 0.50$ were considered satisfactory because they indicate good parameter independence between a node and the adjacent ones (Bralle et al., 1994).

The inversion was completed in three sequential phases:

1 - crustal phases $a$ and $b$ defined the basement velocity field,

2 - mid- and lower crustal reflected phases $b_1$ and $b_2$, and mid-crustal diving waves $b$ detailed the middle and lower crustal interface topography and the mid-crustal velocity variations, and

3 - the Moho reflection and mantle diving waves phases $c$ and $d$ resulted in definition of Moho topography and the uppermost mantle velocity structure.

Two sets of models were produced for every step/layer:

- underparameterized, to ensure stability of the inversion algorithm and to maximise the nodal resolution, and

- overparameterized, derived from the ray-traced model $F$, to estimate the degree of resolution/error of that model. Slight modification to model $F$ had to be made to adapt the format for VMODEL: velocity inversions were removed (in the basement near VIC, and under the high velocity block west of BAS), large lateral gradients were decreased; in the upper mantle the model without the LVZ at c.60 km depth was chosen (see Chapter 6).

Whenever and wherever the complexity of the model or the lack of data resulted in unstable and divergent models, constraints were incorporated from forward modelling, geology and gravity studies.

The range of models derived from different parameterizations converged towards a solution presented as the final model.

Absolute uncertainty of the inverted parameters was estimated in a manner suggested by Zelt and Smith (1992) on the final model. A single parameter value was perturbed from that derived in the final model and held fixed while a further inversion of the final model was carried out; the remainder unperturbed parameters were allowed to vary during the inversion. The perturbation was increased, in both positive and negative directions, until the perturbed model was unable to fit the observed data as well as the unperturbed final model. The goodness of fit was considered in terms of: (1) the number of rays traced
through the model; (2) the result of an $F$ test on the $\chi^2$ values of the perturbed and unperturbed model; and (3) the rms values for the perturbed model differing significantly from the estimated pick uncertainty on the phase considered in the inversion. The level of significance was set to 5%.

Error bounds on upper crustal velocities where the resolution is good (Section 7.2.1) are better than ±0.05 km/s.

The uncertainty on the mid crustal interface is ±1.0 km; the mid crustal velocity has a large value of ±0.6 km/s as it has been inferred from later reflected phases; a smaller value is estimated for the lower crustal velocity, equal to ±0.3 km/s. The geometry of the Moho depth and the uppermost mantle velocity are uncertain to ±2 km and to ±0.4 km/s respectively.

Selected iteration results are included in Appendix C.

### 7.2.1 Upper Crust

The upper crust is represented by a model to 15 km depth (Fig. 7.4), where the topography and the basement cover structure and velocity were fixed during the inversion. Unrealistic results and divergent solutions were obtained if the arrivals from the rift infill were included in the modelling, as discussed previously.

It was impossible also to recover the steep structure of the rift bounding faults, introducing an angle of dip too high for VMODE. The Elgeyo escarpment and Kerio Valley trough would have been completely obliterated in the attempt to produce a smooth and stable inversion. Therefore information from the seismic travel times forward modelling was included, and the model for suprabasement structure was not allowed to alter.

The sedimentary layer is represented with an average velocity of 2.69 to 2.77 km/s; the upper volcanic infill 3.91 to 4.34 km/s; and the lower volcanic infill 5.36 km/s to avoid having to include the large lateral variations resulting in model F and which were unacceptable to VMODE.

$a$ and $b$ arrivals were selected from each shot point to a maximum of 180 km offset, to limit the ray depth penetration within the top 15 km. A total of 573 data points were inverted. Beyond 180 km, arrivals penetrated the deeper crust and were inverted in the
Figure 7.4: Velocity model for the upper crust resulted from the travel-time inversion of line D. Only selected velocity values are included in the diagram. (Velocities in km/s).

next step.

The starting model included a gridded velocity parameterization denser in the vicinity of the rift. The iterations were performed with D between 1 and 5. To estimate the model uniqueness different starting models were inverted, converging towards similar solutions in terms of resolution, goodness of fit and lateral velocity variations.

The model presented in Fig. 7.4 inverts 85% of the arrivals, with $\chi^2$=2.8, and the travel time residual (rms error)=0.10; the $prm$ for the upper velocity values is $\geq$0.60, and less for the lower velocity. 70 model parameters (all velocity nodes) were inverted making it an overparameterized type model.

Underparameterized models showed much higher $prm$, but also gave a higher $\chi^2$ and rms travel time residual for less data points used. The final upper crustal model is taken to be the overparameterized model, taking into account that the resolution is poor where there is no ray coverage, i.e. at the sides of the model.

Fig. 7.5 represents the raypath coverage for the upper crust; the first 30 km from VIC do not fit, probably because the high velocity layer has not been included in the model. From KAP to the east, beyond 240 km along the line and from BAR and TAN to the west...
beyond KAP, the Kerio Valley structure prevents rays from being traced to the surface. Removing the Kerio trough increased the number of data points used; however, the trough was included in the final model.

Observed and calculated travel times are also plotted. The calculated branches result only from rays successfully traced, whereas all the observed data are presented. Vertical bars in the observed data define the picking errors.

The model in Fig. 7.4 was chosen to reflect the geologic complexities inherent in the data, trying to avoid poorly constrained lateral velocity variations.

The lateral velocity variations reproduce model F’s lower velocity variations in the Archaean Nyanza Craton (ANC), increasing towards KAP and under the rift, and decreasing towards CHF (Fig. 3.23). The velocity under KAP is lower than in model F. This might be due to the velocity in the volcanics being higher than in model F, being averaged along the layer in the inverse model. The high velocity block west of BAS cannot be resolved, and one can observe the poor fit between TAN and BAS (from 270 to 300 km), especially for the arrivals from BAS to the west.

The velocity gradients were taken from the results of a reflectivity modelling; gradients from the inversion are higher than from forward modelling especially under KAP but, as velocity resolution from the inversion at depth is poor, it has been decided to use the gradient values modelled by reflectivity (Section 5.3.1).

7.2.2 Lower Crust

The lower crust has been modelled to a depth of 35 km (Fig. 7.6), keeping the upper crustal velocity structure fixed from the previous step while the mid-crustal interface was inverted. The starting model included an homogenous mid-crustal layer with velocity 6.54 to 6.61 km/s, the average from the forward modelling; the grid was uniformly spaced; the lower crust at the first iteration has a velocity from 6.76 to 6.97 km/s.

A total of 223 data points were inverted, including the b arrivals beyond 180 km offset, together with \( b_1 \) and \( b_2 \).

The resolution of the lower crustal boundary, defined only by \( b_2 \) from VIC, resulted in \( prm \) values lower than 0.5, even with an underparameterized model. It was decided
Figure 7.5: Traveltime inversion modelling for shallow velocity model for the a and b phases. a: observed (vertical bars are picking errors) and calculated (lines) travel-times. b: raypath coverage for the phases used in the travel-time inversion.
Figure 7.6: Velocity model for the upper and lower crust resulting from the travel-time inversion of line D. Only selected velocity values are included in the diagram. (Velocities in km/s).

to use its topography from model F (Fig. 7.6). This is another difference from model G, because Braile et al. (1994) have considered neither the $b_2$ phase nor did they introduce such a lower crustal interface.

The model in Fig. 7.6 inverted 86% of the arrivals, with $\chi^2=1.46$, and the travel time residual (rms error)=0.167; the $prm$ values are $\geq 0.80$. In particular, the spatial resolution is better outside the rift, between 0 and 100 km and 250 to 450 km; the velocity resolution is good only on the west side of the model, where $b_2$ diving waves are traced and constrain the model.

The ray diagram in Fig. 7.7 shows how the coverage, and therefore the resolution, is better outside the rift. The mid-crustal boundary clearly deepens to the east between VIC and KAP, under the western shoulder, and rises again towards CHF, under the eastern shoulder. The topographic high under BAR identified in model F (Section 3.4.2) is mainly constrained by data from the axial line interpretation, and is poorly resolved with the data used for this inversion. Braile et al. (1994) used a mid-crustal reflection from BAR to the east to constrain it; the phase was neither picked nor modelled in the present study, possibly due to the slightly different processing procedure used on the BAR record.
The smooth layer boundary used for the inversion avoids the step-like structure seen in Fig 3.22 from model F. Thirty model parameters were inverted making it an overparameterized type model.

Velocity variation in the mid-crust is similar to model F, with 6.5 to 6.6 km/s, and the average was higher towards the east. Nevertheless, in this case, rather than using a 'layer-stripping' approach, because of the limited resolution of the velocity field, the lower crust has not been fixed during the next step of the inversion.

7.2.3 The Moho and the Upper Mantle

The inversion of the Moho boundary has been undertaken incorporating also the lower crustal and upper mantle velocities, using \( c \), \( d \) and \( d_1 \) phases, and a total of 397 observed data points.

To produce the model in Fig. 7.8, 86% of the data points were used, the travel time residual (rms error)=0.22, and \( \chi^2 \) is equal to 1.96; the \( \text{prm} \) values are \( \geq 0.60 \) in the centre of the model.

Moho interface resolution was good from c.50 to 340 km, including 10 nodes. The uppermost mantle reflector, as expected, is poorly resolved, involving only one unreversed reflection; \( d_1 \) has been excluded from the final model. Overall, 40 model parameters were inverted making it an overparameterized type model.

The velocity structure in the uppermost mantle is well resolved at distances between 25 and 400 km from VIC.

In Fig. 7.9 the ray diagram is shown for the \( c \) and \( d \) phases.

During the ray-tracing it was noted that the arrivals to the west of BAR have to be modelled as head waves from the Moho. If only diving waves in the uppermost mantle are traced, no arrivals are seen for phase \( d \).

At about 260 km from VIC the same misfit of c.0.4 s on \( d \) as seen in the Braile et al. (1994) inversion is visible. This phase has been picked with a good level of confidence.
Figure 7.7: Traveltime inversion modelling for mid-crustal velocity model for the $b_1$ and $b_2$ phases. Shallow velocities were fixed using the model in Fig. 7.4. a: observed (vertical bars are picking errors) and calculated (lines) travel-times. b: raypath coverage for the phases used in the travel-time inversion.
CHAPTER 7. INVERSE MODELLING

Figure 7.8: Final velocity model resulted from the travel-time inversion of KRISP 90 cross-rift line D. Only average velocities are reported for the crust, and selected velocity values in the upper mantle (Velocities in km/s).

(0.01 s uncertainty, following the detailed analysis of Chapter 6). As Braile et al. (1994) commented, the advance might be due to the presence of a high velocity block, c.20 km in width, probably located in the rift basin. The misfit could also be due to a 3-D effect in a portion where the line has a sharp bend between BAR and TAN. The shallow seismic model does not include such a feature, because of the ray path coverage ‘in line’. Alternatively, the misfit could be due, as suggested in Section 3.4.1 to the lack of rays diving from VIC through the upper mantle topographic high beneath the rift.

Braile et al. (1994) interpreted the x phase as being an extension of the c phase from BAR. This produced a sharp Moho depth change on both sides of the uppermost mantle upwarp beneath BAR. As discussed in Section 5.3.2, in this analysis x has been interpreted and modelled as a c phase multiply reflected in the rift infill.

To test the model non-uniqueness a different starting model was used, with homogeneous velocities in the lower crust and upper mantle. No lower crustal interface was inserted. The final model resulted in a velocity distribution differing by up to ±0.2 km/s
Figure 7.9: Traveltime inversion modelling for the deep crustal velocity model for the c and d phases. Shallow velocities were fixed using the model in Fig. 7.6. a: observed (vertical bars are picking errors) and calculated (lines) travel-times. b: raypath coverage for the phases used in the travel-time inversion.
in the lower crust and ±0.15 km/s in the uppermost mantle velocities, while depth to the Moho changes by up to ±3 km.

In the unsmoothed model the Moho topographic high is wider than in the F model, extending from c.150 km to c.310 km along the profile.

7.2.4 Summary

Upper crust

As explained by Zelt and Smith (1992), values of $\chi^2$ greater than one might be due to high quality data (small pick uncertainty) sampling small-scale heterogeneities which are not resolvable. The choice of an underparameterised model leads to a low rms and a well resolved model with good ray-coverage, although $\chi^2$ was still greater than one.

In the upper crust, large out-of-plane velocity and structural variations, deviations of the shot-receiver geometry from a straight line and non-ray arrivals identified as $a$ could also increase the $\chi^2$ value.

The differences in velocity field between the inverse and F model have been already commented on. Comparison with model G shows there are no significant differences in the velocity field. The smooth velocity variation with higher values towards the axis of the rift than under the shoulders appears similar in the two models.

Lower crust

No significant differences exist in the mid-crustal boundary between the inverse and F model, considering the well resolved parts of the inverse model and the uncertainty on the boundary parameter values.

An unsmoothed version of the boundary (not presented here) shows similar structure to that seen in model F (Fig. 3.22) and in model G. However the structural high under the rift axis is not seen in the present model.

Model G's structural high in the mid-crust under the rift is due to the modelling of $b_1$ from BAR, which has not been included in this inversion.

Velocities in the mid-crust are lower than in model F, but not beyond the large uncertainties on these values.
The Mooho and the Upper Mantle

In the uppermost mantle the velocity field shows values higher than in model F under the western flank of the rift even taking into account the uncertainty in the respective values. This results from a ray field which excludes the c phase from BAS to the west. A model where this phase is also fitted gives velocities under the western flank compatible with model F.

A lower velocity is found beneath the eastern flank. This feature seems to be the most significant result from the inversion study because it is repeatable for different starting models, even when the starting velocity field is homogeneous throughout the upper mantle. The lowest velocity value is shifted eastward, at c.300 km along the line, compared to model F (Fig. 3.22).

The Moho topography shows differences from model F which are within the error bounds of the boundary parameter values. A topographic high in the Moho is modelled beyond 300 km along the line, further east than in model F: mantle phases from BAR and TAN to the east fit better with the high at c.300 km, while the precritical phase from BAS to the east requires the culmination to be at 350 km. The first solution is the preferred one, because the BAR and TAN phases are visible over a longer range than the BAS phase.

Model G for the Moho structure is compatible with the one presented here from 0 to 100 km and under the rift axis. The depth under the two shoulders is constrained by the c reflection from BAR, the same phase interpreted here as a multiple of c in the basement cover. The depths towards the east end of the profile are poorly resolved in both the model G and the present model. The difference arises primarily from the fact that in Braile et al.’s (1994) inversion no c phase has been considered from TAN.

The low velocities under the eastern flank could be partly explained by the lower crustal velocity being higher in the inverted model than in model F. Moreover the inclusion of d from KAP and from VIC beyond 350 km lowers the velocity in that portion of the model. However, the data are unreversed and the data quality does not justify a modification of the final model resulting from the inversion results here.
7.3 Gravity Modelling

The model resulting from the travel time inversion was examined via gravity modelling using the same procedure as followed for the ray-traced model F.

For clarity the improved model described in Fig. 3.25 will be called hereafter model A, and the one derived from the inversion, model B.

The observed data set was extracted from the catalogue by Swain and Khan (1978), and corrected for a regional anomaly as described in Section 3.5. Models A and B were thus directly comparable, the same program GRAVMAG (Busby, 1987) being used for the gravity modelling. The PC version of the BGS GRAVMAG (Pedley, 1991) is a 2.5-D forward modelling/parameter optimisation gravity and magnetic package.

The gravity theory is based on that described by Rasmussen and Pederson (1979). The program obtains a 2.5-D model using polygons with specified dimensions in an X, Z plane and assigning a finite strike length in the Y direction. The polygons are symmetrically placed about the profile perpendicular to the polygons' strike.

The polygons' dimensions and density can be altered interactively to enable rapid modifications of the model. Constrained optimisation of either polygon coordinates or density accelerates the convergence to the best fitting model and can be used to study the significance of node positions and body densities.

The structural boundaries were taken from the final inverted model B.

An initial model B1 where the densities were simply derived from the inverted seismic velocities using the Nafe-Drake relation (Ludwig et al., 1970) gave a poor fit with the data and was therefore discarded.

Modifications of B1 were applied as follows to give the second model, B2:

- the sedimentary and volcanic layers in the inverted seismic model had been derived from the ray-traced forward model F, which has been used to generate model A. Therefore the same modifications introduced in the shallow part of the ray-traced model to generate model A were inserted here. In particular the densities of the sedimentary units, the lateral subdivision of the volcanic units in the west and central part of the Rift Valley, and the different densities assigned to the basement blocks were reused here in model B2.
- from the mid crust to the Moho the interfaces have been taken from the inversion model B. A V shaped low-velocity uppermost mantle body was limited at the top by the topographic high in the Moho and in the lower part it narrowed, similar to that used in model A.

Half strike lengths in the Y direction were taken from model A described in Section 3.5; there they had been set according to the order of magnitude of the structural length for the various geological units: to 10 km for the sedimentary layers in the Kerio Valley, Lake Baringo and Tangulbei; to 100 km for the volcanic units and the upper crustal units; and to 10000 km for the mid and lower crustal layers and for the uppermost mantle to 60 km depth.

The model was extended laterally to 10000 km length to avoid edge effects.

Interactive gravity modelling of B2 produced an improved model B3 presented in Fig. 7.10. The largest changes between B3 and B1 occur in the rift infill, where the inversion was performed using constant velocity values within these layers (Section 7.2.1). The derived density values were not allowed to differ from the B1 values by more than ±0.2 g/cm$^3$. This value was obtained from Barton (1986; Fig. 1 and Table 1) as the resulting uncertainty following velocity to density conversion, for the densities considered here.

The values used in the shallow part of B3 were derived from model A. Thus the comparison between A and B3 is mainly focused on the deeper part of B3 from the subbasement to the upper mantle.

The fit between the observed and calculated gravity anomaly achieved with the B3 model is slightly better than with model A. This is shown by the comparison of the sum of squares of the differences between the calculated and observed values for the two models.

The misfit visible between 330 and 360 km could be decreased if the high velocity block in the basement (HVB) under BAS is included. The HVB was not introduced in the B models.

Over the range between 60 and 350 km model A gives a better fit to the observed data. As pointed out above, a misfit in B3 could be decreased by introducing the HVB; however, model A without the HVB still achieves a better fit than model B3 (Fig. 7.11).

The largest misfit for B3 between the calculated and observed values occurs between 60
Figure 7.10: The observed and calculated gravity profiles along the cross-rift line D showing the density model B3 derived from the final inverted seismic model. (Densities in gcm$^{-3}$).
and 110 km. To improve B3 the depth of the mid-crustal and lower crustal boundaries was decreased towards the west: this resulted in a comparable overall fit for models A and B3 in the 60-350 km range of the line. Unsmoothed boundaries in the shallow, western part of the inverted model might therefore give a better fit to the gravity data; nevertheless, the final inverted model had to be smoothed to facilitate the ray-tracing through the deeper layers.

The density values used in models A and B3 are reported in Table 7.1, and a full listing of model B2, B3 and for comparison, A, parameters is given in Appendix D:

<table>
<thead>
<tr>
<th>Unit</th>
<th>Vp (km/s) fww</th>
<th>Vp (km/s) inv</th>
<th>( \rho ) model A gcm(^{-3})</th>
<th>( \rho ) model B3 gcm(^{-3})</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sediments</td>
<td>2.6</td>
<td>2.6</td>
<td>2.10</td>
<td>2.10</td>
</tr>
<tr>
<td>Sediments</td>
<td>3.0</td>
<td>2.6</td>
<td>2.30</td>
<td>2.30</td>
</tr>
<tr>
<td>Upper volcanics</td>
<td>3.9</td>
<td>3.85</td>
<td>2.46</td>
<td>2.46</td>
</tr>
<tr>
<td>Lower volcanics</td>
<td>5.2</td>
<td>5.4</td>
<td>2.55</td>
<td>2.55</td>
</tr>
<tr>
<td>Lower volcanics</td>
<td>5.8</td>
<td>5.5</td>
<td>2.50-2.75</td>
<td>2.50-2.75</td>
</tr>
<tr>
<td>Basement</td>
<td>6.1</td>
<td>6.1-6.2</td>
<td>2.70-2.75</td>
<td>2.70-2.75</td>
</tr>
<tr>
<td>Basement</td>
<td>6.3</td>
<td>6.3</td>
<td>2.80-2.82</td>
<td>2.83</td>
</tr>
<tr>
<td>Upper crust</td>
<td>6.6</td>
<td>6.6</td>
<td>2.85</td>
<td>2.88</td>
</tr>
<tr>
<td>Lower crust</td>
<td>6.86</td>
<td>6.9</td>
<td>2.92</td>
<td>2.90</td>
</tr>
<tr>
<td>Upper mantle LVZ</td>
<td>7.6</td>
<td>7.7</td>
<td>3.30</td>
<td>3.22</td>
</tr>
<tr>
<td>Upper mantle - W</td>
<td>8.1</td>
<td>8.1</td>
<td>3.25</td>
<td>3.27</td>
</tr>
<tr>
<td>Upper mantle - E</td>
<td>8.1</td>
<td>7.9</td>
<td>3.35</td>
<td>3.24</td>
</tr>
</tbody>
</table>

Table 7.1: Velocity versus density values used in modelling the gravity data in the forward model A and inverted model B3.

The most noticeable difference between A and B3 is the lateral variation of upper mantle density under the two flanks of the rift; that under the eastern flank being lower than under the western flank. This follows from the velocities beneath the two flanks derived from the inversion.

This might be explained in terms of the difference between the Archaean and Proterozoic lithosphere preserved under the two flanks of the rift to c.60 km depth. It is known that cratons are likely to have a 'cold, thick' lithospheric keel originally reaching the diamond stability depth of c.150 km (Durrheim and Mooney, 1994; Fig. 1.15).

Studies of the gravity anomaly over cratons may suggest the presence of the lithospheric
Figure 7.11: The observed and calculated gravity profiles along the cross-rift line D showing the density model derived from the model in Fig. 3.25 (model A) without the high velocity block beneath BAS. (Densities in g cm$^{-3}$).
keel:

- the effective elastic lithospheric thickness ($T_e$) of the East African Plateau is c.50
  km while other cratonic areas in Africa have values between 64 and 90 km (Ebinger et
  al., 1989), so magmatic processes may have reduced $T_e$ underneath the ANC. This is
  consistent, assuming a link between $T_e$ and thickness of the keel, with Menzies' (1990)
suggestion that plumes and hot asthenosphere can obliterate the original characteristics
of the Archaean keel.

- it should be possible to observe long wavelength positive isostatic anomalies over
  cratons if the keel is still preserved, as happens in the Western African Craton (Lesquer and
  Vasseur, 1992); such a gravity signature is not found over the ANC. The ANC is, as part
  of the East African Plateau, isostatically compensated with the compensation originating
deeper than 60 km (Ebinger et al., 1989; Maguire et al., 1994). The regional anomaly
  correction used here assumes the presence of such deep compensation from Maguire et al.
  (1994).

On a regional scale and for depths greater than 60 km it is not possible to distinguish
between ANC and MOB lithosphere.

However, at shallower depths in the uppermost mantle, a density difference can be
envisaged between Archaean and Proterozoic lithospheric mantle. It is suggested that the
higher density under the ANC might be representative of a 'relic' of the lithospheric keel.

The difference in upper mantle characteristics on the two sides of the rift might also
be consistent with the concept of asymmetric rising of the mantle plume under the rift,
away from the craton margin, as suggested by Smith (1994).

Two scenarios are therefore possible: (1) a pre-existing higher density under the ANC
than under the Mozambique Orogenic Belt (MOB) has been attenuated by a rising plume,
but is still detectable and (2) the lithospheric keel has been completely obliterated by the
plume which is now rising asymmetrically lowering the average density under the MOB
with respect to that under the ANC. A combination of the two processes is also conceivable.
7.4 Discussion and Summary

The application of travel time inversion using the ray-traced model as a starting model achieved a good match with the observed travel times and allowed an estimate of the model resolution and parameter uncertainty. In the upper crust it was not possible to invert observed data obtained from rays travelling across complex structures (e.g., the Kerio trough). Moreover, small scale structures as the HVB under BAS were not resolved by the inversion procedure; the $\chi^2$ value greater than 1 in the upper crust is an indication of that limited resolution. Considering the velocity uncertainty values in the upper crust, the lateral velocity variations presented in the forward model are well resolved.

The resolution is poor at the two extremities of the forward model and over the lower crustal boundary; the mid-crustal layer is overparameterized, and its associated velocities poorly constrained. The Moho boundary and the velocity field in the upper mantle are properly sampled by rays in the forward model. The uncertainty on the Moho depth estimated during forward modelling (Section 3.4.2) is confirmed following the inversion.

An advantage of the Zelt and Smith (1992) ray-tracing used here over that used in the MacRay program is evident modelling phase d from BAR to the west: head waves traced by RAYINVR fit the observed phase where MacRay diving waves were unable to shoot rays to the surface.

The upper mantle velocity uncertainties establish that a LVZ is present under the rift axis; however, the lateral variations might not be as large as presented in Fig. 3.22. The uppermost mantle reflector is not well resolved by ray-tracing.

The most significant differences between the Braile et al. (1994) and the present model result from different identification of the phases picked.

Overall, the final inverted model is less detailed than the forward model from mid-crustal to upper mantle depths, showing that not all the structures presented in the forward model are well resolved by the data. It can be said that no striking differences have been found between the forward and the inverted model. However, the thinning of the lower crust beneath the rift appears greater than in the forward model, but this might be due to the fact that the mid-crustal boundary under the rift was mainly constrained
by the axial line interpretation during forward modelling.

The upper mantle velocity is consistently and significantly lower than in the forward model under the eastern flank and results in a better long wavelength fit of the gravity data. A relationship could be invoked between density and age of the lithosphere, this being older under the western flank; alternatively, the effect of an asymmetrically rising plume could have lowered the density under the eastern flank more than under the western flank.

Another important point arising from the combined inversion/gravity modelling is the following: a step like structure in the mid-crust boundary under the rift's western flank is not sufficiently resolved by the seismic data, and, if included in the model, would result in overparameterization. It was finally decided to include in the inverted model a smoothed mid-crustal boundary, better resolved by the seismic data alone. The smoothed version of the boundary however, gives a fit to the gravity data which is poorer than the unsmoothed step-like version. This suggests that: (1) the thickness of the upper crustal layer in the west, larger in the inverted than in the forward model, might not be real, and (2) a sudden deepening in the mid-crustal boundary does occur at about 100 km from VIC along the profile.
Chapter 8

Discussion and Conclusions

8.1 Introduction

The main points arising from the previous chapters will be recounted here, followed by a discussion concerning their implications for the structural and geodynamic development of the Kenya Rift. Results from the other surveys undertaken during KRISP 90 will be briefly mentioned in relation to the present work.

The importance of the integration of different techniques used in the interpretation of seismic data will also be stressed.

A summary of the conclusions will complete the Chapter.

An attempt will be made to answer specifically the original questions posed by KRISP 90, in particular those put forward prior to the undertaking of the cross-rift line.

8.2 Summary of Findings

The specific questions to be answered by the cross-rift line were mainly related to structural variation between the rift axis and its flanks which was found to change significantly and over a short distance beneath the rift margins.

Rift Infill

One objective was to resolve the graben structure and the depth of the rift infill. An asymmetric basin, with thickest sedimentary accumulation against the rift’s western main
boundary fault, was defined with internal layering in the volcano-sedimentary sequence and including surface sedimentary sub-basins. Thicknesses and velocities are higher than expected from previous seismic surveys (Swain et al., 1981) with the exception of the Kerio Valley infill. The low velocity-low density zone identified under the Kerio Valley had been inferred by the gravity interpretation of Swain et al. (1981), but KRISP 75 seismic data did not constrain the depth of the basement under the Kerio Valley.

On the western shoulder, the synformal shape at the base of the phonolitic plateau is suggestive of an erosional episode preceding the deposition of the Miocene lavas.

**Upper Crust**

P- and S-wave velocities and amplitudes in the basement rocks seem to reflect the lateral transition from Archaean Nyanza Craton (ANC) to Mozambique Orogenic Belt (MOB) terrane, with a further zonation underneath the axis of the rift affected by mafic intrusions. P velocities vary from 6.2 km/s in the ANC to about 6.0 km/s in the MOB except for c.6.2 km/s under the rift axis. S velocities vary from 3.5 km/s in the ANC to c.3.6 km/s in the MOB; this implies an increase in $\sigma$ in the MOB. Study of seismic amplitudes suggests a higher velocity gradient in the ANC (0.003 s$^{-1}$) than in the MOB (0.001 s$^{-1}$). Underneath the rift axis the gradient is 0.002 s$^{-1}$.

**Lower Crust**

Questions concerning possible crustal underplating, and thinning of the crust related to rifting, required study of the lower crustal layer.

A boundary at about 10-15 km has been modelled as a first order discontinuity along the whole line. However, dynamic modelling of the P and S phases suggests that to account for the relatively low amplitudes of the reflections arising from that boundary various explanations should be considered: (1) the effect of attenuation in the crust (2) a laterally varying and/or complex discontinuity (see e.g., Braile and Chiang, 1986); and/or (3) the presence of a transition zone or small scale heterogeneities above the reflector, or random velocity fluctuations might result in a homogenization of the wave field by scattering (Levander and Höliger, 1992).

In the mid-crust there is a mean uniform P-wave velocity of c.6.5 km/s while in the lower crust the velocity is c.6.8 km/s. It is slightly lower under the rift axis possibly...
resulting from thermal processes beneath the rift raising crustal temperatures (Maguire et al., 1994). S velocities of c.3.8 km/s give a high Poisson's ratio (\(\sigma=0.26\)) in the mid-crust which may be due to the presence of fluids at low-pore pressure. In the lower crust S velocities average at c.3.9 km/s, providing a normal value for \(\sigma=0.25\).

Thickness of the lower crust, decreasing under the axis of the rift, will be discussed in relation to the other KRISP 90 lines in Section 8.3.

**Upper Mantle**

One of the principal objectives of the KRISP 90 refraction program was to obtain clear upper mantle seismic phases.

Forward modelling ray-tracing of the KRISP 90 data indicates the Moho topography rises under the axis of the rift over a distance which mirrors its surface expression. Moho depth varies from 30 ± 2 km beneath the rift axis to 38 ± 3 km beneath the western flank and 34 ± 2 km beneath the eastern flank. However, travel time inversion of the P-wave arrivals gave an alternative model where the mantle upwelling is broader and extends under the rift's eastern flank.

The P-wave velocity is everywhere about 8.0 km/s except beneath the rift where it is 7.6 – 7.8 km/s. This anomalously low velocity suggests a 5 – 6% partial melt (Section 3.4.2).

Amplitude modelling of the P and S reflections from the Moho showed, as for the mid-crustal boundary, that the Moho interface may be more complex than a single boundary with a discontinuous velocity increase.

The detection of a reflected phase from within the uppermost mantle allowed the velocity structure to be modelled to 60 km depth under the western flank. Preferred orientation of olivine crystals combined with a phase change in the mantle (spinel/garnet) may be invoked to explain a P-wave velocity value beneath the reflector which could be as high as 8.6±0.1 km/s at 63±2 km depth. Alternatively, the reflection could arise from a relict eclogite layer created during the ANC-Kibaran craton collision, if oceanic crust was present before convergence.
Gravity Data

The results of the joint interpretation of both the seismic and gravity data indicate the East African Plateau has a source of compensation deep in the mantle (> 60 km).

Other Results

Confirmation of previously known facts are the poor energy transmission in the Kenya Rift area (phase $a$ has an amplitude decay of $s^{-1.5}$ to $s^{-2.5}$). There is poor transmission of P- and S-wave energy from shots inside the rift, and for phases travelling through the rift.

Surprisingly, the frequency content of mantle phases is not lost through the uppermost mantle, as might be expected for waves travelling through partial melt. Spectral analysis showed that attenuation by scattering and absorption occurs primarily in the crust beneath the rift.

Clear high-amplitude reflected phases from Moho depths are seen in the S-wave window, precluding an extensive hot regime at lower crustal depth.

More generally, line D as part of the KRISP 90 project was trying to answer fundamental questions about the rifting process: these problems will be addressed in the next section.

8.3 Geodynamic Implications

To enable discussion of the implications resulting from the present interpretation, a summary of the velocity models from KRISP 90 is presented in the fence diagram in Fig. 8.1 (modified from Keller et al., 1994b).

It is also necessary to discuss the implications resulting from the interpretation of the other seismic profiles and the teleseismic results. Model details from KRISP 90 profiles were presented in Section 2.3, therefore the implications alone are discussed here.

The Kerio Valley

The depth and width of the Kerio trough has led to the hypothesis of a separate episode of rifting during the Oligocene. Examples of narrow (40 km) and deep (up to 8 km) basins, bounded by high-angle faults are modelled in the western branch of the
Figure 8.1: Fence diagram summarizing the models from the KRISP 90 seismic studies. For line locations see Fig. 2.1; relative velocity variations from teleseismic delay time studies across the rift at 1°S (after Keller et al., 1994b).
East African Rift System, located in ‘old, cold’ continental lithosphere with a low rate of extension (Ebinger et al., 1991). Could the Kerio Rift have been developed before a significant thermal anomaly reached the central part of the Kenya Rift?

In the central Kenya Rift, the orientation and dip of the Elgeyo fault might simply be due to reactivation of a pre-existing weakness in the basement rocks (Hetzel and Strecker, 1994 and Section 1.3.1). However, as pointed out by Bosworth and Morley (1994), ‘there appeared to be some correlation between heat flow and the boundary fault angle, the lower angle faults being associated with higher heat flow’. In the Mesozoic Anza basin the appearance of basins bounded by high-angle, non rotational faults is related to the last stage of thermal subsidence in the Late Cretaceous/Early Tertiary. Bosworth and Morley (1994) propose an analogy between those basins and the Turkana basin responding to Oligocene/Early Miocene rifting in the Northern Kenya Rift. Not enough is known about the Kerio structure to fit its evolution into an early stage of the Tertiary rifting episode, but it is possible it could be the result of Stage 1 in the process of rift evolution as proposed by Rosendahl (1987).

**The Shape of the Rift Infill**

The asymmetric shape of the rift infill can be related to the advanced stage of rifting reached in the central Kenya Rift. Although at the surface it appears the morphology is of a full graben structure, the depth to the basement implies that the subsidence started with a western dip. As discussed by Keller et al. (1994b), this leads to the conclusion that the upper crust is mainly thinned by simple shear along normal faults. Flexural cantilever modelling has been applied to the northern rift in the light of the KRISP 90 results, demonstrating that extension by simple shear in the brittle upper crust is a viable model also for the Lake Turkana region (Hendrie et al., 1994).

The seismic velocity in the lower volcanic unit of line D indicates the possibility of episodes of mafic intrusion/dyke swarms at shallow depth (between c.2 and 4 km depth) beneath the centre of the graben.

The presence of volcanic activity under the rift axis at the latitude of line D is suggested from the tomography results (Ritter and Achauer, 1994) which modelled high velocity bodies above 5 km depth.
The ANC - MOB Relation and the Mafic Intrusions

The Tibetan-style continent-continent collision (Shackleton and Ries, 1984) is recognized in the significant differences found in the crustal signature on the two sides of the contact between the ANC and MOB terranes extending to lower crustal depth.

The high velocity block to the east of BAS may be indicative of the reworking of Archaean terranes during the MOB orogeny.

It has been suggested that the boundary of the Archaean Craton extends eastward at depth (Le Bas, 1987; Smith, 1994). Despite the fact that no clear indication of such a dipping contact has been derived from the seismic data, the joint interpretation of both the seismic and gravity data requires an eastward dipping interface between a normal (2.7 g/cm$^3$) ANC and a higher density (2.8 g/cm$^3$) MOB at upper crustal level. Preliminary interpretation of the KRISP 94 cross-rift line, at the latitude of Lake Magadi, shows the presence of a low velocity block overlying the ANC. This may represent a slice of MOB overthrust westward during the collision (Birt et al., 1994).

The subdivision into sectors of the Kenyan basement rocks postulated by Mosley (1993) can be recognised in the seismic model.

The signature of the ANC and MOB contact might extend to lower crustal depths. Across the rift the interface detected between 24 and 29 km depth is continuous within the ANC, while in the MOB it is visible only beneath the axial line. However, its continuation along the axis from the south of the rift toward the north "may not be real" (Keller et al., 1994b); in this case, in the south the reflector could still be part of the ANC buried under the MOB.

Mafic intrusions in the shape of sills, as under CHF along the flank line (Prodehl et al., 1994a) represented by diffuse zones of velocity higher than in the surrounding basement have been related to the rifting process and the rising of magma from the mantle to high-level in the upper crust (Macdonald, 1994). These intrusions might also explain why velocities in the upper crust along the axis of the rift are not significantly lower than those beneath the flanks, as might have been expected from the higher temperatures under the axis (Mechie et al., 1994b).
Underplating?

KRISP 90 provided some unexpected results concerning the lower crust. It is supposed that the lower crust thins by extension during rifting, but magmatic addition to the crust could keep the Moho depth unaltered (Keller et al., 1994b and ref. therein).

Along the rift the thickness of the lower crustal layer decreases from south to north consistent with the extension rate being higher in the north. However, the volume of volcanics extruded during the rifting process requires magmatic material to be added to the lower crust (Karson and Curtis, 1989; Macdonald, 1994), in which case the same amount of extension can be hypothesised along the rift, with underplating being more significant in the south. Keller et al. (1994b) try to reconcile the two views suggesting that the volume of underplated material has not replaced the volume lost by extension, this being greater in the north. Morley (1994) suggests that the volume of magma added to the crust could have been overestimated or has been lost to the asthenosphere.

Across the rift the lower crust thins beneath the axis, and the indefinite presence of the interface at c.24 km depth under the axis may indicate the top of a sill-like body intruded into the crust.

The relatively low Poisson's ratio in the mid-crust might be explained by the presence of fluids at low pore pressure, which may be liberated during the prograde metamorphism induced in the lower crust by the process of underplating (Wenzel and Sandmeier, 1992); however, neither is an LVZ detected in the mid-crust, nor are high resolution near-vertical data available to identify the presence of bright spots, related to the accumulation of fluids. On the other hand, the results of magnetotelluric traverses across the rift valley (Section 1.5.6) suggest fluids may be present in the crust above 20 km depth.

It has been suggested that extension acted in the lower crust via a pure shear mechanism (Keller et al., 1994b), consistent with the modelling by Hendrie et al. (1994) for the northern Kenya Rift.

It would appear that the rifting initiated in the thickest portion of the lithosphere, where a low strength region characterized by a high percentage of weak minerals created a preferential site for opening (Vink et al., 1984; Dunbar and Sawyer, 1988). The suture zone between the ANC and the MOB could have been a favoured point of weakness along
which rupture propagated (Vink et al., 1984).

This phenomenon has been related to the Kusznir and Park (1987) model of rifting at zones of low strength in the lithosphere. A thick crust proves to be weaker than a thin crust. As extension proceeds: (1) the geotherm increase and (2) crustal thinning act in opposition, the first softening and the second hardening the lithosphere. The strain rate (Kusznir and Park, 1987) determines how quickly the process will develop: a fast strain rate produces a net weakening; a slow strain rate produces net strengthening, with geometries as illustrated in Fig. 8.2. It follows that from studies of the strain rate and rift geometries the tendency of an extension process to fail may be modelled.

Although no specific study has been done on the strain rate in the Kenya Rift, in
the Turkana Rift, Morley (1994) has suggested the presence of an intermediate to low strain rate; the amount of extension with time suffered by the Kenya Rift is less than the Turkana Rift, and it may be assumed that a model for low strain rate can be applied also in the Kenya Rift. The narrowness of the Kenya graben however is more compatible with the fast strain rate geometry. Modelling of the rifting process in Kenya has to take into account that the whole African continent is in compression (Keller et al., 1994b), therefore the crust will not respond passively to extensional stress; instead, doming of the crust might generate the tensional state of stress around the EAP.

Moho and Upper Mantle Upwelling

In this section will be considered in order: the structure of the Moho, the character of the P and S Moho reflections; the velocity field in the upper mantle and the LVZ within it; and the thermal anomaly related to the LVZ whose shape and size will be also discussed.

Moho topography mirroring the rift graben has been used as an indication that the extension mechanism is 'pure shear' in the lower crust and upper mantle. However the topographic high in the Moho is, from the present work, asymmetric, possibly indicating that the lithosphere under the MOB is more easily deformed than under the ANC.

The slightly lower density in the lithospheric mantle under the eastern flank in comparison with the western flank might also be indicative of a different origin. Archaean lithosphere is believed to be colder and thicker than Proterozoic lithosphere (Durrheim and Mooney, 1994). Heat flow increases away from the centre of the craton, as though the thicker lithosphere under the craton forces mantle heat to escape through the thinner lithosphere of the adjoining mobile belts (Nyblade and Pollack, 1993).

The effect of rifting and extension at Moho depth may be examined via other observations. White (1988) pointed out that in extensional regimes the addition of igneous material to the crust changes the Moho character from a passive strain marker to a new interface at the base of the underplated igneous section and in some cases the Moho deepens. As the character of the Moho reflection c does not seem to change across the rift, such a metamorphosis might not have happened in the central Kenya Rift. The S-wave Moho reflection suggests another implication: a clear and high amplitude u phase supports the idea that the presence of melt in the uppermost mantle is not widespread.
A significant anomalously low velocity zone in the uppermost mantle has been modelled underneath the rift axis, and interpreted in terms of the presence of up to 6% partial melt, indicating the effect of the extensional process underneath the Moho (Achauer et al., 1994; Slack and Davis, 1994; Mechie et al., 1994b). Nevertheless, preliminary results from the KRISP 94 experiment across the southern portion of the rift (Lake Magadi) do not seem to require the presence of such an LVZ in the uppermost mantle (Birt et al., 1994). This supports the contention that the magmatism associated with rifting in the central Kenya Rift originated in the north, and has progressed southwards (Smith, 1994). Moreover, it has been suggested that the thermal anomaly interpreted to occur beneath the Kenya Rift has migrated from west to east. This is suggested from study of the nephelinitic magmas in the Kenya Rift. Such magmas represent small degrees of partial melting from garnet-bearing mantle sources which have rapidly ascended without fractionation in the crust. Their occurrence at the surface can provide an indication of the source migration with time (Le Bas, 1980; Macdonald, 1994).

The asymmetry of the LVZ and its relation to the eastward shift of the thermal anomaly has been discussed by Morley (1994): he suggests it is seen on travel time residual data (Halderman and Davis, 1991), and the LVZ might have 'an initially steep east margin ... modified by eastward migration of the convecting thermal anomaly'.

The size of the anomaly could be related to a small diapir from a 'weak' plume. This may have originated from the 650 km depth transition in the mantle and therefore have a smaller plume head diameter before reaching the lithosphere (≈300 km) than that originating from the core-mantle boundary (≈1000 km) the latter of the type generating Continental Flood Basalts (CFB; e.g., Deccan Traps of India and Karoo of southern Africa) (Griffiths and Campbell, 1990; Campbell and Griffiths, 1990). It has to be said that studies of lithospheric thinning (Ebbing et al., 1989) suggest it is reasonable to treat the areas surrounding the East African Plateau (EAP) as a single rift system; the plume impacting beneath the EAP (Thompson and Gibson, 1994) could have generated the diapir beneath the Kenya Rift.

**Mantle Reflector**

Anisotropy in a cold mantle lithosphere under the western margin of the rift below 60
km could be related to shear stress along more or less horizontal planes directed radially away from the apex of the Kenya Dome. Preferential orientation of olivine could be combined with depletion of basalt and a phase change from spinel to garnet peridotite to account for the high velocity value in the uppermost mantle.

Alternatively, the reflector could be a signature of a shear zone in the mantle resulting from the collision which formed the MOB, or of continental lower crust eclogitized following the same orogeny. Moreover, it could be the top of an eclogitic remnant of oceanic crust subducted during the collision: this implies the presence of a late Precambrian ocean in central Kenya which is still debated (Section 1.3.1).

It cannot be ruled out that the LVZ in the upper mantle modelled along the axis of the rift and interpreted in terms of anomalous mantle material may extend laterally into an LVZ under the rift's western shoulder, but this is not well constrained.

The reflector could also represent the top of a melt fraction percolated into the lithosphere from below and frozen along an isotherm (McKenzie, 1989) during the eastward migration of the thermal anomaly. More constraints on the presence of the LVZ, the thickness and lateral extent of the high velocity layer are required to discriminate between the different hypotheses.

It is interesting to note that preliminary observations of the KRISP 94 record section from VIC into the southern rift cross-line show energy at c.300 km offset immediately following the d phase (P.K.H. Maguire, pers. comm., 1995) possibly equivalent to that defining the reflector discussed here.

8.3.1 Geodynamic Model

In Section 1.2.1 it was mentioned that most authors consent to the Kenya Rift being an 'active' type. A number of results from the present study support this contention: the narrowness of the LVZ under the rift axis; the presence of the low velocity anomalous mantle; the evidence of convective movements in the asthenosphere.

It has been proposed (Maguire et al., 1994) that the Kenya Rift tectonic activity results from a thermal process acting underneath the centre of the Nyanza Craton forming the EAP. Ebinger et al. (1989) showed that there is overcompensation under East Africa,
suggesting the uplift is, in part, compensated dynamically. According to the model produced by Courtney and White (1986), this is further evidence for the presence of a plume under Kenya, as well as the presence of large volumes of melt and partial melt (Latin et al., 1993 and Sections 1.3, 1.4).

A geodynamic model may thus be produced for the central Kenya Rift as follows:

Initial weakness of the lithosphere localised at the MOB collision zone locates the nucleation point of the rifting process; mantle flow from beneath the EAP induces upwelling, the creation of the Kenya Dome, faulting and volcanism. The upwelling is asymmetric and spreads under the Proterozoic lithosphere being younger and weaker than the adjoining ANC. An initial fast strain rate induces a narrow rift and reactivates steep boundary faults in the MOB; the pre-existing zone of linear structures was an important controlling influence on the location of the Kenya Rift.

The strain rate may have decreased with time since the amount of extension observed today is not consistent with the 'softening' model of Kusznir and Park (1987). It is possible that the extensional regime in the rift is opposed by the compressional regime surrounding it, thus preventing rapid continental break-up and broad scale extension (Bott, 1992).

The Kenya Rift is at an early stage of its evolution because heating from the mantle has not diffused laterally and to the surface, because the LVZ is narrow, and because the rift is still seismically active. The rifting process may be considered anomalous because the lower crust seems thinner than expected for the amount of volcanics extruded, i.e. there is some controversy concerning the amount of underplated material.

From seismic evidence up to 5-6% of melt has accumulated in the uppermost mantle; this percentage of melt occurs only locally because it does not represent the percentage required to explain the magmas' composition (for this only 3% is required). The melt penetrating to upper mantle depths could arise from diapirs extracted from a 'weak' plume under the EAP. Melts might radiate from the plume/diapir head consistent with the presence of the LVZ detected along the axis of the rift and the anisotropy invoked to explain the high velocity values in the uppermost mantle radiating from the apex of the Kenya Dome.

Archaean lithosphere might still be preserved under the rift's western flank. There are
very few xenolith analyses from the Nyanza Craton. Those that exist seem to indicate that
there is at least local enrichment in the upper mantle, above 75 km (Ito, 1986). It cannot
be ruled out that down to a depth of 60 km the 'cold keel' typical of an Archaean craton
provides the observed seismic signature, with higher velocities than under the Proterozoic
terrane.

8.3.2 Comparison with Other Rifts

Keller et al. (1994b) provide an exhaustive examination of the four most studied Cenozoic
Rifts: the Rhine graben section of the Central European Rift, the Rio Grande Rift, the
Baikal Rift and the Kenya Rift. For the Baikal Rift however, only a limited amount of
data have been published.

The main points are considered here:

- a feature common to the four rifts is a crust thinned by extension; a correlation
can be found between topography, width of rift zone and crustal thinning by pure shear
extension. In the Kenya Rift (from the apex of the Kenya Dome at Lake Naivasha to the
north; Mechie et al., 1994b) and in the Rio Grande Rift the maximum extension occurs
where the crust appears thinnest, the width is greatest and the elevation of the flanks and
valley floor are lowest (e.g., Keller et al., 1991). The same relationship of extension and
crustal thinning applies at the southern end of Central European Rift (Edel et al., 1975;
Prodehl, 1981);

- considering the lower crust, in the Rio Grande Rift there is a decrease in the axial
lower crustal thickness with increase in extension. The same happens in the Kenya Rift,
where the phenomenon is more pronounced. However, while in the Kenya Rift the thinning
is concentrated in the lower crust, in the Rio Grande Rift it occurs in all crustal layers,
involving the whole crust. The lower crustal velocities do not suggest underplating in the
Rio Grande Rift (e.g., Keller et al., 1991). In the Central European Rift no thinning is
detected (Prodehl, 1981). Thus, no unique relationship can be found between rifting and
lower crustal underplating;

- the overall lithospheric structure shows a gradual transition from flank to axis, in
the Southern Rio Grande Rift (e.g., Davis, 1991) and an abrupt transition in the Kenya,
CHAPTER 8. DISCUSSION AND CONCLUSIONS

Central European (e.g., Prodehl et al., 1992) and the Baikal Rifts (e.g., Ruppel et al., 1992). This difference has been explained in terms of the differences in the pre-rift lithosphere: the Rio Grande Rift initiated in a hot back-arc regime, the Central European Rift in Variscan orogenic terrain, and the Kenya and Baikal Rifts in cool cratonic lithosphere;

- the upper mantle upwarps beneath the axis of the rifts is another common point between the four rifts. However, in the Kenya and Rio Grande Rifts it is associated with a LVZ interpreted in terms of the presence of partial melt. The combination of LVZ and limited crustal thinning results in the two rifts being included in the category of 'active' rifts. The Rhine Graben lacks the mantle low velocity, and has been classified as a 'passive' rift. A comparative study for these three rifts has been presented by Davis et al. (1993). The Lake Baikal Rift is associated with a broad asthenospheric upwarp. A seismic reflection experiment was undertaken in 1992 in the Baikal Rift, but only preliminary data have been published (Scholz et al., 1993).

Gao et al. (1994a) suggest that the Rio Grande and Baikal upper mantle anomalies not being aligned with the graben demonstrate that the plates are in motion; the equivalent Kenya Rift symmetry would confirm that the African plate is almost stationary;

- concerning the mechanism of rifting, for all four rifts considered a symmetric overall crustal structure, with asymmetric basins in the upper crust are observed. Some of the basins are very deep: in the Rhine Graben there is up to 3-7 km of material with P-wave velocity from 2.8 to 4.0 km/s (Prodehl et al., 1976), whereas in the Lake Baikal Rift up to 10 km of sediments have recently been interpreted through modelling (Scholz et al., 1993). The shallow structure of the Kenya Rift has been studied in detail via seismic reflection only beneath Lake Turkana (Dunkelman et al., 1989; Morley et al., 1992; Hendrie et al., 1994) and in the Kerio Valley (Pope, 1992). These and KRISP data indicate the presence of many small complex basins and some large asymmetrical ones;

- upper mantle anisotropy has been revealed in south-western Germany, in an area including the Rhine Graben (Fuchs, 1983). Beneath the Baikal Rift it is suggested olivine crystals are orientated horizontally and normal to the rift's axis, similar to that resulting from mantle flow beneath the mid-ocean rifts; in the south of the same rift the orientation of the olivine is related to the stress induced by the India-Asia collision (Gao et al.,...
CHAPTER 8. DISCUSSION AND CONCLUSIONS

1994b). This attests to the possibility of different mechanisms acting in the same area on the direction of flow; caution should therefore be used in the interpretation of anisotropy in the upper mantle beneath the Kenya Rift, and the high velocity modelled under the western flank might not be related to the velocity structure under the axis.

It is of interest to note particular features in other regions of continental extension relevant to the model derived from the cross-rift line study:

In the north-western Basin and Range Province in the western USA: mafic intrusions are suggested to exist in the crust, where maximum extension occurs associated with a possible thermal anomaly; however, as suggested in the present study, the velocities are also not apparently affected by the thermal effects (Benz et al., 1990). An LVZ in the mantle is detected in a region that suffered compression and thickening of the crust prior to the large-scale Cenozoic extension (Benz et al., 1990), and where there is evidence for active magmatic underplating at the base of the crust despite the scarcity of volcanic activity in the last 6 Ma (Jarchow et al., 1993). Similarly, in the Basin and Range region of south-eastern California, Wilson et al. (1991) describe features similar to the ones modelled beneath the Kenya Rift:

- dyke swarms in the upper crust and a high velocity (6.4 km/s) block at 5 km depth, interpreted as a mafic intrusion;
- crustal thickness which does not reflect the amount of extension either (1) because the crust was of variable thickness before extension or (2) because new material has been added by lateral ductile flow and magmatism.

Of particular interest is the widespread presence of mafic intrusions and the evidence of underplating in this continental area. This has not been proved in the Kenya Rift but is suggested by the results from KRISP 90. As in the Kenya Rift, rifting seems to develop favourably in thick and weak portions of the crust.

The crustal structure underlying the Miocene CFB Columbia Plateau, north of the Basin and Range Province, is described by Catchings and Mooney (1988) as being typical of a continental rift. Evidence for an Eocene rifting episode includes: (1) a central graben, filled with more than 5 km of sediments beneath 1.5 to 3.5 km of continental flood basalts, (2) high velocity (7.5 km/s) thickened lower crust, (3) palaeomagnetic data requiring
extension in the area and (4) the presence of highly extended Eocene metamorphic core complexes. An upper mantle reflector at c.50 km depth has been modelled as the top of a LVZ related to the presence of partial melt; the LVZ as well as points (1) and (2) are also proposed to exist under the Kenya Rift.

Inversion of teleseismic arrival times in the Central African Shear Zone in Cameroon resulted in evidence for a narrow asthenospheric upwelling from c.190 to c.120 km depth and 'frozen-in' anisotropy in the subcrustal lithosphere. It is proposed that the West African Rift represents reopening of a suture zone and that the anisotropy derives from the olivine orientation in the oceanic lithosphere which existed prior to suturing (Fairhead and Binks, 1991; Plomerová et al., 1993). It is tempting to propose a similar model for the Kenya Rift, but there is no definite evidence for the presence of oceanic crust before the ANC-Kibaran craton collision started.

8.4 Conclusions

Kinematic and dynamic modelling of the P- and S-wave fields recorded during KRISP 90 along a seismic profile across the rift has been integrated with traveltime inversion and gravity modelling along the same profile.

The result is a picture down to c.60 km, at which depth detailed processing of relevant seismic phases and petrological modelling has also been undertaken.

Some of the most significant results are catalogued here:

- the crust and upper mantle (lithosphere) are significantly different for the ANC and the MOB: variations are seen in the upper crustal P and S velocity structure; a lower crustal boundary is clearly identified only in the ANC portion of the crust; the crust appears thickened on passing from ANC to MOB; the average upper mantle velocity is higher in the ANC; an upper mantle reflector is seen in the ANC portion of the lithospheric mantle;

- the orogenic signature is still preserved in the upper crust; a high velocity block in the MOB terrane could be indicative of orogenic reworking of older terranes; orogenic reactivated trends are visible in the present day structural pattern; a mid crustal and
lower crustal step-like structure across the ANC-MOB boundary testifies to the westward vergence of the orogen;

- the thermal anomaly and the influence of partial melt is of limited extent: the maximum attenuation of seismic waves takes place beneath the rift at upper crustal depths, whereas mantle P phases are not attenuated travelling beneath the rift, nor S-waves in the lower crust;

- the general structure of the crust across the rift is consistent with a model of extension via simple shear in the upper crust and pure shear in the lower crust and upper mantle;

- a sharp lateral transition in the upper mantle is observed between rift axis and flanks; to the west a high velocity (low temperature) region is immediately adjacent to an anomalously low velocity (higher temperature) region beneath the rift axis, as required by the modelling of the upper mantle reflector;

- assuming the hypothesis that subduction took place prior to collision forming the MOB, the reflector at c.60 km depth under the western flank may represent relict oceanic subducted crust not remelted by the rising mantle anomaly, or eclogitized continental crust. Otherwise, the reflector might represent the top of a frozen melt layer preserving its anisotropic state indicating an horizontal direction of flow away from the apex of the uplifted region;

- the size of the mantle thermal anomaly suggests the presence of a small diapir under the Kenya Rift, radiating from a ‘weak’ plume seated under the East African Plateau which dynamically maintains its elevation; the diapir appears to have spread asymmetrically towards the MOB lithosphere to the east.
Writing notes by the Kerio River
Bibliography


Bateman, H., 1910. The solution of the integral equation connecting the velocity of propagation of an earthquake-wave in the interior of the earth with the disturbance travels to take off to the different stations on the earth's surface. Phil. Mag., 19, 576-587. Also in Phys. Z., 11, 96-99.


BIBLIOGRAPHY

331


BIBLIOGRAPHY

BIBLIOGRAPHY


BIBLIOGRAPHY


Pope, D.A., 1992, Analyses and interpretation of the seismic reflection profiles from the Kerio Valley, Kenya Rift, Msc. dissertation, Univ. of Leeds, UK.


BIBLIOGRAPHY


BIBLIOGRAPHY


### KRISP 90 Recording Site Coordinates - Line D

<table>
<thead>
<tr>
<th>SITE</th>
<th>LATITUDE-N</th>
<th>LONGITUDE-E</th>
<th>HEIGHT m</th>
<th>SITE</th>
<th>LATITUDE-N</th>
<th>LONGITUDE-E</th>
<th>HEIGHT m</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.48</td>
<td>36.04.47</td>
<td>649</td>
<td>0.47</td>
<td>36.37.69</td>
<td>1845</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.49</td>
<td>36.01.75</td>
<td>707</td>
<td>0.44</td>
<td>36.36.55</td>
<td>1830</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.48</td>
<td>37.59.51</td>
<td>695</td>
<td>0.42</td>
<td>36.35.21</td>
<td>1855</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.49</td>
<td>37.57.45</td>
<td>707</td>
<td>0.43</td>
<td>36.33.65</td>
<td>1875</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.49</td>
<td>37.55.05</td>
<td>732</td>
<td>0.43</td>
<td>36.32.72</td>
<td>1915</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.48</td>
<td>37.52.43</td>
<td>777</td>
<td>0.44</td>
<td>36.31.66</td>
<td>1970</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.47</td>
<td>37.49.75</td>
<td>777</td>
<td>0.45</td>
<td>36.30.38</td>
<td>1950</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.46</td>
<td>37.47.46</td>
<td>823</td>
<td>0.45</td>
<td>36.29.06</td>
<td>1845</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.45</td>
<td>37.44.50</td>
<td>915</td>
<td>0.45</td>
<td>36.28.16</td>
<td>1840</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.44</td>
<td>37.42.57</td>
<td>869</td>
<td>0.45</td>
<td>36.27.35</td>
<td>1855</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.43</td>
<td>37.40.19</td>
<td>838</td>
<td>0.45</td>
<td>36.26.79</td>
<td>1860</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.42</td>
<td>37.37.69</td>
<td>884</td>
<td>0.46</td>
<td>36.26.26</td>
<td>1865</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.44</td>
<td>37.35.19</td>
<td>915</td>
<td>0.46</td>
<td>36.25.91</td>
<td>1870</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.46</td>
<td>37.33.00</td>
<td>911</td>
<td>0.46</td>
<td>36.25.35</td>
<td>1805</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.47</td>
<td>37.30.40</td>
<td>936</td>
<td>0.46</td>
<td>36.24.78</td>
<td>1760</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.48</td>
<td>37.28.23</td>
<td>990</td>
<td>0.46</td>
<td>36.24.26</td>
<td>1740</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.51</td>
<td>37.26.13</td>
<td>1120</td>
<td>0.46</td>
<td>36.23.73</td>
<td>1715</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.51</td>
<td>36.22.98</td>
<td>1190</td>
<td>0.46</td>
<td>36.23.11</td>
<td>1700</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.52</td>
<td>37.20.71</td>
<td>1190</td>
<td>0.46</td>
<td>36.22.61</td>
<td>1670</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.54</td>
<td>37.18.40</td>
<td>1230</td>
<td>0.47</td>
<td>36.22.07</td>
<td>1660</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.57</td>
<td>37.15.92</td>
<td>1260</td>
<td>0.47</td>
<td>36.21.55</td>
<td>1620</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.56</td>
<td>37.13.99</td>
<td>1195</td>
<td>0.47</td>
<td>36.21.06</td>
<td>1600</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.55</td>
<td>37.11.54</td>
<td>1175</td>
<td>0.47</td>
<td>36.20.48</td>
<td>1560</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.55</td>
<td>37.08.68</td>
<td>1270</td>
<td>0.47</td>
<td>36.19.88</td>
<td>1440</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.54</td>
<td>37.06.29</td>
<td>1180</td>
<td>0.48</td>
<td>36.19.36</td>
<td>1430</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.53</td>
<td>37.03.50</td>
<td>1250</td>
<td>0.47</td>
<td>36.18.64</td>
<td>1390</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.49</td>
<td>37.05.85</td>
<td>1190</td>
<td>0.47</td>
<td>36.18.28</td>
<td>1330</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.51</td>
<td>37.05.07</td>
<td>1220</td>
<td>0.48</td>
<td>36.17.74</td>
<td>1290</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.52</td>
<td>37.04.61</td>
<td>1245</td>
<td>0.48</td>
<td>36.17.07</td>
<td>1215</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.53</td>
<td>37.02.54</td>
<td>1265</td>
<td>0.48</td>
<td>36.16.59</td>
<td>1200</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.52</td>
<td>37.01.44</td>
<td>1290</td>
<td>0.48</td>
<td>36.16.13</td>
<td>1185</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.50</td>
<td>37.00.03</td>
<td>1315</td>
<td>0.48</td>
<td>36.15.60</td>
<td>1175</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.49</td>
<td>36.58.76</td>
<td>1425</td>
<td>0.48</td>
<td>36.15.07</td>
<td>1165</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.46</td>
<td>36.57.88</td>
<td>1445</td>
<td>0.49</td>
<td>35.14.52</td>
<td>1155</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.46</td>
<td>36.56.26</td>
<td>1540</td>
<td>0.48</td>
<td>36.13.87</td>
<td>1155</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.46</td>
<td>36.55.16</td>
<td>1660</td>
<td>0.48</td>
<td>36.13.46</td>
<td>1155</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.47</td>
<td>36.54.07</td>
<td>1710</td>
<td>0.47</td>
<td>36.12.95</td>
<td>1155</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.48</td>
<td>36.52.73</td>
<td>1720</td>
<td>0.46</td>
<td>36.12.33</td>
<td>1120</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.49</td>
<td>36.51.70</td>
<td>1740</td>
<td>0.46</td>
<td>36.11.81</td>
<td>1070</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.50</td>
<td>36.50.46</td>
<td>1745</td>
<td>0.45</td>
<td>36.11.29</td>
<td>1060</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.51</td>
<td>36.49.27</td>
<td>1760</td>
<td>0.45</td>
<td>36.10.75</td>
<td>1035</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.52</td>
<td>36.48.03</td>
<td>1755</td>
<td>0.44</td>
<td>36.10.21</td>
<td>1065</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.54</td>
<td>36.46.89</td>
<td>1770</td>
<td>0.44</td>
<td>36.09.67</td>
<td>1055</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.55</td>
<td>36.44.88</td>
<td>1800</td>
<td>0.43</td>
<td>36.09.14</td>
<td>1060</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.50</td>
<td>36.43.92</td>
<td>1800</td>
<td>0.43</td>
<td>36.08.76</td>
<td>1080</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.50</td>
<td>36.42.98</td>
<td>1800</td>
<td>0.42</td>
<td>36.08.02</td>
<td>1030</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.49</td>
<td>36.41.92</td>
<td>1790</td>
<td>0.42</td>
<td>36.07.52</td>
<td>1015</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.48</td>
<td>36.40.50</td>
<td>1800</td>
<td>0.42</td>
<td>36.06.98</td>
<td>1025</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.48</td>
<td>36.38.92</td>
<td>1840</td>
<td>0.42</td>
<td>36.06.32</td>
<td>1055</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
### Appendix A

**KRISP 90 Recording Site Coordinates - Line D**

<table>
<thead>
<tr>
<th>SITE</th>
<th>LATITUDE-N</th>
<th>LONGITUDE-E</th>
<th>HEIGHT (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>99</td>
<td>0.42.03</td>
<td>36.05.66</td>
<td>1080</td>
</tr>
<tr>
<td>100</td>
<td>0.41.75</td>
<td>36.05.26</td>
<td>1015</td>
</tr>
<tr>
<td>101</td>
<td>0.44.52</td>
<td>36.04.83</td>
<td>1160</td>
</tr>
<tr>
<td>102</td>
<td>0.44.61</td>
<td>36.04.32</td>
<td>1120</td>
</tr>
<tr>
<td>103</td>
<td>0.44.63</td>
<td>36.03.76</td>
<td>1080</td>
</tr>
<tr>
<td>104</td>
<td>0.44.42</td>
<td>36.03.21</td>
<td>1060</td>
</tr>
<tr>
<td>105</td>
<td>0.44.01</td>
<td>36.02.66</td>
<td>995</td>
</tr>
<tr>
<td>106</td>
<td>0.37.84</td>
<td>36.02.13</td>
<td>975</td>
</tr>
<tr>
<td>107</td>
<td>0.37.38</td>
<td>36.01.61</td>
<td>1015</td>
</tr>
<tr>
<td>108</td>
<td>0.36.74</td>
<td>36.01.05</td>
<td>1160</td>
</tr>
<tr>
<td>109</td>
<td>0.36.46</td>
<td>36.00.51</td>
<td>1080</td>
</tr>
<tr>
<td>110</td>
<td>0.35.72</td>
<td>36.00.03</td>
<td>1060</td>
</tr>
<tr>
<td>111</td>
<td>0.34.60</td>
<td>35.59.70</td>
<td>1005</td>
</tr>
<tr>
<td>112</td>
<td>0.34.12</td>
<td>35.59.46</td>
<td>1020</td>
</tr>
<tr>
<td>113</td>
<td>0.33.09</td>
<td>35.59.27</td>
<td>1025</td>
</tr>
<tr>
<td>114</td>
<td>0.31.98</td>
<td>35.59.12</td>
<td>1035</td>
</tr>
<tr>
<td>115</td>
<td>0.30.43</td>
<td>35.58.42</td>
<td>1090</td>
</tr>
<tr>
<td>116</td>
<td>0.30.56</td>
<td>35.57.81</td>
<td>1105</td>
</tr>
<tr>
<td>117</td>
<td>0.28.32</td>
<td>35.57.35</td>
<td>1128</td>
</tr>
<tr>
<td>118</td>
<td>0.28.51</td>
<td>35.56.81</td>
<td>1143</td>
</tr>
<tr>
<td>119</td>
<td>0.28.59</td>
<td>35.56.26</td>
<td>1165</td>
</tr>
<tr>
<td>120</td>
<td>0.28.39</td>
<td>35.55.70</td>
<td>1204</td>
</tr>
<tr>
<td>121</td>
<td>0.29.74</td>
<td>35.55.18</td>
<td>1189</td>
</tr>
<tr>
<td>122</td>
<td>0.28.89</td>
<td>35.54.48</td>
<td>1250</td>
</tr>
<tr>
<td>123</td>
<td>0.29.79</td>
<td>35.54.23</td>
<td>1226</td>
</tr>
<tr>
<td>124</td>
<td>0.29.60</td>
<td>35.53.61</td>
<td>1256</td>
</tr>
<tr>
<td>125</td>
<td>0.29.36</td>
<td>35.53.01</td>
<td>1281</td>
</tr>
<tr>
<td>126</td>
<td>0.29.35</td>
<td>35.52.45</td>
<td>1311</td>
</tr>
<tr>
<td>127</td>
<td>0.29.20</td>
<td>35.51.93</td>
<td>1320</td>
</tr>
<tr>
<td>128</td>
<td>0.28.88</td>
<td>35.51.39</td>
<td>1387</td>
</tr>
<tr>
<td>129</td>
<td>0.28.67</td>
<td>35.50.86</td>
<td>1403</td>
</tr>
<tr>
<td>130</td>
<td>0.29.98</td>
<td>35.50.25</td>
<td>1540</td>
</tr>
<tr>
<td>131</td>
<td>0.29.93</td>
<td>35.49.93</td>
<td>1585</td>
</tr>
<tr>
<td>132</td>
<td>0.30.18</td>
<td>35.49.27</td>
<td>1728</td>
</tr>
<tr>
<td>133</td>
<td>0.29.65</td>
<td>35.48.69</td>
<td>1860</td>
</tr>
<tr>
<td>134</td>
<td>0.29.10</td>
<td>35.48.11</td>
<td>1921</td>
</tr>
<tr>
<td>135</td>
<td>0.28.56</td>
<td>35.47.72</td>
<td>1997</td>
</tr>
<tr>
<td>136</td>
<td>0.29.12</td>
<td>35.47.04</td>
<td>1921</td>
</tr>
<tr>
<td>137</td>
<td>0.28.81</td>
<td>35.46.52</td>
<td>1857</td>
</tr>
<tr>
<td>138</td>
<td>0.29.48</td>
<td>35.45.96</td>
<td>1967</td>
</tr>
<tr>
<td>139</td>
<td>0.29.95</td>
<td>35.45.47</td>
<td>1997</td>
</tr>
<tr>
<td>140</td>
<td>0.29.88</td>
<td>35.44.88</td>
<td>2060</td>
</tr>
<tr>
<td>141</td>
<td>0.29.91</td>
<td>35.44.34</td>
<td>1970</td>
</tr>
<tr>
<td>142</td>
<td>0.30.19</td>
<td>35.44.82</td>
<td>2080</td>
</tr>
<tr>
<td>143</td>
<td>0.29.75</td>
<td>35.43.28</td>
<td>1820</td>
</tr>
<tr>
<td>144</td>
<td>0.29.34</td>
<td>35.42.66</td>
<td>1720</td>
</tr>
<tr>
<td>145</td>
<td>0.29.12</td>
<td>35.42.21</td>
<td>1660</td>
</tr>
<tr>
<td>146</td>
<td>0.28.24</td>
<td>35.41.76</td>
<td>1565</td>
</tr>
<tr>
<td>147</td>
<td>0.27.77</td>
<td>35.41.09</td>
<td>1430</td>
</tr>
</tbody>
</table>
### KRISP 90 Recording Site Coordinates - Line D

<table>
<thead>
<tr>
<th>SITE</th>
<th>LATITUDE-N</th>
<th>LONGITUDE-E</th>
<th>HEIGHT m</th>
<th>SITE</th>
<th>LATITUDE-N</th>
<th>LONGITUDE-E</th>
<th>HEIGHT m</th>
</tr>
</thead>
<tbody>
<tr>
<td>197</td>
<td>0.11.61</td>
<td>35.00.71</td>
<td>1860</td>
<td>224</td>
<td>0.00.21</td>
<td>34.31.79</td>
<td>1440</td>
</tr>
<tr>
<td>198</td>
<td>0.11.37</td>
<td>34.59.53</td>
<td>1870</td>
<td>225</td>
<td>0.00.25S</td>
<td>34.30.64</td>
<td>1420</td>
</tr>
<tr>
<td>199</td>
<td>0.11.30</td>
<td>34.58.50</td>
<td>1780</td>
<td>226</td>
<td>0.01.40S</td>
<td>34.29.51</td>
<td>1340</td>
</tr>
<tr>
<td>200</td>
<td>0.11.48</td>
<td>34.57.58</td>
<td>1745</td>
<td>227</td>
<td>0.03.38S</td>
<td>34.26.48</td>
<td>1370</td>
</tr>
<tr>
<td>201</td>
<td>0.11.49</td>
<td>34.56.80</td>
<td>1700</td>
<td>228</td>
<td>0.03.58S</td>
<td>34.23.90</td>
<td>1345</td>
</tr>
<tr>
<td>202</td>
<td>0.10.49</td>
<td>34.55.62</td>
<td>1700</td>
<td>229</td>
<td>0.04.37S</td>
<td>34.21.89</td>
<td>1320</td>
</tr>
<tr>
<td>203</td>
<td>0.09.62</td>
<td>34.54.50</td>
<td>1720</td>
<td>230</td>
<td>0.05.85S</td>
<td>34.19.36</td>
<td>1340</td>
</tr>
<tr>
<td>204</td>
<td>0.08.96</td>
<td>34.53.28</td>
<td>1665</td>
<td>231</td>
<td>0.05.57S</td>
<td>34.17.75</td>
<td>1265</td>
</tr>
<tr>
<td>205</td>
<td>0.08.07</td>
<td>34.52.15</td>
<td>1710</td>
<td>232</td>
<td>0.06.97S</td>
<td>34.16.48</td>
<td>1235</td>
</tr>
<tr>
<td>206</td>
<td>0.07.28</td>
<td>34.50.71</td>
<td>1680</td>
<td>233</td>
<td>0.07.95S</td>
<td>34.13.81</td>
<td>1195</td>
</tr>
<tr>
<td>207</td>
<td>0.06.61</td>
<td>34.49.90</td>
<td>1700</td>
<td>234</td>
<td>0.08.87S</td>
<td>34.11.20</td>
<td>1165</td>
</tr>
<tr>
<td>208</td>
<td>0.04.86</td>
<td>34.48.90</td>
<td>1700</td>
<td>235</td>
<td>0.10.01S</td>
<td>34.09.30</td>
<td>1190</td>
</tr>
<tr>
<td>209</td>
<td>0.04.48</td>
<td>34.48.02</td>
<td>1720</td>
<td>236</td>
<td>0.11.72S</td>
<td>34.09.22</td>
<td>1140</td>
</tr>
<tr>
<td>210</td>
<td>0.04.00</td>
<td>34.46.66</td>
<td>1700</td>
<td>237</td>
<td>0.24.61</td>
<td>35.34.83</td>
<td>2720</td>
</tr>
<tr>
<td>211</td>
<td>0.03.58</td>
<td>34.45.72</td>
<td>1665</td>
<td>238</td>
<td>0.25.68</td>
<td>35.37.10</td>
<td>1240</td>
</tr>
<tr>
<td>212</td>
<td>0.03.06</td>
<td>34.44.36</td>
<td>1665</td>
<td>239</td>
<td>0.27.79</td>
<td>35.40.22</td>
<td>1340</td>
</tr>
<tr>
<td>213</td>
<td>0.02.51</td>
<td>34.43.04</td>
<td>1640</td>
<td>240</td>
<td>0.30.56</td>
<td>35.57.81</td>
<td>1105</td>
</tr>
<tr>
<td>214</td>
<td>0.02.04</td>
<td>34.42.32</td>
<td>1670</td>
<td>241</td>
<td>0.37.83</td>
<td>36.02.12</td>
<td>1000</td>
</tr>
<tr>
<td>215</td>
<td>0.02.35</td>
<td>34.42.52</td>
<td>1610</td>
<td>242</td>
<td>0.36.45</td>
<td>36.03.10</td>
<td>980</td>
</tr>
<tr>
<td>216</td>
<td>0.01.99</td>
<td>34.40.54</td>
<td>1800</td>
<td>243</td>
<td>0.37.23</td>
<td>36.03.90</td>
<td>980</td>
</tr>
<tr>
<td>217</td>
<td>0.01.95</td>
<td>34.39.56</td>
<td>1570</td>
<td>244</td>
<td>0.38.51</td>
<td>36.04.22</td>
<td>980</td>
</tr>
<tr>
<td>218</td>
<td>0.01.77</td>
<td>34.38.10</td>
<td>1545</td>
<td>245</td>
<td>0.38.75</td>
<td>36.05.76</td>
<td>1000</td>
</tr>
<tr>
<td>219</td>
<td>0.01.58</td>
<td>34.36.96</td>
<td>1580</td>
<td>246</td>
<td>0.41.87</td>
<td>36.05.36</td>
<td>1040</td>
</tr>
<tr>
<td>220</td>
<td>0.01.30</td>
<td>34.35.66</td>
<td>1520</td>
<td>247</td>
<td>0.41.12</td>
<td>35.30.52</td>
<td>2340</td>
</tr>
<tr>
<td>221</td>
<td>0.01.77</td>
<td>34.34.81</td>
<td>1515</td>
<td>248</td>
<td>0.37.10</td>
<td>35.49.34</td>
<td>2180</td>
</tr>
<tr>
<td>222</td>
<td>0.02.15</td>
<td>34.33.62</td>
<td>1500</td>
<td>249</td>
<td>0.43.65</td>
<td>35.50.30</td>
<td>2200</td>
</tr>
<tr>
<td>223</td>
<td>0.00.68</td>
<td>34.32.66</td>
<td>1420</td>
<td>250</td>
<td>0.25.33</td>
<td>35.57.68</td>
<td>1158</td>
</tr>
</tbody>
</table>

**NB:** Latitudes are in degrees, minutes and decimals. Heights refer to maps produced by the Survey of Kenya.
Appendix A

C ******************************************************************** C
C PROGRAM TOPO
C
C THIS IS A PROGRAM FOR DETERMINING THE TOPOGRAPHY ALONG A SEISMIC
C LINE WHOSE STATION LOCATION IS EXTRACTED FROM THE OUTPUT OF TRAN4
C PROGRAM; THE DISTANCES ALONG THE LINE ARE THE AVERAGE OF THE
C DISTANCES BETWEEN EACH STATIONS AND EVERY SHOT POINT.
C
C LANGUAGE : FORTRAN 77
C
C AUTHOR : A. PARKER, R. MASOTTI
C
C DATE : 25/2/1992
C
C ******************************************************************** C
C
PROGRAM TOPO

IMPLICIT DOUBLE PRECISION (A-H,O-Z)

C array dimension of input and output
DIMENSION ISNT(250),ISTNAM(250,5),DIST(250,5),INOST(250,5),
  DIS(250),SH(250,2,5)

PARAMETER (MAX=250)

CHARACTER*20 FILl,FIL2,FIL3,FIL4,FIL5
CHARACTER*20 FILOUT

C shot points distances along the line
SHFl=236.000
SHF2=000.000
SHF3=155.000
SHF4=264.000
SHF5=351.000

C input and output files
WRITE (6,'(A20)') 'enter topo output filename'
READ (5,'(A20)') FILOUT
OPEN (7,FILE='skarl21.out',STATUS='OLD')
OPEN (8,FILE='skarl22.out',STATUS='OLD')
OPEN (9,FILE='skarl23.out',STATUS='OLD')
OPEN (10,FILE='skarl25.out',STATUS='OLD')
OPEN (11,FILE='skarl27.out',STATUS='OLD')
OPEN (12,FILE=FILOUT,STATUS='NEW')

C reading the number of station in each file
READ (7,*),'enter number of stations in each file'
READ (8,*),N2
READ (9,*),N3
READ (10,*),N4
READ (11,*),N5

C loops for the creation of the initial distances array
DO 5 I=1,N1
  READ (7,990) ISNT(I),ISTNAM(I,1),DIST(I,1)
  DO 5 J=1,MAX
    DIS(I,J)=DIST(I,J)
5 CONTINUE

990 FORMAT (A20,5E14.7)
5 CONTINUE

J = 1
DO 10 I = 1,MAX
   IF (ISTNAM(I,1) .EQ. J) THEN
      SH(I,1,1) = DIST(I,1)
      SH(I,2,1) = SHF1
      J = J + 1
   ENDIF
10 CONTINUE

DO 15 I = 1,N2
   READ (8,990) ISNT(I),ISTNAM(I,2),DIST(I,2)
15 CONTINUE

J = 1
DO 20 I = 1,MAX
   IF (ISTNAM(I,2) .EQ. J) THEN
      SH(I,1,2) = DIST(I,2)
      SH(I,2,2) = SHF2
      J = J + 1
   ENDIF
20 CONTINUE

DO 25 I = 1,N3
   READ (9,990) ISNT(I),ISTNAM(I,3),DIST(I,3)
25 CONTINUE

1 = 1
DO 30 I = UMAX
   IF (ISTNAM(I,3) .EQ. J) THEN
      SH(I,1,3) = DIST(I,3)
      SH(I,2,3) = SHF3
      J = J + 1
   ENDIF
30 CONTINUE

DO 35 I = 1,N4
   READ (10,990) ISNT(I),ISTNAM(I,4),DIST(I,4)
35 CONTINUE

J = 1
DO 40 I = UMAX
   IF (ISTNAM(I,4) .EQ. J) THEN
      SH(I,1,4) = DIST(I,4)
      SH(I,2,4) = SHF4
      J = J + 1
   ENDIF
40 CONTINUE

DO 45 I = 1,N5
   READ (11,990) ISNT(I),ISTNAM(I,5),DIST(I,5)
45 CONTINUE

J = 1
DO 50 I = UMAX
   IF (ISTNAM(I,5) .EQ. J) THEN
      SH(I,1,5) = DIST(I,5)
SH(2,5) = SHF5
J = J + 1
ENDIF
50 CONTINUE

DO 60 I = 1,MAX
NUM = 5
DO 65 J = 1,5
IF (SH(I,J) .EQ. 0.0) NUM = NUM - 1
65 CONTINUE
IF (NUM .NE. 0) THEN
DIS(I) = (SH(I,1,1) + SH(I,1,2) + SH(I,1,3) + SH(I,1,4) + SH(I,1,5) +
          SH(I,2,1) + SH(I,2,2) + SH(I,2,3) + SH(I,2,4) + SH(I,2,5))/NUM
ENDIF
60 CONTINUE

WRITE (12,991) (I,DIS(I))

WRITE (12,992) (SHFl)
WRITE (12,992) (SHF2)
WRITE (12,992) (SHF3)
WRITE (12,992) (SHF4)
WRITE (12,992) (SHF5)

990 FORMAT(2I6,F10.3)
991 FORMAT(16,F10.3)
992 FORMAT(2X,F7.3)

STOP
END
Appendix A

**Topography along KRISP 90 Line D**

<table>
<thead>
<tr>
<th>SITE</th>
<th>DIST km</th>
<th>HEIGHT m</th>
<th>SITE</th>
<th>DIST km</th>
<th>HEIGHT m</th>
</tr>
</thead>
<tbody>
<tr>
<td>2</td>
<td>453.555</td>
<td>710</td>
<td>58</td>
<td>282.776</td>
<td>1840</td>
</tr>
<tr>
<td>3</td>
<td>449.395</td>
<td>690</td>
<td>59</td>
<td>281.394</td>
<td>1850</td>
</tr>
<tr>
<td>5</td>
<td>441.401</td>
<td>730</td>
<td>60</td>
<td>280.386</td>
<td>1860</td>
</tr>
<tr>
<td>6</td>
<td>436.297</td>
<td>780</td>
<td>61</td>
<td>279.530</td>
<td>1860</td>
</tr>
<tr>
<td>7</td>
<td>431.195</td>
<td>790</td>
<td>62</td>
<td>278.936</td>
<td>1870</td>
</tr>
<tr>
<td>8</td>
<td>424.400</td>
<td>820</td>
<td>63</td>
<td>278.104</td>
<td>1800</td>
</tr>
<tr>
<td>9</td>
<td>422.055</td>
<td>910</td>
<td>64</td>
<td>276.590</td>
<td>1760</td>
</tr>
<tr>
<td>11</td>
<td>413.113</td>
<td>840</td>
<td>66</td>
<td>275.198</td>
<td>1710</td>
</tr>
<tr>
<td>12</td>
<td>408.272</td>
<td>880</td>
<td>67</td>
<td>274.131</td>
<td>1700</td>
</tr>
<tr>
<td>13</td>
<td>404.081</td>
<td>910</td>
<td>68</td>
<td>273.368</td>
<td>1670</td>
</tr>
<tr>
<td>15</td>
<td>395.837</td>
<td>940</td>
<td>69</td>
<td>272.528</td>
<td>1660</td>
</tr>
<tr>
<td>16</td>
<td>392.262</td>
<td>990</td>
<td>70</td>
<td>271.629</td>
<td>1620</td>
</tr>
<tr>
<td>17</td>
<td>389.275</td>
<td>1120</td>
<td>71</td>
<td>270.783</td>
<td>1600</td>
</tr>
<tr>
<td>18</td>
<td>383.681</td>
<td>1190</td>
<td>73</td>
<td>268.934</td>
<td>1440</td>
</tr>
<tr>
<td>19</td>
<td>378.879</td>
<td>1190</td>
<td>75</td>
<td>265.669</td>
<td>1390</td>
</tr>
<tr>
<td>21</td>
<td>373.863</td>
<td>1260</td>
<td>76</td>
<td>266.940</td>
<td>1330</td>
</tr>
<tr>
<td>22</td>
<td>369.211</td>
<td>1190</td>
<td>79</td>
<td>263.310</td>
<td>1200</td>
</tr>
<tr>
<td>23</td>
<td>365.877</td>
<td>1170</td>
<td>80</td>
<td>263.152</td>
<td>1180</td>
</tr>
<tr>
<td>25</td>
<td>358.549</td>
<td>1180</td>
<td>81</td>
<td>262.286</td>
<td>1170</td>
</tr>
<tr>
<td>26</td>
<td>347.228</td>
<td>1250</td>
<td>82</td>
<td>261.440</td>
<td>1160</td>
</tr>
<tr>
<td>27</td>
<td>350.910</td>
<td>1190</td>
<td>83</td>
<td>260.582</td>
<td>1150</td>
</tr>
<tr>
<td>28</td>
<td>348.666</td>
<td>1220</td>
<td>84</td>
<td>259.418</td>
<td>1150</td>
</tr>
<tr>
<td>30</td>
<td>345.688</td>
<td>1260</td>
<td>85</td>
<td>258.403</td>
<td>1150</td>
</tr>
<tr>
<td>31</td>
<td>343.866</td>
<td>1280</td>
<td>86</td>
<td>257.132</td>
<td>1150</td>
</tr>
<tr>
<td>32</td>
<td>341.216</td>
<td>1310</td>
<td>88</td>
<td>254.264</td>
<td>1070</td>
</tr>
<tr>
<td>33</td>
<td>338.671</td>
<td>1420</td>
<td>89</td>
<td>252.728</td>
<td>1060</td>
</tr>
<tr>
<td>34</td>
<td>338.353</td>
<td>1440</td>
<td>90</td>
<td>251.666</td>
<td>1030</td>
</tr>
<tr>
<td>35</td>
<td>333.425</td>
<td>1540</td>
<td>91</td>
<td>251.519</td>
<td>1060</td>
</tr>
<tr>
<td>36</td>
<td>331.563</td>
<td>1660</td>
<td>92</td>
<td>249.192</td>
<td>1050</td>
</tr>
<tr>
<td>37</td>
<td>329.727</td>
<td>1710</td>
<td>93</td>
<td>248.047</td>
<td>1060</td>
</tr>
<tr>
<td>38</td>
<td>327.575</td>
<td>1720</td>
<td>94</td>
<td>246.868</td>
<td>1080</td>
</tr>
<tr>
<td>39</td>
<td>325.980</td>
<td>1740</td>
<td>95</td>
<td>245.096</td>
<td>1030</td>
</tr>
<tr>
<td>40</td>
<td>324.164</td>
<td>1740</td>
<td>96</td>
<td>244.186</td>
<td>1010</td>
</tr>
<tr>
<td>41</td>
<td>322.298</td>
<td>1760</td>
<td>97</td>
<td>243.374</td>
<td>1020</td>
</tr>
<tr>
<td>42</td>
<td>320.494</td>
<td>1750</td>
<td>98</td>
<td>242.458</td>
<td>1050</td>
</tr>
<tr>
<td>44</td>
<td>315.828</td>
<td>1800</td>
<td>99</td>
<td>241.374</td>
<td>1080</td>
</tr>
<tr>
<td>45</td>
<td>312.638</td>
<td>1800</td>
<td>100</td>
<td>240.246</td>
<td>1010</td>
</tr>
<tr>
<td>46</td>
<td>310.741</td>
<td>1800</td>
<td>101</td>
<td>237.396</td>
<td>1160</td>
</tr>
<tr>
<td>47</td>
<td>308.697</td>
<td>1790</td>
<td>102</td>
<td>236.716</td>
<td>1120</td>
</tr>
<tr>
<td>48</td>
<td>305.805</td>
<td>1800</td>
<td>103</td>
<td>235.940</td>
<td>1060</td>
</tr>
<tr>
<td>49</td>
<td>302.754</td>
<td>1840</td>
<td>104</td>
<td>235.132</td>
<td>1060</td>
</tr>
<tr>
<td>50</td>
<td>300.238</td>
<td>1840</td>
<td>105</td>
<td>234.278</td>
<td>990</td>
</tr>
<tr>
<td>51</td>
<td>297.408</td>
<td>1830</td>
<td>107</td>
<td>230.444</td>
<td>980</td>
</tr>
<tr>
<td>52</td>
<td>294.525</td>
<td>1850</td>
<td>108</td>
<td>229.052</td>
<td>980</td>
</tr>
<tr>
<td>53</td>
<td>291.810</td>
<td>1870</td>
<td>109</td>
<td>227.942</td>
<td>980</td>
</tr>
<tr>
<td>54</td>
<td>290.251</td>
<td>1910</td>
<td>111</td>
<td>225.174</td>
<td>1000</td>
</tr>
<tr>
<td>55</td>
<td>289.510</td>
<td>1970</td>
<td>112</td>
<td>224.394</td>
<td>1020</td>
</tr>
<tr>
<td>56</td>
<td>286.532</td>
<td>1950</td>
<td>113</td>
<td>223.234</td>
<td>1020</td>
</tr>
<tr>
<td>57</td>
<td>284.488</td>
<td>1840</td>
<td>114</td>
<td>222.044</td>
<td>1030</td>
</tr>
</tbody>
</table>
### Appendix A

Topography along KRISP 90 Line D

<table>
<thead>
<tr>
<th>SITE</th>
<th>DIST km</th>
<th>HEIGHT m</th>
<th>SITE</th>
<th>DIST km</th>
<th>HEIGHT m</th>
</tr>
</thead>
<tbody>
<tr>
<td>115</td>
<td>219.606</td>
<td>1090</td>
<td>169</td>
<td>158.892</td>
<td>2290</td>
</tr>
<tr>
<td>116</td>
<td>218.756</td>
<td>1100</td>
<td>170</td>
<td>157.660</td>
<td>2270</td>
</tr>
<tr>
<td>117</td>
<td>216.106</td>
<td>1130</td>
<td>171</td>
<td>155.478</td>
<td>2250</td>
</tr>
<tr>
<td>118</td>
<td>215.438</td>
<td>1140</td>
<td>172</td>
<td>152.760</td>
<td>2210</td>
</tr>
<tr>
<td>119</td>
<td>214.650</td>
<td>1160</td>
<td>173</td>
<td>151.530</td>
<td>2220</td>
</tr>
<tr>
<td>120</td>
<td>213.600</td>
<td>1200</td>
<td>174</td>
<td>148.290</td>
<td>2200</td>
</tr>
<tr>
<td>121</td>
<td>213.896</td>
<td>1190</td>
<td>175</td>
<td>147.012</td>
<td>2200</td>
</tr>
<tr>
<td>122</td>
<td>212.888</td>
<td>1250</td>
<td>176</td>
<td>138.553</td>
<td>2110</td>
</tr>
<tr>
<td>123</td>
<td>212.400</td>
<td>1230</td>
<td>177</td>
<td>135.961</td>
<td>2200</td>
</tr>
<tr>
<td>124</td>
<td>211.238</td>
<td>1260</td>
<td>178</td>
<td>132.093</td>
<td>2170</td>
</tr>
<tr>
<td>125</td>
<td>210.218</td>
<td>1280</td>
<td>179</td>
<td>124.685</td>
<td>2130</td>
</tr>
<tr>
<td>126</td>
<td>208.611</td>
<td>1310</td>
<td>180</td>
<td>123.553</td>
<td>2110</td>
</tr>
<tr>
<td>127</td>
<td>208.168</td>
<td>1320</td>
<td>181</td>
<td>125.963</td>
<td>2090</td>
</tr>
<tr>
<td>128</td>
<td>207.160</td>
<td>1300</td>
<td>182</td>
<td>134.618</td>
<td>2120</td>
</tr>
<tr>
<td>129</td>
<td>206.146</td>
<td>1400</td>
<td>183</td>
<td>133.066</td>
<td>2080</td>
</tr>
<tr>
<td>130</td>
<td>205.970</td>
<td>1540</td>
<td>184</td>
<td>131.336</td>
<td>2030</td>
</tr>
<tr>
<td>131</td>
<td>205.396</td>
<td>1580</td>
<td>185</td>
<td>129.954</td>
<td>2000</td>
</tr>
<tr>
<td>132</td>
<td>204.470</td>
<td>1720</td>
<td>186</td>
<td>128.420</td>
<td>1980</td>
</tr>
<tr>
<td>133</td>
<td>203.108</td>
<td>1680</td>
<td>187</td>
<td>126.564</td>
<td>1970</td>
</tr>
<tr>
<td>134</td>
<td>201.730</td>
<td>1920</td>
<td>188</td>
<td>125.388</td>
<td>1990</td>
</tr>
<tr>
<td>135</td>
<td>200.678</td>
<td>2000</td>
<td>191</td>
<td>122.149</td>
<td>2010</td>
</tr>
<tr>
<td>136</td>
<td>199.940</td>
<td>1920</td>
<td>192</td>
<td>118.560</td>
<td>1980</td>
</tr>
<tr>
<td>137</td>
<td>198.840</td>
<td>1860</td>
<td>193</td>
<td>116.861</td>
<td>1980</td>
</tr>
<tr>
<td>139</td>
<td>197.166</td>
<td>2000</td>
<td>195</td>
<td>112.717</td>
<td>1890</td>
</tr>
<tr>
<td>140</td>
<td>196.826</td>
<td>2060</td>
<td>197</td>
<td>108.735</td>
<td>1860</td>
</tr>
<tr>
<td>141</td>
<td>195.928</td>
<td>1970</td>
<td>198</td>
<td>106.476</td>
<td>1870</td>
</tr>
<tr>
<td>142</td>
<td>195.480</td>
<td>2080</td>
<td>199</td>
<td>104.663</td>
<td>1780</td>
</tr>
<tr>
<td>143</td>
<td>194.014</td>
<td>1820</td>
<td>200</td>
<td>103.202</td>
<td>1740</td>
</tr>
<tr>
<td>144</td>
<td>193.012</td>
<td>1720</td>
<td>202</td>
<td>99.307</td>
<td>1700</td>
</tr>
<tr>
<td>145</td>
<td>191.752</td>
<td>1660</td>
<td>203</td>
<td>98.663</td>
<td>1720</td>
</tr>
<tr>
<td>146</td>
<td>190.366</td>
<td>1560</td>
<td>204</td>
<td>94.108</td>
<td>1660</td>
</tr>
<tr>
<td>147</td>
<td>186.896</td>
<td>1420</td>
<td>205</td>
<td>77.416</td>
<td>1710</td>
</tr>
<tr>
<td>148</td>
<td>187.878</td>
<td>1360</td>
<td>206</td>
<td>88.546</td>
<td>1660</td>
</tr>
<tr>
<td>150</td>
<td>186.128</td>
<td>1220</td>
<td>207</td>
<td>86.694</td>
<td>1700</td>
</tr>
<tr>
<td>151</td>
<td>184.910</td>
<td>1150</td>
<td>208</td>
<td>83.750</td>
<td>1700</td>
</tr>
<tr>
<td>152</td>
<td>183.906</td>
<td>1140</td>
<td>209</td>
<td>81.980</td>
<td>1720</td>
</tr>
<tr>
<td>153</td>
<td>183.140</td>
<td>1150</td>
<td>210</td>
<td>79.320</td>
<td>1700</td>
</tr>
<tr>
<td>155</td>
<td>180.827</td>
<td>1260</td>
<td>211</td>
<td>77.550</td>
<td>1660</td>
</tr>
<tr>
<td>156</td>
<td>176.452</td>
<td>2270</td>
<td>212</td>
<td>74.746</td>
<td>1660</td>
</tr>
<tr>
<td>157</td>
<td>175.010</td>
<td>2740</td>
<td>213</td>
<td>72.090</td>
<td>1640</td>
</tr>
<tr>
<td>158</td>
<td>174.150</td>
<td>2890</td>
<td>214</td>
<td>70.526</td>
<td>1670</td>
</tr>
<tr>
<td>159</td>
<td>173.448</td>
<td>2840</td>
<td>215</td>
<td>69.369</td>
<td>1610</td>
</tr>
<tr>
<td>160</td>
<td>170.584</td>
<td>2600</td>
<td>216</td>
<td>67.456</td>
<td>1600</td>
</tr>
<tr>
<td>162</td>
<td>169.612</td>
<td>2540</td>
<td>217</td>
<td>65.242</td>
<td>1570</td>
</tr>
<tr>
<td>163</td>
<td>168.662</td>
<td>2540</td>
<td>218</td>
<td>63.136</td>
<td>1540</td>
</tr>
<tr>
<td>164</td>
<td>167.816</td>
<td>2510</td>
<td>219</td>
<td>61.056</td>
<td>1580</td>
</tr>
<tr>
<td>167</td>
<td>163.648</td>
<td>2410</td>
<td>221</td>
<td>57.514</td>
<td>1510</td>
</tr>
<tr>
<td>168</td>
<td>160.374</td>
<td>2300</td>
<td>224</td>
<td>51.253</td>
<td>1440</td>
</tr>
</tbody>
</table>
## Appendix A

### Topography along KRISP 90 Line D

<table>
<thead>
<tr>
<th>SITE</th>
<th>DIST km</th>
<th>HEIGHT m</th>
</tr>
</thead>
<tbody>
<tr>
<td>225</td>
<td>49.088</td>
<td>1420</td>
</tr>
<tr>
<td>226</td>
<td>44.470</td>
<td>1340</td>
</tr>
<tr>
<td>227</td>
<td>39.660</td>
<td>1370</td>
</tr>
<tr>
<td>228</td>
<td>35.132</td>
<td>1340</td>
</tr>
<tr>
<td>229</td>
<td>31.146</td>
<td>1320</td>
</tr>
<tr>
<td>230</td>
<td>25.778</td>
<td>1340</td>
</tr>
<tr>
<td>231</td>
<td>23.272</td>
<td>1280</td>
</tr>
<tr>
<td>232</td>
<td>14.374</td>
<td>1190</td>
</tr>
<tr>
<td>233</td>
<td>9.847</td>
<td>1180</td>
</tr>
<tr>
<td>234</td>
<td>5.853</td>
<td>1190</td>
</tr>
<tr>
<td>235</td>
<td>4.299</td>
<td>1140</td>
</tr>
<tr>
<td>237</td>
<td>176.020</td>
<td>2720</td>
</tr>
<tr>
<td>238</td>
<td>180.634</td>
<td>1240</td>
</tr>
<tr>
<td>239</td>
<td>187.430</td>
<td>1340</td>
</tr>
<tr>
<td>240</td>
<td>218.766</td>
<td>1100</td>
</tr>
<tr>
<td>241</td>
<td>231.602</td>
<td>1000</td>
</tr>
<tr>
<td>242</td>
<td>232.158</td>
<td>980</td>
</tr>
<tr>
<td>243</td>
<td>234.082</td>
<td>980</td>
</tr>
<tr>
<td>244</td>
<td>235.790</td>
<td>980</td>
</tr>
<tr>
<td>245</td>
<td>238.376</td>
<td>1000</td>
</tr>
<tr>
<td>246</td>
<td>237.668</td>
<td>1040</td>
</tr>
<tr>
<td>247</td>
<td>180.490</td>
<td>2340</td>
</tr>
<tr>
<td>248</td>
<td>209.188</td>
<td>2190</td>
</tr>
<tr>
<td>249</td>
<td>213.974</td>
<td>2200</td>
</tr>
<tr>
<td>250</td>
<td>213.948</td>
<td>1160</td>
</tr>
</tbody>
</table>

**NB:** Heights are rounded to the nearest 10 m; Not all recording sites are represented here.
### Appendix B

#### Time-term Depths along Line D

**VICVOL. DEP** (from VIC to c.130km)

<table>
<thead>
<tr>
<th>STATION</th>
<th>Z(TIMT)</th>
<th>Z(ERR)</th>
<th>Z(ST ERR)</th>
</tr>
</thead>
<tbody>
<tr>
<td>229</td>
<td>0.07748</td>
<td>0.11180</td>
<td>0.12910</td>
</tr>
<tr>
<td>228</td>
<td>0.13316</td>
<td>0.02083</td>
<td>0.02405</td>
</tr>
<tr>
<td>227</td>
<td>0.16471</td>
<td>0.00451</td>
<td>0.00521</td>
</tr>
<tr>
<td>221</td>
<td>0.08003</td>
<td>0.00263</td>
<td>0.00303</td>
</tr>
<tr>
<td>219</td>
<td>0.15391</td>
<td>0.02221</td>
<td>0.02566</td>
</tr>
<tr>
<td>218</td>
<td>0.15920</td>
<td>0.00150</td>
<td>0.00174</td>
</tr>
<tr>
<td>217</td>
<td>0.15074</td>
<td>0.02142</td>
<td>0.02473</td>
</tr>
<tr>
<td>216</td>
<td>0.13592</td>
<td>0.1395</td>
<td>0.1611</td>
</tr>
<tr>
<td>215</td>
<td>0.13592</td>
<td>0.00622</td>
<td>0.00718</td>
</tr>
<tr>
<td>214</td>
<td>0.13002</td>
<td>0.00534</td>
<td>0.00616</td>
</tr>
<tr>
<td>213</td>
<td>0.17614</td>
<td>0.04912</td>
<td>0.05672</td>
</tr>
<tr>
<td>212</td>
<td>0.17191</td>
<td>0.01554</td>
<td>0.01793</td>
</tr>
<tr>
<td>210</td>
<td>0.23542</td>
<td>0.01310</td>
<td>0.01514</td>
</tr>
<tr>
<td>209</td>
<td>0.26823</td>
<td>0.00667</td>
<td>0.00768</td>
</tr>
<tr>
<td>208</td>
<td>0.19519</td>
<td>0.00487</td>
<td>0.00581</td>
</tr>
<tr>
<td>207</td>
<td>0.14650</td>
<td>0.01315</td>
<td>0.01520</td>
</tr>
<tr>
<td>204</td>
<td>0.21255</td>
<td>0.01558</td>
<td>0.01914</td>
</tr>
<tr>
<td>203</td>
<td>0.23902</td>
<td>0.02943</td>
<td>0.03398</td>
</tr>
<tr>
<td>200</td>
<td>0.17191</td>
<td>0.02943</td>
<td>0.03398</td>
</tr>
<tr>
<td>189</td>
<td>0.09929</td>
<td>0.00301</td>
<td>0.00347</td>
</tr>
<tr>
<td>188</td>
<td>0.18059</td>
<td>0.08024</td>
<td>0.09264</td>
</tr>
<tr>
<td>187</td>
<td>0.19858</td>
<td>0.07513</td>
<td>0.08676</td>
</tr>
<tr>
<td>186</td>
<td>0.10734</td>
<td>0.04082</td>
<td>0.04715</td>
</tr>
<tr>
<td>184</td>
<td>0.14333</td>
<td>0.03468</td>
<td>0.04003</td>
</tr>
<tr>
<td>193</td>
<td>0.14333</td>
<td>0.03468</td>
<td>0.04003</td>
</tr>
<tr>
<td>192</td>
<td>0.14544</td>
<td>0.07414</td>
<td>0.08559</td>
</tr>
<tr>
<td>191</td>
<td>0.20176</td>
<td>0.01099</td>
<td>0.01266</td>
</tr>
<tr>
<td>188</td>
<td>0.30062</td>
<td>0.03089</td>
<td>0.03567</td>
</tr>
<tr>
<td>187</td>
<td>0.02893</td>
<td>0.01806</td>
<td>0.02085</td>
</tr>
<tr>
<td>186</td>
<td>0.22948</td>
<td>0.01609</td>
<td>0.01916</td>
</tr>
<tr>
<td>185</td>
<td>0.19741</td>
<td>0.02887</td>
<td>0.03102</td>
</tr>
<tr>
<td>999</td>
<td>0.57944</td>
<td>0.00779</td>
<td>0.00779</td>
</tr>
<tr>
<td>998</td>
<td>0.40203</td>
<td>0.00779</td>
<td>0.00779</td>
</tr>
</tbody>
</table>

**KAPELG. DEP** (from KAP to the Elgeyo escarpment)

<table>
<thead>
<tr>
<th>STATION</th>
<th>Z(TIMT)</th>
<th>Z(ERR)</th>
<th>Z(ST ERR)</th>
</tr>
</thead>
<tbody>
<tr>
<td>189</td>
<td>0.55830</td>
<td>0.40722</td>
<td>0.43191</td>
</tr>
<tr>
<td>179</td>
<td>0.56706</td>
<td>0.33864</td>
<td>0.35911</td>
</tr>
<tr>
<td>177</td>
<td>0.86652</td>
<td>0.31037</td>
<td>0.32917</td>
</tr>
<tr>
<td>176</td>
<td>0.94159</td>
<td>0.23120</td>
<td>0.24522</td>
</tr>
<tr>
<td>175</td>
<td>1.21536</td>
<td>0.40308</td>
<td>0.42753</td>
</tr>
<tr>
<td>174</td>
<td>0.85298</td>
<td>0.20767</td>
<td>0.22045</td>
</tr>
<tr>
<td>168</td>
<td>1.56340</td>
<td>0.04030</td>
<td>0.05685</td>
</tr>
</tbody>
</table>

*VELOCITIES USED (km/s): 2.0000  6.0989*

*VELOCITIES USED (km/s): 4.8000  6.0152*
### Appendix B

**Time-term Depths along Line D**

<table>
<thead>
<tr>
<th>Station</th>
<th>Z(TIMT)</th>
<th>Z(Err)</th>
<th>Z(Err)</th>
</tr>
</thead>
<tbody>
<tr>
<td>167</td>
<td>1.59128</td>
<td>0.09494</td>
<td>0.10967</td>
</tr>
<tr>
<td>164</td>
<td>1.75056</td>
<td>0.22587</td>
<td>0.28083</td>
</tr>
<tr>
<td>163</td>
<td>2.65053</td>
<td>0.83211</td>
<td>0.96082</td>
</tr>
<tr>
<td>162</td>
<td>1.96241</td>
<td>0.30894</td>
<td>0.35672</td>
</tr>
<tr>
<td>161</td>
<td>2.01418</td>
<td>0.31507</td>
<td>0.36373</td>
</tr>
<tr>
<td>159</td>
<td>2.27302</td>
<td>0.51943</td>
<td>0.59979</td>
</tr>
<tr>
<td>156</td>
<td>2.34391</td>
<td>0.90650</td>
<td>1.04675</td>
</tr>
<tr>
<td>238</td>
<td>2.36382</td>
<td>1.19645</td>
<td>1.28609</td>
</tr>
<tr>
<td>999</td>
<td>-0.22300</td>
<td>0.12289</td>
<td>0.12416</td>
</tr>
<tr>
<td>998</td>
<td>0.58777</td>
<td>0.09653</td>
<td>0.09653</td>
</tr>
<tr>
<td>997</td>
<td>6.02742</td>
<td>0.90960</td>
<td>0.96777</td>
</tr>
</tbody>
</table>

**ELGBAR.DEP (from the Elgeyo escarpment to BAR)**

**VELOCITIES USED (km/s):** 3.8000 6.0000

<table>
<thead>
<tr>
<th>Station</th>
<th>Z(TIMT)</th>
<th>Z(Err)</th>
<th>Z(Err)</th>
</tr>
</thead>
<tbody>
<tr>
<td>153</td>
<td>2.81264</td>
<td>0.44316</td>
<td>0.51171</td>
</tr>
<tr>
<td>152</td>
<td>2.75617</td>
<td>0.44031</td>
<td>0.50842</td>
</tr>
<tr>
<td>151</td>
<td>3.15390</td>
<td>0.25828</td>
<td>0.29620</td>
</tr>
<tr>
<td>150</td>
<td>3.38223</td>
<td>0.03599</td>
<td>0.04519</td>
</tr>
<tr>
<td>148</td>
<td>3.41170</td>
<td>0.11932</td>
<td>0.13778</td>
</tr>
<tr>
<td>147</td>
<td>3.43870</td>
<td>0.12256</td>
<td>0.14152</td>
</tr>
<tr>
<td>146</td>
<td>3.47799</td>
<td>0.26707</td>
<td>0.30842</td>
</tr>
<tr>
<td>145</td>
<td>3.33088</td>
<td>0.16494</td>
<td>0.19047</td>
</tr>
<tr>
<td>144</td>
<td>3.50499</td>
<td>0.17550</td>
<td>0.20265</td>
</tr>
<tr>
<td>143</td>
<td>3.40924</td>
<td>0.24468</td>
<td>0.28254</td>
</tr>
<tr>
<td>141</td>
<td>3.77752</td>
<td>0.31343</td>
<td>0.36194</td>
</tr>
<tr>
<td>140</td>
<td>3.78979</td>
<td>0.22971</td>
<td>0.26521</td>
</tr>
<tr>
<td>137</td>
<td>3.76328</td>
<td>0.36626</td>
<td>0.42293</td>
</tr>
<tr>
<td>136</td>
<td>3.82956</td>
<td>0.18561</td>
<td>0.21434</td>
</tr>
<tr>
<td>135</td>
<td>3.82824</td>
<td>0.09888</td>
<td>0.11534</td>
</tr>
<tr>
<td>134</td>
<td>3.59632</td>
<td>0.00795</td>
<td>0.00918</td>
</tr>
<tr>
<td>133</td>
<td>3.50548</td>
<td>0.09791</td>
<td>0.11308</td>
</tr>
<tr>
<td>132</td>
<td>3.57423</td>
<td>0.08549</td>
<td>0.09870</td>
</tr>
<tr>
<td>131</td>
<td>3.45147</td>
<td>0.20913</td>
<td>0.24149</td>
</tr>
<tr>
<td>124</td>
<td>3.81695</td>
<td>0.00795</td>
<td>0.00918</td>
</tr>
<tr>
<td>123</td>
<td>3.72252</td>
<td>0.05102</td>
<td>0.05892</td>
</tr>
<tr>
<td>122</td>
<td>3.88324</td>
<td>0.02396</td>
<td>0.02765</td>
</tr>
<tr>
<td>119</td>
<td>3.72448</td>
<td>0.02229</td>
<td>0.02578</td>
</tr>
<tr>
<td>117</td>
<td>4.15512</td>
<td>0.17584</td>
<td>0.20304</td>
</tr>
<tr>
<td>999</td>
<td>-0.13749</td>
<td>0.07577</td>
<td>0.07655</td>
</tr>
<tr>
<td>998</td>
<td>0.36236</td>
<td>0.05951</td>
<td>0.05951</td>
</tr>
<tr>
<td>997</td>
<td>3.71614</td>
<td>0.05961</td>
<td>0.05966</td>
</tr>
</tbody>
</table>
### Appendix B

**Time-term Depths along Line D**

**BARTAN.DEP (from BAR to TAN)**

<table>
<thead>
<tr>
<th>VELOCITIES USED (km/s):</th>
<th>2.4000</th>
<th>3.9000</th>
<th>6.0777</th>
</tr>
</thead>
</table>

<table>
<thead>
<tr>
<th>STATION</th>
<th>Z(TIMT)</th>
<th>Z(ERR)</th>
<th>Z(ST ERR)</th>
</tr>
</thead>
<tbody>
<tr>
<td>119</td>
<td>3.91298</td>
<td>0.08877</td>
<td>0.09426</td>
</tr>
<tr>
<td>117</td>
<td>4.24656</td>
<td>0.14140</td>
<td>0.15009</td>
</tr>
<tr>
<td>240</td>
<td>5.03270</td>
<td>0.01341</td>
<td>0.01580</td>
</tr>
<tr>
<td>115</td>
<td>4.93557</td>
<td>0.08602</td>
<td>0.09965</td>
</tr>
<tr>
<td>114</td>
<td>4.99964</td>
<td>0.13698</td>
<td>0.15848</td>
</tr>
<tr>
<td>113</td>
<td>5.01134</td>
<td>0.11648</td>
<td>0.13484</td>
</tr>
<tr>
<td>112</td>
<td>4.88845</td>
<td>0.06919</td>
<td>0.08029</td>
</tr>
<tr>
<td>111</td>
<td>4.55725</td>
<td>0.12040</td>
<td>0.13931</td>
</tr>
<tr>
<td>109</td>
<td>3.42981</td>
<td>-0.90356</td>
<td>-0.89278</td>
</tr>
<tr>
<td>108</td>
<td>3.20404</td>
<td>-0.93107</td>
<td>-0.92451</td>
</tr>
<tr>
<td>107</td>
<td>2.74741</td>
<td>-0.59765</td>
<td>-0.57487</td>
</tr>
<tr>
<td>241</td>
<td>2.51146</td>
<td>-0.56333</td>
<td>-0.49986</td>
</tr>
<tr>
<td>99</td>
<td>3.43143</td>
<td>0.15091</td>
<td>0.17455</td>
</tr>
<tr>
<td>98</td>
<td>3.22193</td>
<td>0.03126</td>
<td>0.03639</td>
</tr>
<tr>
<td>97</td>
<td>3.27889</td>
<td>0.04138</td>
<td>0.04809</td>
</tr>
<tr>
<td>96</td>
<td>3.36836</td>
<td>0.03314</td>
<td>0.03858</td>
</tr>
<tr>
<td>95</td>
<td>3.37347</td>
<td>0.02246</td>
<td>0.02626</td>
</tr>
<tr>
<td>94</td>
<td>3.17383</td>
<td>0.10387</td>
<td>0.12025</td>
</tr>
<tr>
<td>92</td>
<td>2.96141</td>
<td>0.14694</td>
<td>0.18993</td>
</tr>
<tr>
<td>88</td>
<td>3.51483</td>
<td>0.01514</td>
<td>0.01778</td>
</tr>
<tr>
<td>86</td>
<td>3.19428</td>
<td>-0.72284</td>
<td>-0.71419</td>
</tr>
<tr>
<td>85</td>
<td>2.86680</td>
<td>-0.65979</td>
<td>-0.64138</td>
</tr>
<tr>
<td>83</td>
<td>2.87771</td>
<td>-0.80687</td>
<td>-0.80520</td>
</tr>
<tr>
<td>82</td>
<td>2.75713</td>
<td>-0.80099</td>
<td>-0.78634</td>
</tr>
<tr>
<td>81</td>
<td>3.56486</td>
<td>0.08450</td>
<td>0.09787</td>
</tr>
<tr>
<td>79</td>
<td>3.53008</td>
<td>0.07667</td>
<td>0.08882</td>
</tr>
<tr>
<td>75</td>
<td>3.34397</td>
<td>0.05175</td>
<td>0.06009</td>
</tr>
<tr>
<td>73</td>
<td>3.00887</td>
<td>0.05338</td>
<td>0.06192</td>
</tr>
<tr>
<td>70</td>
<td>2.79835</td>
<td>0.00588</td>
<td>0.00705</td>
</tr>
<tr>
<td>69</td>
<td>2.38291</td>
<td>0.34222</td>
<td>0.37053</td>
</tr>
<tr>
<td>68</td>
<td>2.39613</td>
<td>0.32344</td>
<td>0.35263</td>
</tr>
<tr>
<td>67</td>
<td>2.45105</td>
<td>0.27417</td>
<td>0.29090</td>
</tr>
<tr>
<td>66</td>
<td>2.43783</td>
<td>0.11689</td>
<td>0.12411</td>
</tr>
<tr>
<td>63</td>
<td>2.26443</td>
<td>0.11323</td>
<td>0.12019</td>
</tr>
<tr>
<td>62</td>
<td>2.28477</td>
<td>0.00619</td>
<td>0.00746</td>
</tr>
<tr>
<td>61</td>
<td>2.32189</td>
<td>0.08501</td>
<td>0.09848</td>
</tr>
<tr>
<td>60</td>
<td>2.19172</td>
<td>0.06411</td>
<td>0.07433</td>
</tr>
<tr>
<td>59</td>
<td>2.07324</td>
<td>0.04707</td>
<td>0.05607</td>
</tr>
<tr>
<td>58</td>
<td>2.01527</td>
<td>0.01763</td>
<td>0.01880</td>
</tr>
<tr>
<td>57</td>
<td>2.02035</td>
<td>0.04220</td>
<td>0.04911</td>
</tr>
<tr>
<td>55</td>
<td>1.85407</td>
<td>0.02216</td>
<td>0.02587</td>
</tr>
<tr>
<td>54</td>
<td>1.61508</td>
<td>0.13199</td>
<td>0.14008</td>
</tr>
<tr>
<td>51</td>
<td>1.75492</td>
<td>0.08323</td>
<td>0.08841</td>
</tr>
<tr>
<td>50</td>
<td>1.81390</td>
<td>0.15345</td>
<td>0.16291</td>
</tr>
<tr>
<td>998</td>
<td>0.92708</td>
<td>0.07285</td>
<td>0.07306</td>
</tr>
<tr>
<td>997</td>
<td>3.14440</td>
<td>-0.94830</td>
<td>-0.94830</td>
</tr>
</tbody>
</table>
### Appendix B

Time-term Depths along Line D

<table>
<thead>
<tr>
<th>Station</th>
<th>Z(TimT)</th>
<th>Z(ERR)</th>
<th>Z(Err)</th>
</tr>
</thead>
<tbody>
<tr>
<td>996</td>
<td>3.33736</td>
<td>0.04514</td>
<td>0.04514</td>
</tr>
<tr>
<td>995</td>
<td>-0.43671</td>
<td>0.04423</td>
<td>0.04428</td>
</tr>
</tbody>
</table>

**BARBAS.DEP (from BAR to BAS)**

**VELOCITIES USED (km/s): 2.4000 3.9000 6.1224**

<table>
<thead>
<tr>
<th>STATION</th>
<th>Z(TimT)</th>
<th>Z(ERR)</th>
<th>Z(Err)</th>
</tr>
</thead>
<tbody>
<tr>
<td>82</td>
<td>2.80372</td>
<td>-0.84018</td>
<td>-0.83218</td>
</tr>
<tr>
<td>81</td>
<td>3.60736</td>
<td>0.03853</td>
<td>0.04481</td>
</tr>
<tr>
<td>79</td>
<td>3.58941</td>
<td>0.02533</td>
<td>0.02985</td>
</tr>
<tr>
<td>75</td>
<td>3.38272</td>
<td>0.00347</td>
<td>0.00428</td>
</tr>
<tr>
<td>73</td>
<td>3.04578</td>
<td>0.01807</td>
<td>0.01885</td>
</tr>
<tr>
<td>70</td>
<td>2.83329</td>
<td>0.07481</td>
<td>0.08670</td>
</tr>
<tr>
<td>69</td>
<td>2.32837</td>
<td>0.42699</td>
<td>0.45299</td>
</tr>
<tr>
<td>68</td>
<td>2.34304</td>
<td>0.41995</td>
<td>0.44555</td>
</tr>
<tr>
<td>67</td>
<td>2.39920</td>
<td>0.35934</td>
<td>0.38130</td>
</tr>
<tr>
<td>66</td>
<td>2.38807</td>
<td>0.16662</td>
<td>0.16623</td>
</tr>
<tr>
<td>63</td>
<td>2.22111</td>
<td>0.14761</td>
<td>0.15667</td>
</tr>
<tr>
<td>62</td>
<td>2.28739</td>
<td>0.14427</td>
<td>0.16689</td>
</tr>
<tr>
<td>60</td>
<td>2.20442</td>
<td>0.20114</td>
<td>0.23256</td>
</tr>
<tr>
<td>59</td>
<td>2.03695</td>
<td>0.14476</td>
<td>0.15368</td>
</tr>
<tr>
<td>61</td>
<td>2.32786</td>
<td>0.22254</td>
<td>0.25725</td>
</tr>
<tr>
<td>58</td>
<td>1.98231</td>
<td>0.08032</td>
<td>0.09307</td>
</tr>
<tr>
<td>57</td>
<td>1.86038</td>
<td>0.04537</td>
<td>0.05265</td>
</tr>
<tr>
<td>56</td>
<td>1.97523</td>
<td>0.08032</td>
<td>0.09307</td>
</tr>
<tr>
<td>55</td>
<td>1.69596</td>
<td>0.04628</td>
<td>0.05371</td>
</tr>
<tr>
<td>54</td>
<td>1.60084</td>
<td>0.03504</td>
<td>0.03732</td>
</tr>
<tr>
<td>51</td>
<td>1.75313</td>
<td>0.16421</td>
<td>0.17428</td>
</tr>
<tr>
<td>50</td>
<td>1.81687</td>
<td>0.03282</td>
<td>0.03494</td>
</tr>
<tr>
<td>49</td>
<td>1.61855</td>
<td>0.09216</td>
<td>0.09788</td>
</tr>
<tr>
<td>48</td>
<td>1.67876</td>
<td>0.10314</td>
<td>0.10952</td>
</tr>
<tr>
<td>47</td>
<td>1.65346</td>
<td>0.04688</td>
<td>0.04982</td>
</tr>
<tr>
<td>46</td>
<td>1.70456</td>
<td>0.08629</td>
<td>0.09166</td>
</tr>
<tr>
<td>45</td>
<td>1.37368</td>
<td>0.18667</td>
<td>0.19810</td>
</tr>
<tr>
<td>44</td>
<td>1.06503</td>
<td>0.03965</td>
<td>0.04607</td>
</tr>
<tr>
<td>42</td>
<td>0.97653</td>
<td>0.09474</td>
<td>0.10967</td>
</tr>
<tr>
<td>41</td>
<td>1.01042</td>
<td>0.09383</td>
<td>0.09965</td>
</tr>
<tr>
<td>40</td>
<td>0.84903</td>
<td>0.08129</td>
<td>0.08635</td>
</tr>
<tr>
<td>39</td>
<td>0.76252</td>
<td>0.11680</td>
<td>0.12404</td>
</tr>
<tr>
<td>38</td>
<td>0.83436</td>
<td>0.10884</td>
<td>0.11346</td>
</tr>
<tr>
<td>37</td>
<td>0.83199</td>
<td>0.03150</td>
<td>0.03383</td>
</tr>
<tr>
<td>36</td>
<td>0.60518</td>
<td>0.05159</td>
<td>0.05483</td>
</tr>
<tr>
<td>35</td>
<td>0.42759</td>
<td>0.06525</td>
<td>0.07562</td>
</tr>
<tr>
<td>34</td>
<td>0.42102</td>
<td>0.13861</td>
<td>0.14716</td>
</tr>
<tr>
<td>33</td>
<td>0.34310</td>
<td>0.17013</td>
<td>0.18060</td>
</tr>
<tr>
<td>32</td>
<td>0.04815</td>
<td>0.14625</td>
<td>0.15525</td>
</tr>
<tr>
<td>31</td>
<td>0.05928</td>
<td>0.14159</td>
<td>0.15029</td>
</tr>
<tr>
<td>30</td>
<td>-0.05304</td>
<td>0.16618</td>
<td>0.17635</td>
</tr>
<tr>
<td>26</td>
<td>0.08710</td>
<td>0.28087</td>
<td>0.29803</td>
</tr>
</tbody>
</table>
### Appendix B

#### Time-term Depths along Line D

<table>
<thead>
<tr>
<th>Station</th>
<th>Z(TMT)</th>
<th>Z(ERR)</th>
<th>Z(Err)</th>
</tr>
</thead>
<tbody>
<tr>
<td>28</td>
<td>0.03601</td>
<td>0.23003</td>
<td>0.24409</td>
</tr>
<tr>
<td>997</td>
<td>3.17575</td>
<td>-0.94311</td>
<td>-0.94311</td>
</tr>
<tr>
<td>996</td>
<td>3.61292</td>
<td>0.05806</td>
<td>0.05806</td>
</tr>
<tr>
<td>995</td>
<td>0.13517</td>
<td>0.02290</td>
<td>0.02290</td>
</tr>
</tbody>
</table>

VOLCHF.DEP (from c.330 km to CHF)

VELOCITIES USED (km/s): 2.0000 6.0286

<table>
<thead>
<tr>
<th>Station</th>
<th>Z(TMT)</th>
<th>Z(ERR)</th>
<th>Z(Err)</th>
</tr>
</thead>
<tbody>
<tr>
<td>37</td>
<td>0.20098</td>
<td>0.05506</td>
<td>0.06358</td>
</tr>
<tr>
<td>36</td>
<td>0.17787</td>
<td>0.06563</td>
<td>0.07888</td>
</tr>
<tr>
<td>34</td>
<td>0.14925</td>
<td>0.09986</td>
<td>0.01151</td>
</tr>
<tr>
<td>33</td>
<td>0.13780</td>
<td>0.01656</td>
<td>0.01910</td>
</tr>
<tr>
<td>32</td>
<td>-0.00848</td>
<td>0.01821</td>
<td>0.02103</td>
</tr>
<tr>
<td>31</td>
<td>-0.01102</td>
<td>0.02955</td>
<td>0.03413</td>
</tr>
<tr>
<td>30</td>
<td>-0.03943</td>
<td>0.04887</td>
<td>0.05641</td>
</tr>
<tr>
<td>28</td>
<td>0.02968</td>
<td>0.05063</td>
<td>0.05845</td>
</tr>
<tr>
<td>26</td>
<td>0.04961</td>
<td>0.01483</td>
<td>0.01690</td>
</tr>
<tr>
<td>25</td>
<td>0.16621</td>
<td>0.07950</td>
<td>0.09182</td>
</tr>
<tr>
<td>22</td>
<td>0.21794</td>
<td>0.04272</td>
<td>0.04933</td>
</tr>
<tr>
<td>13</td>
<td>0.37852</td>
<td>0.03066</td>
<td>0.03541</td>
</tr>
<tr>
<td>9</td>
<td>0.36253</td>
<td>0.04906</td>
<td>0.05665</td>
</tr>
<tr>
<td>3</td>
<td>0.10134</td>
<td>0.08368</td>
<td>0.09663</td>
</tr>
<tr>
<td>2</td>
<td>0.08459</td>
<td>0.07202</td>
<td>0.08315</td>
</tr>
<tr>
<td>996</td>
<td>1.28094</td>
<td>0.01702</td>
<td>0.01705</td>
</tr>
<tr>
<td>995</td>
<td>-0.04473</td>
<td>0.01003</td>
<td>0.01003</td>
</tr>
<tr>
<td>994</td>
<td>0.02099</td>
<td>0.02580</td>
<td>0.02801</td>
</tr>
</tbody>
</table>
### Appendix C

Inversion Iteration Results for Line D

<table>
<thead>
<tr>
<th>Iteration</th>
<th>Comments</th>
<th>$\chi^2$</th>
<th>No rays</th>
<th>vol. nodes</th>
<th>int. nodes</th>
<th>rms error</th>
<th>D</th>
</tr>
</thead>
<tbody>
<tr>
<td>UCs</td>
<td>phases: a, b</td>
<td></td>
<td>573 picks</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>v016</td>
<td></td>
<td>6.104</td>
<td>460</td>
<td>66</td>
<td>0.175</td>
<td>5</td>
<td></td>
</tr>
<tr>
<td>v017</td>
<td></td>
<td>3.343</td>
<td>474</td>
<td>66</td>
<td>0.121</td>
<td>5</td>
<td></td>
</tr>
<tr>
<td>v018</td>
<td>sol. div.</td>
<td>10.988</td>
<td>506</td>
<td>70</td>
<td>0.213</td>
<td>1</td>
<td></td>
</tr>
<tr>
<td>v019</td>
<td></td>
<td>3.385</td>
<td>478</td>
<td>70</td>
<td>0.125</td>
<td>1</td>
<td></td>
</tr>
<tr>
<td>v020</td>
<td>smoothed</td>
<td>2.805</td>
<td>486</td>
<td>70</td>
<td>0.106</td>
<td></td>
<td></td>
</tr>
<tr>
<td>v021</td>
<td>pref. sol.</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

| MCs       | phases: h, h₁, (b₂) | 223 picks |         |            |            |            |    |
| v038      | smoothed | 1.065   | 194     | 20         | 10         | 0.197      | 1  |
| v039      | pref. sol. | 1.642   | 200     | 20         | 10         | 0.175      |    |

| LCs+UM    | phases: c, d₁, (d₂) | 397 picks |         |            |            |            |    |
| v054      | smoothed | 2.038   | 370     | 50         | 20         | 0.220      | 1  |
| v055      | solution div. | 2.579   | 288     | 50         | 20         | 0.250      | 10 |
| v056      | smoothed | 2.483   | 322     | 50         | 20         | 0.246      | 20 |
| v057      | smoothed | 2.222   | 316     | 50         | 20         | 0.241      | 10 |
| v058      | smoothed | 2.205   | 325     | 50         | 20         | 0.236      | 5  |
| v059      | smoothed | 2.715   | 330     | 50         | 20         | 0.282      | 5  |
| v060      | smoothed | 2.450   | 303     | 50         | 20         | 0.257      | 1  |
| v062      | smoothed | 5.776   | 364     | 30         | 10         | 0.335      | 1  |
| v063      | pref. sol. | 1.957   | 343     | 30         | 10         | 0.225      |    |

**UCs** = upper crust; **MCs** = mid-crust; **LCs** = lower crust; **UM** = upper mantle.

*vel.* = velocity; *int.* = interface; *D* = damping factor; *sol. div.* = diverging solution; *pref. sol.* = preferred solution.
Appendix D

GRAVMAG MODEL LISTINGS

Model Title: KRISP 90 EW Line (Inverted Model - B2)

Profile Orientation: .00000 Degrees

Earth's Magnetic Field Declination Inclination
38.20000 A/m -7.000 Degrees 68.000 Degrees

Gravity Calculation Surface: .000 Km Above Observations

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>Backgnd</td>
<td>2.7000</td>
<td>.0000</td>
<td>.0000 -7.000</td>
<td>68.000</td>
<td>10.00</td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>2.1000</td>
<td>.0000</td>
<td>.0000 -7.000</td>
<td>68.000</td>
<td>10.00</td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>2.1000</td>
<td>.0000</td>
<td>.0000 -7.000</td>
<td>68.000</td>
<td>10.00</td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>2.3000</td>
<td>.0000</td>
<td>.0000 -7.000</td>
<td>68.000</td>
<td>10.00</td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>2.4600</td>
<td>.0000</td>
<td>.0000 -7.000</td>
<td>68.000</td>
<td>100.00</td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>2.5000</td>
<td>.0000</td>
<td>.0000 -7.000</td>
<td>68.000</td>
<td>100.00</td>
<td></td>
</tr>
<tr>
<td>6</td>
<td>2.6000</td>
<td>.0000</td>
<td>.0000 -7.000</td>
<td>68.000</td>
<td>100.00</td>
<td></td>
</tr>
<tr>
<td>7</td>
<td>2.8000</td>
<td>.0000</td>
<td>.0000 -7.000</td>
<td>68.000</td>
<td>100.00</td>
<td></td>
</tr>
<tr>
<td>8</td>
<td>2.8600</td>
<td>.0000</td>
<td>.0000 -7.000</td>
<td>68.000</td>
<td>10000.00</td>
<td></td>
</tr>
<tr>
<td>9</td>
<td>2.9200</td>
<td>.0000</td>
<td>.0000 -7.000</td>
<td>68.000</td>
<td>10000.00</td>
<td></td>
</tr>
<tr>
<td>10</td>
<td>3.2500</td>
<td>.0000</td>
<td>.0000 -7.000</td>
<td>68.000</td>
<td>10000.00</td>
<td></td>
</tr>
<tr>
<td>11</td>
<td>0.9400</td>
<td>.0000</td>
<td>.0000 -7.000</td>
<td>68.000</td>
<td>10000.00</td>
<td></td>
</tr>
<tr>
<td>12</td>
<td>3.2000</td>
<td>.0000</td>
<td>.0000 -7.000</td>
<td>68.000</td>
<td>10000.00</td>
<td></td>
</tr>
<tr>
<td>13</td>
<td>2.8200</td>
<td>.0000</td>
<td>.0000 0.000</td>
<td>0.000</td>
<td>0.000</td>
<td></td>
</tr>
<tr>
<td>14</td>
<td>2.5500</td>
<td>.0000</td>
<td>.0000 -7.000</td>
<td>68.000</td>
<td>100.00</td>
<td></td>
</tr>
<tr>
<td>15</td>
<td>2.7500</td>
<td>.0000</td>
<td>.0000 -7.000</td>
<td>68.000</td>
<td>100.00</td>
<td></td>
</tr>
<tr>
<td>16</td>
<td>2.7500</td>
<td>.0000</td>
<td>.0000 -7.000</td>
<td>68.000</td>
<td>100.00</td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>1 Outline</td>
<td>180.600</td>
<td>-0.7700</td>
<td>180.3573 -1.2735</td>
<td>181.5600 -1.2100</td>
<td></td>
<td></td>
</tr>
<tr>
<td>2 Outline</td>
<td>182.620</td>
<td>-1.1620</td>
<td>184.6100 -1.1360</td>
<td>185.6835 -1.1945</td>
<td></td>
<td></td>
</tr>
<tr>
<td>3 Outline</td>
<td>184.500</td>
<td>-0.8000</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>1 Outline</td>
<td>214.1000</td>
<td>-0.3000</td>
<td>213.3701 -1.2123</td>
<td>214.3100 -1.1960</td>
<td></td>
<td></td>
</tr>
<tr>
<td>2 Outline</td>
<td>215.8000</td>
<td>-1.1420</td>
<td>220.1400 -1.0490</td>
<td>222.8200 -1.0230</td>
<td></td>
<td></td>
</tr>
<tr>
<td>3 Outline</td>
<td>227.6500</td>
<td>-1.0050</td>
<td>231.4400 -0.9740</td>
<td>239.2700 -0.9670</td>
<td></td>
<td></td>
</tr>
<tr>
<td>4 Outline</td>
<td>240.5740</td>
<td>-1.0263</td>
<td>238.5000 -0.5300</td>
<td>234.6000 -0.4900</td>
<td></td>
<td></td>
</tr>
<tr>
<td>5 Outline</td>
<td>234.1000</td>
<td>-0.5700</td>
<td>227.2000 -0.6900</td>
<td>225.5000 -0.4200</td>
<td></td>
<td></td>
</tr>
<tr>
<td>6 Outline</td>
<td>218.9000</td>
<td>-0.1800</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>1 Outline</td>
<td>264.3000</td>
<td>-0.8700</td>
<td>256.8000 -1.1000</td>
<td>257.3975 -1.1142</td>
<td></td>
<td></td>
</tr>
<tr>
<td>2 Outline</td>
<td>259.4800</td>
<td>-1.1510</td>
<td>263.0500 -1.1850</td>
<td>264.9000 -1.2220</td>
<td></td>
<td></td>
</tr>
<tr>
<td>3 Outline</td>
<td>265.0612</td>
<td>-1.2337</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
### Appendix D

**GRAVMAG MODEL LISTINGS**

**Polygon 4 Outline**

<table>
<thead>
<tr>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>181.2000</td>
<td>2.0900</td>
<td>180.6000</td>
<td>-0.7700</td>
<td>184.5000</td>
<td>-0.8000</td>
</tr>
<tr>
<td>185.6835</td>
<td>-1.1945</td>
<td>186.0600</td>
<td>-1.2150</td>
<td>187.6500</td>
<td>-1.3410</td>
</tr>
<tr>
<td>188.9800</td>
<td>-1.4510</td>
<td>191.8000</td>
<td>-1.6680</td>
<td>194.5700</td>
<td>-1.8950</td>
</tr>
<tr>
<td>197.2100</td>
<td>-1.9460</td>
<td>199.7100</td>
<td>-1.8850</td>
<td>202.9400</td>
<td>-1.8800</td>
</tr>
<tr>
<td>205.4200</td>
<td>-1.5960</td>
<td>206.5900</td>
<td>-1.3720</td>
<td>208.2500</td>
<td>-1.3180</td>
</tr>
<tr>
<td>210.5700</td>
<td>-1.2620</td>
<td>212.6400</td>
<td>-1.2250</td>
<td>213.7301</td>
<td>-1.2123</td>
</tr>
<tr>
<td>214.1000</td>
<td>-0.3000</td>
<td>218.9000</td>
<td>-0.1800</td>
<td>225.5000</td>
<td>-0.4200</td>
</tr>
<tr>
<td>227.2000</td>
<td>-0.6900</td>
<td>234.1000</td>
<td>-0.5700</td>
<td>234.6000</td>
<td>-0.4900</td>
</tr>
<tr>
<td>238.5000</td>
<td>-0.5300</td>
<td>240.5740</td>
<td>-0.1025</td>
<td>241.5800</td>
<td>-1.0720</td>
</tr>
<tr>
<td>244.6200</td>
<td>-1.0140</td>
<td>251.0000</td>
<td>-1.0520</td>
<td>254.2700</td>
<td>-1.0590</td>
</tr>
<tr>
<td>256.6476</td>
<td>-1.0100</td>
<td>256.8000</td>
<td>-1.0500</td>
<td>264.3000</td>
<td>-0.8700</td>
</tr>
<tr>
<td>265.0612</td>
<td>-1.2357</td>
<td>266.7800</td>
<td>-1.3580</td>
<td>269.7600</td>
<td>-1.5090</td>
</tr>
<tr>
<td>274.3600</td>
<td>-1.7010</td>
<td>278.3600</td>
<td>-1.8650</td>
<td>280.8700</td>
<td>-1.8510</td>
</tr>
<tr>
<td>282.8800</td>
<td>-1.8450</td>
<td>285.8200</td>
<td>-1.9470</td>
<td>288.9000</td>
<td>-1.9470</td>
</tr>
<tr>
<td>296.9000</td>
<td>-1.8200</td>
<td>301.0300</td>
<td>-1.8370</td>
<td>307.3500</td>
<td>-1.7840</td>
</tr>
<tr>
<td>312.4600</td>
<td>-1.7910</td>
<td>320.4300</td>
<td>-1.7600</td>
<td>332.8600</td>
<td>-1.7230</td>
</tr>
<tr>
<td>330.5600</td>
<td>-1.7090</td>
<td>333.4000</td>
<td>-1.6794</td>
<td>333.4000</td>
<td>-1.6000</td>
</tr>
<tr>
<td>312.7000</td>
<td>-1.3400</td>
<td>311.7000</td>
<td>-1.0500</td>
<td>283.6000</td>
<td>-0.4600</td>
</tr>
<tr>
<td>273.3000</td>
<td>-0.1700</td>
<td>269.1000</td>
<td>-0.0500</td>
<td>267.1000</td>
<td>0.8200</td>
</tr>
<tr>
<td>260.3000</td>
<td>0.8300</td>
<td>258.1000</td>
<td>1.4200</td>
<td>251.7000</td>
<td>1.4700</td>
</tr>
<tr>
<td>242.4000</td>
<td>1.6200</td>
<td>240.1000</td>
<td>2.1500</td>
<td>224.0000</td>
<td>2.4500</td>
</tr>
<tr>
<td>212.9000</td>
<td>2.2500</td>
<td>209.0000</td>
<td>1.5700</td>
<td>193.7000</td>
<td>1.3100</td>
</tr>
<tr>
<td>186.3000</td>
<td>1.4700</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**Polygon 5 Outline**

<table>
<thead>
<tr>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>181.2000</td>
<td>2.0900</td>
<td>186.3000</td>
<td>1.4700</td>
<td>195.7000</td>
<td>1.3100</td>
</tr>
<tr>
<td>209.0000</td>
<td>1.5700</td>
<td>212.9000</td>
<td>2.2500</td>
<td>224.0000</td>
<td>2.4500</td>
</tr>
<tr>
<td>240.1000</td>
<td>2.1500</td>
<td>210.8000</td>
<td>3.4900</td>
<td>208.4000</td>
<td>3.5300</td>
</tr>
<tr>
<td>199.0000</td>
<td>3.5300</td>
<td>194.3000</td>
<td>4.4700</td>
<td>193.2000</td>
<td>8.0000</td>
</tr>
<tr>
<td>184.5000</td>
<td></td>
<td>8.0000</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**Polygon 6 Outline**

<table>
<thead>
<tr>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>136.2000</td>
<td>-1.9300</td>
<td>132.2150</td>
<td>-1.9884</td>
<td>134.6700</td>
<td>-2.0700</td>
</tr>
<tr>
<td>139.0500</td>
<td>-2.1790</td>
<td>144.3300</td>
<td>-2.1910</td>
<td>150.7200</td>
<td>-2.2020</td>
</tr>
<tr>
<td>164.6600</td>
<td>-2.3870</td>
<td>167.0000</td>
<td>-2.4260</td>
<td>169.2000</td>
<td>-2.4860</td>
</tr>
<tr>
<td>174.7400</td>
<td>-2.6470</td>
<td>176.0800</td>
<td>-2.6760</td>
<td>177.6196</td>
<td>-2.1664</td>
</tr>
<tr>
<td>159.0000</td>
<td>-1.0900</td>
<td>154.9000</td>
<td>-0.5400</td>
<td>150.6000</td>
<td>-0.8400</td>
</tr>
<tr>
<td>146.6000</td>
<td>-1.8000</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**Polygon 7 Outline**

<table>
<thead>
<tr>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>151.3570</td>
<td>18.4671</td>
<td>131.1732</td>
<td>1.7700</td>
<td>110.9483</td>
<td>-1.8576</td>
</tr>
<tr>
<td>112.1600</td>
<td>-1.8740</td>
<td>115.9400</td>
<td>-1.9700</td>
<td>119.5000</td>
<td>-2.0030</td>
</tr>
<tr>
<td>124.5100</td>
<td>-1.9590</td>
<td>131.1500</td>
<td>-1.9530</td>
<td>131.4880</td>
<td>-1.9642</td>
</tr>
<tr>
<td>136.2000</td>
<td>-1.9300</td>
<td>146.6000</td>
<td>-1.8000</td>
<td>150.6000</td>
<td>-0.8400</td>
</tr>
<tr>
<td>154.9000</td>
<td>-0.5400</td>
<td>159.0000</td>
<td>-1.0900</td>
<td>177.6196</td>
<td>-2.1664</td>
</tr>
<tr>
<td>180.3100</td>
<td>-1.2760</td>
<td>180.3573</td>
<td>-1.2735</td>
<td>180.6000</td>
<td>-0.7700</td>
</tr>
<tr>
<td>181.2000</td>
<td>2.0900</td>
<td>184.5000</td>
<td>8.0000</td>
<td>191.7000</td>
<td>18.5200</td>
</tr>
</tbody>
</table>
### Appendix D

#### GRAVMAG MODEL LISTINGS

<table>
<thead>
<tr>
<th>Polygon 8 Outline</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>-10000.0000</td>
<td>14.5300</td>
<td>0.0000</td>
<td>14.5300</td>
<td>10.3000</td>
<td>14.5300</td>
<td></td>
</tr>
<tr>
<td>47.2000</td>
<td>16.0700</td>
<td>85.1000</td>
<td>16.9700</td>
<td>92.9000</td>
<td>17.0000</td>
<td></td>
</tr>
<tr>
<td>103.7000</td>
<td>18.0000</td>
<td>191.7000</td>
<td>18.5200</td>
<td>204.1000</td>
<td>18.8000</td>
<td></td>
</tr>
<tr>
<td>212.8000</td>
<td>18.9000</td>
<td>253.7000</td>
<td>18.3000</td>
<td>269.0000</td>
<td>17.6000</td>
<td></td>
</tr>
<tr>
<td>298.0000</td>
<td>17.0000</td>
<td>308.8000</td>
<td>16.3000</td>
<td>381.1000</td>
<td>12.0000</td>
<td></td>
</tr>
<tr>
<td>387.7000</td>
<td>12.0000</td>
<td>432.2000</td>
<td>11.4800</td>
<td>453.5000</td>
<td>9.7200</td>
<td></td>
</tr>
<tr>
<td>499.4000</td>
<td>8.9791</td>
<td>10000.0000</td>
<td>7.0000</td>
<td>10000.0000</td>
<td>28.5000</td>
<td></td>
</tr>
<tr>
<td>503.0913</td>
<td>29.5673</td>
<td>453.5000</td>
<td>29.5500</td>
<td>398.3000</td>
<td>29.4000</td>
<td></td>
</tr>
<tr>
<td>351.5000</td>
<td>29.4000</td>
<td>281.3000</td>
<td>26.8800</td>
<td>263.9000</td>
<td>25.4000</td>
<td></td>
</tr>
<tr>
<td>239.3000</td>
<td>24.6000</td>
<td>197.5000</td>
<td>25.0000</td>
<td>176.2000</td>
<td>26.4000</td>
<td></td>
</tr>
<tr>
<td>100.0000</td>
<td>26.7500</td>
<td>87.7000</td>
<td>26.0000</td>
<td>51.1000</td>
<td>25.5000</td>
<td></td>
</tr>
<tr>
<td>18.7000</td>
<td>25.4000</td>
<td>0.0000</td>
<td>25.4000</td>
<td>-10000.0000</td>
<td>25.3700</td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Polygon 9 Outline</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>-10000.0000</td>
<td>25.3700</td>
<td>0.0000</td>
<td>25.4000</td>
<td>18.7000</td>
<td>25.4000</td>
<td></td>
</tr>
<tr>
<td>51.1000</td>
<td>25.5000</td>
<td>87.7000</td>
<td>26.0000</td>
<td>100.0000</td>
<td>26.7900</td>
<td></td>
</tr>
<tr>
<td>176.2000</td>
<td>26.4000</td>
<td>197.5000</td>
<td>25.0000</td>
<td>239.3000</td>
<td>24.6000</td>
<td></td>
</tr>
<tr>
<td>263.9000</td>
<td>25.2000</td>
<td>281.3000</td>
<td>26.8800</td>
<td>351.5000</td>
<td>29.4000</td>
<td></td>
</tr>
<tr>
<td>398.3000</td>
<td>29.4000</td>
<td>453.5000</td>
<td>29.5500</td>
<td>503.0913</td>
<td>29.5673</td>
<td></td>
</tr>
<tr>
<td>10000.0000</td>
<td>28.5000</td>
<td>10000.0000</td>
<td>30.0000</td>
<td>498.6625</td>
<td>33.5958</td>
<td></td>
</tr>
<tr>
<td>453.5000</td>
<td>34.0000</td>
<td>403.1100</td>
<td>36.7300</td>
<td>352.7200</td>
<td>34.7500</td>
<td></td>
</tr>
<tr>
<td>302.3300</td>
<td>34.2400</td>
<td>251.9400</td>
<td>28.1500</td>
<td>201.5600</td>
<td>30.2200</td>
<td></td>
</tr>
<tr>
<td>151.1700</td>
<td>36.8700</td>
<td>100.7800</td>
<td>36.2800</td>
<td>50.3900</td>
<td>34.8700</td>
<td></td>
</tr>
<tr>
<td>0.0000</td>
<td>33.6200</td>
<td>-10000.0000</td>
<td>33.6200</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Polygon 10 Outline</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>-10000.0000</td>
<td>33.6200</td>
<td>0.0000</td>
<td>33.6200</td>
<td>50.3900</td>
<td>34.8700</td>
<td></td>
</tr>
<tr>
<td>100.7800</td>
<td>36.2800</td>
<td>151.1700</td>
<td>36.8700</td>
<td>201.5600</td>
<td>30.2200</td>
<td></td>
</tr>
<tr>
<td>251.9400</td>
<td>28.1500</td>
<td>302.3300</td>
<td>34.2400</td>
<td>352.7200</td>
<td>34.7500</td>
<td></td>
</tr>
<tr>
<td>403.1100</td>
<td>36.7300</td>
<td>453.5000</td>
<td>34.0000</td>
<td>498.6625</td>
<td>34.0000</td>
<td></td>
</tr>
<tr>
<td>10000.0000</td>
<td>30.0000</td>
<td>10000.0000</td>
<td>60.0000</td>
<td>-10000.0000</td>
<td>60.0000</td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Polygon 11 Outline</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>-10000.0000</td>
<td>60.0000</td>
<td>10000.0000</td>
<td>60.0000</td>
<td>10000.0000</td>
<td>70.0000</td>
<td></td>
</tr>
<tr>
<td>-10000.0000</td>
<td>70.0000</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Polygon 12 Outline</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>151.1700</td>
<td>36.8700</td>
<td>201.5600</td>
<td>30.2200</td>
<td>251.9400</td>
<td>28.1500</td>
<td></td>
</tr>
<tr>
<td>302.3300</td>
<td>34.2400</td>
<td>352.7200</td>
<td>34.7500</td>
<td>243.4043</td>
<td>59.6570</td>
<td></td>
</tr>
<tr>
<td>210.6448</td>
<td>59.8406</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Polygon 13 Outline</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>193.2000</td>
<td>8.8000</td>
<td>194.5000</td>
<td>4.4700</td>
<td>199.0000</td>
<td>3.5300</td>
<td></td>
</tr>
<tr>
<td>208.4000</td>
<td>3.5300</td>
<td>210.8000</td>
<td>3.4900</td>
<td>215.3000</td>
<td>4.2800</td>
<td></td>
</tr>
<tr>
<td>236.1000</td>
<td>4.1800</td>
<td>238.4000</td>
<td>3.9400</td>
<td>239.8000</td>
<td>3.6400</td>
<td></td>
</tr>
</tbody>
</table>
### Appendix D

#### GRAVMAG MODEL LISTINGS

<table>
<thead>
<tr>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>259.9000</td>
<td>3.6400</td>
<td>263.3000</td>
<td>3.2000</td>
<td>265.9000</td>
<td>2.6500</td>
</tr>
<tr>
<td>270.7000</td>
<td>2.5100</td>
<td>281.0848</td>
<td>17.3600</td>
<td>269.0000</td>
<td>17.6000</td>
</tr>
<tr>
<td>253.7000</td>
<td>18.3000</td>
<td>212.8000</td>
<td>18.9000</td>
<td>204.1000</td>
<td>18.8000</td>
</tr>
<tr>
<td>191.7000</td>
<td>18.5200</td>
<td>184.5000</td>
<td>8.8000</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

#### Polygon 14 Outline

<table>
<thead>
<tr>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>267.1000</td>
<td>0.8200</td>
<td>269.1000</td>
<td>-0.0500</td>
<td>273.3000</td>
<td>-0.1700</td>
</tr>
<tr>
<td>283.6000</td>
<td>-0.4600</td>
<td>311.7000</td>
<td>-1.0500</td>
<td>312.7000</td>
<td>-1.3400</td>
</tr>
<tr>
<td>253.7000</td>
<td>18.3000</td>
<td>212.8000</td>
<td>18.9000</td>
<td>204.1000</td>
<td>18.8000</td>
</tr>
<tr>
<td>191.7000</td>
<td>18.5200</td>
<td>184.5000</td>
<td>8.8000</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

#### Polygon 15 Outline

<table>
<thead>
<tr>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>270.7000</td>
<td>2.5100</td>
<td>275.5000</td>
<td>1.9300</td>
<td>275.9000</td>
<td>1.4200</td>
</tr>
<tr>
<td>305.6000</td>
<td>-0.2400</td>
<td>310.5000</td>
<td>-0.3300</td>
<td>311.9000</td>
<td>-0.8600</td>
</tr>
<tr>
<td>326.7000</td>
<td>-1.1400</td>
<td>333.4000</td>
<td>-1.6000</td>
<td>336.6258</td>
<td>-1.5807</td>
</tr>
<tr>
<td>338.3500</td>
<td>-1.4600</td>
<td>342.0500</td>
<td>-1.4280</td>
<td>344.5500</td>
<td>-1.2850</td>
</tr>
<tr>
<td>350.0400</td>
<td>-1.3400</td>
<td>355.4400</td>
<td>-1.2430</td>
<td>356.3300</td>
<td>-1.1820</td>
</tr>
<tr>
<td>360.6400</td>
<td>-1.2700</td>
<td>365.6300</td>
<td>-1.1660</td>
<td>371.8300</td>
<td>-1.2410</td>
</tr>
<tr>
<td>377.8300</td>
<td>-1.2480</td>
<td>381.1000</td>
<td>12.0000</td>
<td>308.8000</td>
<td>16.3000</td>
</tr>
<tr>
<td>298.0000</td>
<td>17.0000</td>
<td>281.0848</td>
<td>17.3600</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

#### Polygon 16 Outline

<table>
<thead>
<tr>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>240.1000</td>
<td>2.1500</td>
<td>242.4000</td>
<td>1.6200</td>
<td>251.7000</td>
<td>1.4700</td>
</tr>
<tr>
<td>270.7000</td>
<td>2.5100</td>
<td>265.9000</td>
<td>2.6500</td>
<td>263.3000</td>
<td>3.2000</td>
</tr>
<tr>
<td>259.9000</td>
<td>3.6400</td>
<td>238.8000</td>
<td>3.6400</td>
<td>238.4000</td>
<td>3.9400</td>
</tr>
</tbody>
</table>

#### Calculated Gravity Anomaly (gravity units)

<table>
<thead>
<tr>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>99.9903</td>
<td>86.9262</td>
<td>77.5970</td>
<td>66.5502</td>
<td>44.9697</td>
<td>29.1172</td>
</tr>
<tr>
<td>-59.5710</td>
<td>-73.1649</td>
<td>-98.2816</td>
<td>-95.0811</td>
<td>-107.9223</td>
<td>-121.3267</td>
</tr>
<tr>
<td>-134.0453</td>
<td>-155.5700</td>
<td>-177.0607</td>
<td>-190.2852</td>
<td>-199.6793</td>
<td>-205.3898</td>
</tr>
<tr>
<td>-90.2991</td>
<td>6.6761</td>
<td>67.6500</td>
<td>133.8336</td>
<td>203.5488</td>
<td>208.4562</td>
</tr>
<tr>
<td>319.0258</td>
<td>329.7405</td>
<td>355.2110</td>
<td>307.2029</td>
<td>285.7573</td>
<td>111.9867</td>
</tr>
<tr>
<td>-259.7562</td>
<td>-221.0579</td>
<td>-165.0312</td>
<td>-108.5762</td>
<td>-54.8661</td>
<td>2.2943</td>
</tr>
<tr>
<td>35.7259</td>
<td>54.0835</td>
<td>73.0562</td>
<td>81.3388</td>
<td>2.5470</td>
<td>2.6941</td>
</tr>
<tr>
<td>45.5706</td>
<td>82.4873</td>
<td>165.0328</td>
<td>194.2623</td>
<td>266.6040</td>
<td>315.0273</td>
</tr>
<tr>
<td>345.4124</td>
<td>355.3674</td>
<td>354.5080</td>
<td>348.6880</td>
<td>338.6634</td>
<td>333.7891</td>
</tr>
<tr>
<td>304.0905</td>
<td>242.3389</td>
<td>165.8018</td>
<td>137.0580</td>
<td>129.3709</td>
<td>119.9449</td>
</tr>
<tr>
<td>96.1199</td>
<td>81.9267</td>
<td>72.1346</td>
<td>69.0139</td>
<td>98.9851</td>
<td>140.2030</td>
</tr>
<tr>
<td>169.9869</td>
<td>178.9073</td>
<td>214.9471</td>
<td>247.8161</td>
<td>249.4850</td>
<td>251.4317</td>
</tr>
<tr>
<td>251.9699</td>
<td>255.6889</td>
<td>258.7595</td>
<td>259.6429</td>
<td>263.6725</td>
<td>263.5054</td>
</tr>
<tr>
<td>262.2424</td>
<td>242.4903</td>
<td>168.0854</td>
<td>129.3899</td>
<td>103.5412</td>
<td>94.8911</td>
</tr>
</tbody>
</table>
### Appendix D

#### GRAVMAG MODEL LISTINGS

<p>| | | | | | | | |</p>
<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>89.7890</td>
<td>90.8679</td>
<td>94.6930</td>
<td>99.2430</td>
<td>114.8684</td>
<td>142.3811</td>
<td></td>
<td></td>
</tr>
<tr>
<td>184.1053</td>
<td>224.1499</td>
<td>260.9823</td>
<td>290.5642</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**Observed Gravity Anomaly (gravity units)**

<p>| | | | | | | | |</p>
<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>414.4050</td>
<td>416.5210</td>
<td>466.6060</td>
<td>485.6945</td>
<td>501.8478</td>
<td>276.9558</td>
<td></td>
<td></td>
</tr>
<tr>
<td>282.5298</td>
<td>268.0792</td>
<td>228.1145</td>
<td>263.5272</td>
<td>271.6047</td>
<td>241.7880</td>
<td></td>
<td></td>
</tr>
<tr>
<td>229.1805</td>
<td>206.3660</td>
<td>204.8733</td>
<td>66.9957</td>
<td>20.1903</td>
<td>-28.6985</td>
<td></td>
<td></td>
</tr>
<tr>
<td>-113.3850</td>
<td>-201.6600</td>
<td>-154.8600</td>
<td>-154.8361</td>
<td>-71.0490</td>
<td>-60.0750</td>
<td></td>
<td></td>
</tr>
<tr>
<td>18.6665</td>
<td>27.0225</td>
<td>77.0305</td>
<td>129.3075</td>
<td>178.8195</td>
<td>203.1000</td>
<td></td>
<td></td>
</tr>
<tr>
<td>340.0750</td>
<td>342.7500</td>
<td>325.0000</td>
<td>325.0750</td>
<td>322.4000</td>
<td>187.1125</td>
<td></td>
<td></td>
</tr>
<tr>
<td>-1.4500</td>
<td>-142.7750</td>
<td>-229.2625</td>
<td>-265.0750</td>
<td>-281.0625</td>
<td>-275.7250</td>
<td></td>
<td></td>
</tr>
<tr>
<td>-248.0150</td>
<td>-205.6548</td>
<td>-125.0768</td>
<td>-84.2643</td>
<td>-52.6746</td>
<td>49.1715</td>
<td></td>
<td></td>
</tr>
<tr>
<td>101.2118</td>
<td>148.4313</td>
<td>156.5453</td>
<td>141.0780</td>
<td>96.2808</td>
<td>60.6844</td>
<td></td>
<td></td>
</tr>
<tr>
<td>10.5155</td>
<td>32.0265</td>
<td>156.9363</td>
<td>272.0480</td>
<td>316.9965</td>
<td>318.2610</td>
<td></td>
<td></td>
</tr>
<tr>
<td>325.0291</td>
<td>334.9500</td>
<td>315.2465</td>
<td>305.5659</td>
<td>209.2476</td>
<td>157.4550</td>
<td></td>
<td></td>
</tr>
<tr>
<td>108.6010</td>
<td>74.4919</td>
<td>88.3180</td>
<td>128.5181</td>
<td>128.3685</td>
<td>137.4440</td>
<td></td>
<td></td>
</tr>
<tr>
<td>148.2410</td>
<td>154.7195</td>
<td>102.0950</td>
<td>84.9766</td>
<td>96.4924</td>
<td>142.1736</td>
<td></td>
<td></td>
</tr>
<tr>
<td>226.0568</td>
<td>229.3136</td>
<td>247.9460</td>
<td>282.9034</td>
<td>320.7738</td>
<td>378.0613</td>
<td></td>
<td></td>
</tr>
<tr>
<td>390.1238</td>
<td>411.8790</td>
<td>398.5940</td>
<td>371.8518</td>
<td>347.8140</td>
<td>341.7192</td>
<td></td>
<td></td>
</tr>
<tr>
<td>314.9642</td>
<td>280.8143</td>
<td>202.6643</td>
<td>142.9342</td>
<td>113.4640</td>
<td>93.5520</td>
<td></td>
<td></td>
</tr>
<tr>
<td>62.4640</td>
<td>80.0320</td>
<td>84.2321</td>
<td>68.9600</td>
<td>88.5599</td>
<td>59.7520</td>
<td></td>
<td></td>
</tr>
<tr>
<td>150.6887</td>
<td>242.5062</td>
<td>370.7015</td>
<td>412.5563</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
**Appendix D**

**GRAVMAG MODEL LISTINGS**

**Model Title:** KRISP 90 EW Line (Inverted Model - B3)

**Profile Orientation:** 0.0000 Degrees

**Earth's Magnetic Field:**
- Declination: 38.2000 A/m
- Inclination: -7.000 Degrees

**Gravity Calculation Surface:** 0.000 Km Above Observations

<table>
<thead>
<tr>
<th>Polygon Number</th>
<th>Density (g/cm³)</th>
<th>Susc</th>
<th>Rem. Declin. A/m</th>
<th>Rem. Inclin. Degrees</th>
<th>Rem. Strike Degrees</th>
<th>Half Km</th>
</tr>
</thead>
<tbody>
<tr>
<td>Backgnd</td>
<td>2.7000</td>
<td>0.0000</td>
<td>0.0000</td>
<td>-7.000</td>
<td>68.000</td>
<td>10.00</td>
</tr>
<tr>
<td>1</td>
<td>2.1000</td>
<td>0.0000</td>
<td>0.0000</td>
<td>-7.000</td>
<td>68.000</td>
<td>10.00</td>
</tr>
<tr>
<td>2</td>
<td>2.1000</td>
<td>0.0000</td>
<td>0.0000</td>
<td>-7.000</td>
<td>68.000</td>
<td>10.00</td>
</tr>
<tr>
<td>3</td>
<td>2.3000</td>
<td>0.0000</td>
<td>0.0000</td>
<td>-7.000</td>
<td>68.000</td>
<td>10.00</td>
</tr>
<tr>
<td>4</td>
<td>2.4600</td>
<td>0.0000</td>
<td>0.0000</td>
<td>-7.000</td>
<td>68.000</td>
<td>100.00</td>
</tr>
<tr>
<td>5</td>
<td>2.5000</td>
<td>0.0000</td>
<td>0.0000</td>
<td>-7.000</td>
<td>68.000</td>
<td>100.00</td>
</tr>
<tr>
<td>6</td>
<td>2.6000</td>
<td>0.0000</td>
<td>0.0000</td>
<td>-7.000</td>
<td>68.000</td>
<td>100.00</td>
</tr>
<tr>
<td>7</td>
<td>2.7961</td>
<td>0.0000</td>
<td>0.0000</td>
<td>-7.000</td>
<td>68.000</td>
<td>100.00</td>
</tr>
<tr>
<td>8</td>
<td>2.8526</td>
<td>0.0000</td>
<td>0.0000</td>
<td>-7.000</td>
<td>68.000</td>
<td>10000.00</td>
</tr>
<tr>
<td>9</td>
<td>2.9016</td>
<td>0.0000</td>
<td>0.0000</td>
<td>-7.000</td>
<td>68.000</td>
<td>10000.00</td>
</tr>
<tr>
<td>10</td>
<td>3.2726</td>
<td>0.0000</td>
<td>0.0000</td>
<td>-7.000</td>
<td>68.000</td>
<td>10000.00</td>
</tr>
<tr>
<td>11</td>
<td>0.9400</td>
<td>0.0000</td>
<td>0.0000</td>
<td>-7.000</td>
<td>68.000</td>
<td>10000.00</td>
</tr>
<tr>
<td>12</td>
<td>3.2194</td>
<td>0.0000</td>
<td>0.0000</td>
<td>-7.000</td>
<td>68.000</td>
<td>10000.00</td>
</tr>
<tr>
<td>13</td>
<td>2.8278</td>
<td>0.0000</td>
<td>0.0000</td>
<td>0.000</td>
<td>0.000</td>
<td>200.00</td>
</tr>
<tr>
<td>14</td>
<td>2.5500</td>
<td>0.0000</td>
<td>0.0000</td>
<td>-7.000</td>
<td>68.000</td>
<td>100.00</td>
</tr>
<tr>
<td>15</td>
<td>2.7642</td>
<td>0.0000</td>
<td>0.0000</td>
<td>0.000</td>
<td>0.000</td>
<td>100.00</td>
</tr>
<tr>
<td>16</td>
<td>2.7500</td>
<td>0.0000</td>
<td>0.0000</td>
<td>-7.000</td>
<td>68.000</td>
<td>100.00</td>
</tr>
<tr>
<td>17</td>
<td>3.2376</td>
<td>0.0000</td>
<td>0.0000</td>
<td>-7.000</td>
<td>68.000</td>
<td>10000.00</td>
</tr>
</tbody>
</table>

**Polygon 1 Outline**

<table>
<thead>
<tr>
<th>X</th>
<th>Z</th>
<th>180.6000</th>
<th>-0.7700</th>
<th>180.3573</th>
<th>-1.2735</th>
<th>181.5600</th>
<th>-1.2100</th>
</tr>
</thead>
<tbody>
<tr>
<td>Z</td>
<td>X</td>
<td>182.6200</td>
<td>-1.1620</td>
<td>184.6100</td>
<td>-1.1360</td>
<td>185.6835</td>
<td>-1.1945</td>
</tr>
<tr>
<td>X</td>
<td>Z</td>
<td>184.5000</td>
<td>-0.8000</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**Polygon 2 Outline**

<table>
<thead>
<tr>
<th>X</th>
<th>Z</th>
<th>214.1000</th>
<th>-0.3000</th>
<th>213.3701</th>
<th>-1.2123</th>
<th>214.3100</th>
<th>-1.1960</th>
</tr>
</thead>
<tbody>
<tr>
<td>Z</td>
<td>X</td>
<td>215.8600</td>
<td>-1.1420</td>
<td>220.1400</td>
<td>-1.0490</td>
<td>222.8200</td>
<td>-1.0230</td>
</tr>
<tr>
<td>X</td>
<td>Z</td>
<td>227.6500</td>
<td>-1.0060</td>
<td>231.4400</td>
<td>-0.9740</td>
<td>239.2700</td>
<td>-0.9670</td>
</tr>
<tr>
<td>Z</td>
<td>X</td>
<td>240.5740</td>
<td>-1.0263</td>
<td>238.5000</td>
<td>-0.5300</td>
<td>234.6000</td>
<td>-0.4900</td>
</tr>
<tr>
<td>X</td>
<td>Z</td>
<td>234.1000</td>
<td>-0.5700</td>
<td>227.2000</td>
<td>-0.6900</td>
<td>225.5000</td>
<td>-0.4200</td>
</tr>
<tr>
<td>Z</td>
<td>X</td>
<td>218.9000</td>
<td>-0.1800</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**Polygon 3 Outline**

<table>
<thead>
<tr>
<th>X</th>
<th>Z</th>
<th>264.3000</th>
<th>-0.8700</th>
<th>256.8000</th>
<th>-1.1000</th>
<th>257.3975</th>
<th>-1.1142</th>
</tr>
</thead>
<tbody>
<tr>
<td>Z</td>
<td>X</td>
<td>259.4800</td>
<td>-1.1510</td>
<td>263.0500</td>
<td>-1.1850</td>
<td>264.9000</td>
<td>-1.2220</td>
</tr>
<tr>
<td>X</td>
<td>Z</td>
<td>265.0612</td>
<td>-1.2337</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
### Appendix D

**GRAVMAG MODEL LISTINGS**

<table>
<thead>
<tr>
<th>Polygon</th>
<th>Outline 4</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>181.2000</td>
<td>2.0900</td>
<td>180.6000</td>
<td>-0.7700</td>
<td>184.5000</td>
<td>-0.8000</td>
</tr>
<tr>
<td>185.6835</td>
<td>-1.1945</td>
<td>186.0600</td>
<td>-1.2150</td>
<td>187.6500</td>
<td>-1.3410</td>
</tr>
<tr>
<td>188.9800</td>
<td>-1.4510</td>
<td>191.8000</td>
<td>-1.6680</td>
<td>194.5700</td>
<td>-1.8950</td>
</tr>
<tr>
<td>197.2100</td>
<td>-1.9460</td>
<td>199.7100</td>
<td>-1.8850</td>
<td>202.9400</td>
<td>-1.8900</td>
</tr>
<tr>
<td>205.4200</td>
<td>-1.5960</td>
<td>206.5900</td>
<td>-1.3720</td>
<td>208.2500</td>
<td>-1.3180</td>
</tr>
<tr>
<td>210.5700</td>
<td>-1.2620</td>
<td>212.6400</td>
<td>-1.2250</td>
<td>213.3700</td>
<td>-1.2123</td>
</tr>
<tr>
<td>214.1000</td>
<td>-0.3000</td>
<td>218.9000</td>
<td>-0.1800</td>
<td>225.5000</td>
<td>-0.4200</td>
</tr>
<tr>
<td>227.2000</td>
<td>-0.6900</td>
<td>234.1000</td>
<td>-0.5700</td>
<td>234.6000</td>
<td>-0.4900</td>
</tr>
<tr>
<td>238.3000</td>
<td>0.5300</td>
<td>240.5700</td>
<td>0.0263</td>
<td>241.5800</td>
<td>-1.0720</td>
</tr>
<tr>
<td>244.6200</td>
<td>-1.0140</td>
<td>251.0000</td>
<td>-1.0520</td>
<td>254.2700</td>
<td>-1.0590</td>
</tr>
<tr>
<td>256.6746</td>
<td>-1.1010</td>
<td>256.8000</td>
<td>-1.1000</td>
<td>264.3000</td>
<td>-0.8700</td>
</tr>
<tr>
<td>265.0612</td>
<td>-1.2337</td>
<td>266.7800</td>
<td>-1.3580</td>
<td>269.7600</td>
<td>-1.5090</td>
</tr>
<tr>
<td>274.3600</td>
<td>-1.7010</td>
<td>278.3600</td>
<td>-1.8650</td>
<td>280.8700</td>
<td>-1.8510</td>
</tr>
<tr>
<td>282.8800</td>
<td>-1.8450</td>
<td>285.8200</td>
<td>-1.9470</td>
<td>288.8900</td>
<td>-1.9470</td>
</tr>
<tr>
<td>296.9000</td>
<td>-1.8200</td>
<td>301.0300</td>
<td>-1.8370</td>
<td>307.3500</td>
<td>-1.7840</td>
</tr>
<tr>
<td>312.4600</td>
<td>-1.7910</td>
<td>320.4300</td>
<td>-1.7600</td>
<td>322.8600</td>
<td>-1.7230</td>
</tr>
<tr>
<td>330.5600</td>
<td>-1.7090</td>
<td>333.4000</td>
<td>-1.6794</td>
<td>333.4000</td>
<td>-1.6000</td>
</tr>
<tr>
<td>312.7000</td>
<td>-1.3400</td>
<td>311.7000</td>
<td>-1.0500</td>
<td>283.6000</td>
<td>-0.4600</td>
</tr>
<tr>
<td>273.3000</td>
<td>-0.1700</td>
<td>269.1000</td>
<td>-0.0500</td>
<td>267.1000</td>
<td>0.8200</td>
</tr>
<tr>
<td>260.3000</td>
<td>0.8300</td>
<td>258.1000</td>
<td>1.4200</td>
<td>251.7000</td>
<td>1.4700</td>
</tr>
<tr>
<td>242.0000</td>
<td>1.6200</td>
<td>240.1000</td>
<td>2.1500</td>
<td>224.0000</td>
<td>2.4500</td>
</tr>
<tr>
<td>212.9000</td>
<td>2.2500</td>
<td>209.0000</td>
<td>1.5700</td>
<td>195.7000</td>
<td>1.3100</td>
</tr>
<tr>
<td>186.3000</td>
<td>1.4700</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Polygon</th>
<th>Outline 5</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>181.2000</td>
<td>2.0900</td>
<td>186.3000</td>
<td>1.4700</td>
<td>195.7000</td>
<td>1.3100</td>
</tr>
<tr>
<td>209.0000</td>
<td>1.5700</td>
<td>212.9000</td>
<td>2.2500</td>
<td>224.0000</td>
<td>2.4500</td>
</tr>
<tr>
<td>240.1000</td>
<td>2.1500</td>
<td>210.8000</td>
<td>3.4900</td>
<td>208.4000</td>
<td>3.5300</td>
</tr>
<tr>
<td>199.0000</td>
<td>3.5300</td>
<td>194.5000</td>
<td>4.4700</td>
<td>193.2000</td>
<td>8.8000</td>
</tr>
<tr>
<td>184.5000</td>
<td>8.8000</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Polygon</th>
<th>Outline 6</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>136.2000</td>
<td>-1.9300</td>
<td>132.2150</td>
<td>-1.9884</td>
<td>134.6700</td>
<td>-2.0700</td>
</tr>
<tr>
<td>139.0500</td>
<td>-2.1790</td>
<td>144.3300</td>
<td>-2.1910</td>
<td>150.7200</td>
<td>-2.2020</td>
</tr>
<tr>
<td>164.6600</td>
<td>-2.3870</td>
<td>167.0000</td>
<td>-2.4260</td>
<td>169.2000</td>
<td>-2.4860</td>
</tr>
<tr>
<td>174.7400</td>
<td>-2.6470</td>
<td>176.0800</td>
<td>-2.6760</td>
<td>177.6196</td>
<td>-2.1644</td>
</tr>
<tr>
<td>159.0000</td>
<td>-1.0900</td>
<td>154.9000</td>
<td>-0.5400</td>
<td>150.6000</td>
<td>-0.8400</td>
</tr>
<tr>
<td>146.6000</td>
<td>-1.8000</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Polygon</th>
<th>Outline 7</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>152.2586</td>
<td>17.4975</td>
<td>155.6323</td>
<td>-0.5160</td>
<td>112.1721</td>
<td>1.8743</td>
</tr>
<tr>
<td>112.1600</td>
<td>1.8740</td>
<td>115.9400</td>
<td>-1.9700</td>
<td>119.5000</td>
<td>-2.0030</td>
</tr>
<tr>
<td>124.5100</td>
<td>-1.9390</td>
<td>131.1500</td>
<td>-1.9530</td>
<td>131.4880</td>
<td>-1.9642</td>
</tr>
<tr>
<td>136.2000</td>
<td>-1.9300</td>
<td>146.6000</td>
<td>-1.8000</td>
<td>150.6000</td>
<td>-0.8400</td>
</tr>
<tr>
<td>154.9000</td>
<td>-0.5400</td>
<td>159.0000</td>
<td>-1.0900</td>
<td>177.6196</td>
<td>-2.1644</td>
</tr>
<tr>
<td>180.3100</td>
<td>-1.2760</td>
<td>180.3573</td>
<td>-1.2735</td>
<td>180.6000</td>
<td>-0.7700</td>
</tr>
<tr>
<td>181.2000</td>
<td>2.0900</td>
<td>184.5000</td>
<td>8.8000</td>
<td>191.7000</td>
<td>18.5200</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Polygon</th>
<th>Outline 8</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
### Appendix D

#### GRAVMAG MODEL LISTINGS

<table>
<thead>
<tr>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>-10000.000</td>
<td>14.5300</td>
<td>0.0000</td>
<td>14.5300</td>
<td>10.3000</td>
<td>14.5300</td>
</tr>
<tr>
<td>47.2000</td>
<td>16.0700</td>
<td>85.1000</td>
<td>16.9700</td>
<td>92.9000</td>
<td>17.0000</td>
</tr>
<tr>
<td>103.7000</td>
<td>18.0000</td>
<td>191.7000</td>
<td>18.5200</td>
<td>204.1000</td>
<td>18.8000</td>
</tr>
<tr>
<td>212.8000</td>
<td>18.9000</td>
<td>253.7000</td>
<td>18.3000</td>
<td>269.0000</td>
<td>17.6600</td>
</tr>
<tr>
<td>398.0000</td>
<td>17.0000</td>
<td>308.8000</td>
<td>16.5000</td>
<td>381.1000</td>
<td>12.0000</td>
</tr>
<tr>
<td>387.7000</td>
<td>12.0000</td>
<td>432.2000</td>
<td>11.4800</td>
<td>453.5000</td>
<td>9.7200</td>
</tr>
<tr>
<td>499.4000</td>
<td>8.9791</td>
<td>10000.000</td>
<td>7.0000</td>
<td>10000.000</td>
<td>28.5000</td>
</tr>
<tr>
<td>503.0913</td>
<td>29.5673</td>
<td>453.5000</td>
<td>29.5500</td>
<td>398.3000</td>
<td>29.4000</td>
</tr>
<tr>
<td>100.0000</td>
<td>0.0000</td>
<td>191.7000</td>
<td>18.5200</td>
<td>204.1000</td>
<td>18.8000</td>
</tr>
<tr>
<td>18.7000</td>
<td>25.4000</td>
<td>0.0000</td>
<td>25.4000</td>
<td>-10000.000</td>
<td>25.3700</td>
</tr>
</tbody>
</table>

**Polygon 9 Outline**

<table>
<thead>
<tr>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>-10000.000</td>
<td>25.3700</td>
<td>0.0000</td>
<td>25.4000</td>
<td>18.7000</td>
<td>25.4000</td>
</tr>
<tr>
<td>51.1000</td>
<td>25.5000</td>
<td>87.7000</td>
<td>26.0000</td>
<td>100.0000</td>
<td>26.7500</td>
</tr>
<tr>
<td>176.2000</td>
<td>26.4000</td>
<td>197.5000</td>
<td>25.0000</td>
<td>239.3000</td>
<td>24.6000</td>
</tr>
<tr>
<td>263.9000</td>
<td>25.2000</td>
<td>281.3000</td>
<td>26.8800</td>
<td>351.5000</td>
<td>29.4000</td>
</tr>
<tr>
<td>398.3000</td>
<td>29.4000</td>
<td>453.5000</td>
<td>29.5500</td>
<td>503.0913</td>
<td>29.5673</td>
</tr>
<tr>
<td>10000.000</td>
<td>28.5000</td>
<td>10000.000</td>
<td>30.0000</td>
<td>498.6625</td>
<td>33.5958</td>
</tr>
<tr>
<td>453.5000</td>
<td>34.0000</td>
<td>403.1100</td>
<td>36.7300</td>
<td>352.7200</td>
<td>34.7500</td>
</tr>
<tr>
<td>302.3300</td>
<td>34.2400</td>
<td>251.9400</td>
<td>28.1500</td>
<td>201.5600</td>
<td>30.2200</td>
</tr>
<tr>
<td>151.1700</td>
<td>36.8700</td>
<td>100.7800</td>
<td>36.2800</td>
<td>50.3900</td>
<td>34.8700</td>
</tr>
<tr>
<td>0.0000</td>
<td>33.6200</td>
<td>-10000.000</td>
<td>33.6200</td>
<td>453.5000</td>
<td>34.0000</td>
</tr>
<tr>
<td>33.6200</td>
<td>0.0000</td>
<td>33.6200</td>
<td>50.3900</td>
<td>34.8700</td>
<td></td>
</tr>
<tr>
<td>100.0000</td>
<td>60.0000</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**Polygon 10 Outline**

<table>
<thead>
<tr>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>-10000.000</td>
<td>33.6200</td>
<td>0.0000</td>
<td>33.6200</td>
<td>50.3900</td>
<td>34.8700</td>
</tr>
<tr>
<td>100.7800</td>
<td>36.2800</td>
<td>151.1700</td>
<td>36.8700</td>
<td>210.6448</td>
<td>60.0000</td>
</tr>
<tr>
<td>-10000.000</td>
<td>60.0000</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**Polygon 11 Outline**

<table>
<thead>
<tr>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>-10000.000</td>
<td>60.0000</td>
<td>10000.000</td>
<td>60.0000</td>
<td>10000.000</td>
<td>70.0000</td>
</tr>
<tr>
<td>-10000.000</td>
<td>70.0000</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**Polygon 12 Outline**

<table>
<thead>
<tr>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>151.1700</td>
<td>36.8700</td>
<td>201.5600</td>
<td>30.2200</td>
<td>251.9400</td>
<td>28.1500</td>
</tr>
<tr>
<td>302.3300</td>
<td>34.2400</td>
<td>352.7200</td>
<td>34.7500</td>
<td>243.4043</td>
<td>60.0000</td>
</tr>
<tr>
<td>210.6448</td>
<td>60.0000</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**Polygon 13 Outline**

<table>
<thead>
<tr>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>193.2000</td>
<td>8.8000</td>
<td>194.5000</td>
<td>4.4700</td>
<td>199.0000</td>
<td>3.5300</td>
</tr>
<tr>
<td>208.4000</td>
<td>3.5300</td>
<td>210.8000</td>
<td>3.4900</td>
<td>215.3000</td>
<td>4.2800</td>
</tr>
<tr>
<td>236.1000</td>
<td>4.1800</td>
<td>238.4000</td>
<td>3.9400</td>
<td>239.8000</td>
<td>3.6400</td>
</tr>
<tr>
<td>259.9000</td>
<td>3.6400</td>
<td>263.3000</td>
<td>3.2000</td>
<td>269.0000</td>
<td>2.6500</td>
</tr>
<tr>
<td>270.7000</td>
<td>2.5100</td>
<td>281.0848</td>
<td>17.3600</td>
<td>269.0000</td>
<td>17.6000</td>
</tr>
<tr>
<td>253.7000</td>
<td>18.3000</td>
<td>212.8000</td>
<td>18.9000</td>
<td>204.1000</td>
<td>18.8000</td>
</tr>
<tr>
<td>191.7000</td>
<td>18.5200</td>
<td>184.5000</td>
<td>8.8000</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Appendix D

**GRAVMAG MODEL LISTINGS**

### Polygon 14 Outline

<table>
<thead>
<tr>
<th></th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>267.1000</td>
<td>0.8200</td>
<td>269.1000</td>
<td>-0.0500</td>
<td>273.3000</td>
<td>-0.1700</td>
<td></td>
</tr>
<tr>
<td>333.4000</td>
<td>-1.6000</td>
<td>326.7000</td>
<td>-1.1400</td>
<td>311.9000</td>
<td>-0.8600</td>
<td></td>
</tr>
<tr>
<td>310.5000</td>
<td>-0.3300</td>
<td>305.6000</td>
<td>-0.2400</td>
<td>302.3000</td>
<td>0.2400</td>
<td></td>
</tr>
<tr>
<td>283.3000</td>
<td>0.5700</td>
<td>278.5000</td>
<td>0.7800</td>
<td>275.9000</td>
<td>1.4200</td>
<td></td>
</tr>
<tr>
<td>275.3000</td>
<td>1.9500</td>
<td>270.7000</td>
<td>2.5100</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

### Polygon 15 Outline

<table>
<thead>
<tr>
<th></th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>270.7000</td>
<td>2.5100</td>
<td>275.5000</td>
<td>1.9300</td>
<td>275.9000</td>
<td>1.4200</td>
<td></td>
</tr>
<tr>
<td>305.6000</td>
<td>-0.2400</td>
<td>310.5000</td>
<td>-0.3300</td>
<td>311.9000</td>
<td>-0.8600</td>
<td></td>
</tr>
<tr>
<td>333.4000</td>
<td>-1.1400</td>
<td>333.4000</td>
<td>-1.6000</td>
<td>336.6258</td>
<td>-1.5807</td>
<td></td>
</tr>
<tr>
<td>338.5500</td>
<td>-1.4600</td>
<td>342.0500</td>
<td>-1.4280</td>
<td>344.5500</td>
<td>-1.2850</td>
<td></td>
</tr>
<tr>
<td>350.0400</td>
<td>-1.2460</td>
<td>353.4400</td>
<td>-1.2430</td>
<td>356.3300</td>
<td>-1.1820</td>
<td></td>
</tr>
<tr>
<td>360.6400</td>
<td>-1.2790</td>
<td>365.6300</td>
<td>-1.1660</td>
<td>371.8300</td>
<td>-1.2410</td>
<td></td>
</tr>
<tr>
<td>377.8300</td>
<td>-1.2480</td>
<td>393.6405</td>
<td>11.5207</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>387.7000</td>
<td>12.0000</td>
<td>381.1000</td>
<td>12.0000</td>
<td>308.8000</td>
<td>16.3000</td>
<td></td>
</tr>
<tr>
<td>298.0000</td>
<td>17.0000</td>
<td>281.0848</td>
<td>17.3600</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

### Polygon 16 Outline

<table>
<thead>
<tr>
<th></th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>240.1000</td>
<td>2.1500</td>
<td>242.4000</td>
<td>1.6200</td>
<td>251.7000</td>
<td>1.4700</td>
<td></td>
</tr>
<tr>
<td>258.1000</td>
<td>1.4200</td>
<td>260.3000</td>
<td>0.8300</td>
<td>267.1000</td>
<td>0.8200</td>
<td></td>
</tr>
<tr>
<td>270.7000</td>
<td>2.5100</td>
<td>265.9000</td>
<td>2.6500</td>
<td>263.3000</td>
<td>3.2000</td>
<td></td>
</tr>
<tr>
<td>259.9000</td>
<td>3.6400</td>
<td>239.8000</td>
<td>3.6400</td>
<td>238.4000</td>
<td>3.9400</td>
<td></td>
</tr>
</tbody>
</table>

### Polygon 17 Outline

<table>
<thead>
<tr>
<th></th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>151.1700</td>
<td>36.8700</td>
<td>201.5600</td>
<td>30.2200</td>
<td>251.9400</td>
<td>28.1500</td>
<td></td>
</tr>
<tr>
<td>302.3000</td>
<td>34.2400</td>
<td>352.7200</td>
<td>34.7500</td>
<td>403.1100</td>
<td>36.7300</td>
<td></td>
</tr>
<tr>
<td>453.5000</td>
<td>34.0000</td>
<td>498.6625</td>
<td>34.0000</td>
<td>10000.0000</td>
<td>30.0000</td>
<td></td>
</tr>
<tr>
<td>10000.0000</td>
<td>60.0000</td>
<td>210.6448</td>
<td>60.0000</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**Calculated Gravity Anomaly (gravity units)**

<table>
<thead>
<tr>
<th></th>
<th>345.6741</th>
<th>330.2186</th>
<th>319.1494</th>
<th>308.8754</th>
<th>261.4865</th>
</tr>
</thead>
<tbody>
<tr>
<td>244.3926</td>
<td>227.4126</td>
<td>215.2665</td>
<td>203.6124</td>
<td>184.4918</td>
<td>168.1382</td>
</tr>
<tr>
<td>156.2719</td>
<td>140.0887</td>
<td>130.5944</td>
<td>113.6550</td>
<td>98.0482</td>
<td>81.7581</td>
</tr>
<tr>
<td>65.9911</td>
<td>38.9596</td>
<td>10.7137</td>
<td>-7.0695</td>
<td>-20.0722</td>
<td>-29.5702</td>
</tr>
<tr>
<td>-42.0227</td>
<td>-59.6826</td>
<td>-63.9711</td>
<td>-67.8556</td>
<td>-58.1185</td>
<td>-49.5226</td>
</tr>
<tr>
<td>-36.9199</td>
<td>-5.3271</td>
<td>-48.9035</td>
<td>-139.7169</td>
<td>230.8038</td>
<td>246.5868</td>
</tr>
<tr>
<td>351.3491</td>
<td>359.4241</td>
<td>362.5869</td>
<td>329.3145</td>
<td>366.6794</td>
<td>130.6704</td>
</tr>
<tr>
<td>-78.2788</td>
<td>-148.4629</td>
<td>-221.2905</td>
<td>-222.0046</td>
<td>-245.1843</td>
<td>-253.7849</td>
</tr>
<tr>
<td>-240.2231</td>
<td>-201.8855</td>
<td>-146.8677</td>
<td>-91.5693</td>
<td>-40.1895</td>
<td>14.5273</td>
</tr>
<tr>
<td>46.5703</td>
<td>63.2114</td>
<td>79.4907</td>
<td>85.1265</td>
<td>4.2173</td>
<td>2.5801</td>
</tr>
<tr>
<td>40.7603</td>
<td>75.1973</td>
<td>153.3413</td>
<td>179.8823</td>
<td>248.3281</td>
<td>296.2437</td>
</tr>
<tr>
<td>325.8320</td>
<td>335.2959</td>
<td>334.6079</td>
<td>331.0034</td>
<td>323.2485</td>
<td>320.3833</td>
</tr>
<tr>
<td>293.0576</td>
<td>236.0322</td>
<td>169.0737</td>
<td>148.9033</td>
<td>145.6719</td>
<td>139.2339</td>
</tr>
<tr>
<td>119.0811</td>
<td>108.1343</td>
<td>105.1982</td>
<td>105.4214</td>
<td>141.5679</td>
<td>188.3159</td>
</tr>
</tbody>
</table>
### Appendix D

#### GRAVMAG MODEL LISTINGS

<table>
<thead>
<tr>
<th>Value 1</th>
<th>Value 2</th>
<th>Value 3</th>
<th>Value 4</th>
<th>Value 5</th>
<th>Value 6</th>
</tr>
</thead>
<tbody>
<tr>
<td>224.0010</td>
<td>234.5737</td>
<td>276.1616</td>
<td>312.5835</td>
<td>314.9111</td>
<td>318.3711</td>
</tr>
<tr>
<td>318.8940</td>
<td>324.4668</td>
<td>328.4819</td>
<td>329.7588</td>
<td>334.9829</td>
<td>334.6216</td>
</tr>
<tr>
<td>332.8325</td>
<td>307.3062</td>
<td>211.4912</td>
<td>161.2085</td>
<td>127.8525</td>
<td>115.6934</td>
</tr>
<tr>
<td>108.3218</td>
<td>107.7710</td>
<td>111.1885</td>
<td>115.7065</td>
<td>131.4595</td>
<td>160.9707</td>
</tr>
<tr>
<td>205.6299</td>
<td>249.5015</td>
<td>289.6899</td>
<td>321.8179</td>
<td>340.6750</td>
<td>342.7500</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Value 1</th>
<th>Value 2</th>
<th>Value 3</th>
<th>Value 4</th>
<th>Value 5</th>
<th>Value 6</th>
</tr>
</thead>
<tbody>
<tr>
<td>Observed Gravity Anomaly (gravity units)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>414.4050</td>
<td>416.5210</td>
<td>461.3200</td>
<td>485.8945</td>
<td>501.8478</td>
<td>276.9558</td>
</tr>
<tr>
<td>282.5298</td>
<td>268.6792</td>
<td>228.1145</td>
<td>263.5272</td>
<td>271.6047</td>
<td>241.7880</td>
</tr>
<tr>
<td>229.1805</td>
<td>206.3660</td>
<td>201.8733</td>
<td>66.9857</td>
<td>29.1903</td>
<td>-28.6985</td>
</tr>
<tr>
<td>-113.3850</td>
<td>-201.6000</td>
<td>-154.8600</td>
<td>-154.8361</td>
<td>-71.0490</td>
<td>-60.0750</td>
</tr>
<tr>
<td>-18.6665</td>
<td>27.0225</td>
<td>77.0305</td>
<td>129.3075</td>
<td>178.8195</td>
<td>203.1000</td>
</tr>
<tr>
<td>340.6750</td>
<td>342.7500</td>
<td>325.0000</td>
<td>325.0750</td>
<td>322.4000</td>
<td>187.1125</td>
</tr>
<tr>
<td>-1.4500</td>
<td>-142.7750</td>
<td>-229.2625</td>
<td>-265.0750</td>
<td>-281.0625</td>
<td>-278.7250</td>
</tr>
<tr>
<td>-248.0150</td>
<td>-205.6548</td>
<td>-125.0768</td>
<td>-84.2643</td>
<td>-52.6746</td>
<td>49.1715</td>
</tr>
<tr>
<td>101.2118</td>
<td>148.4313</td>
<td>156.5453</td>
<td>141.0780</td>
<td>96.2808</td>
<td>60.6944</td>
</tr>
<tr>
<td>10.5155</td>
<td>32.0265</td>
<td>156.9363</td>
<td>272.0480</td>
<td>316.9965</td>
<td>318.2610</td>
</tr>
<tr>
<td>325.0291</td>
<td>334.9500</td>
<td>315.2465</td>
<td>305.5659</td>
<td>200.2476</td>
<td>157.4550</td>
</tr>
<tr>
<td>108.6010</td>
<td>74.9191</td>
<td>88.3180</td>
<td>128.5181</td>
<td>128.3685</td>
<td>137.4440</td>
</tr>
<tr>
<td>148.2410</td>
<td>134.7195</td>
<td>102.0950</td>
<td>84.9766</td>
<td>96.4924</td>
<td>142.1736</td>
</tr>
<tr>
<td>226.0568</td>
<td>229.3136</td>
<td>247.9460</td>
<td>282.9034</td>
<td>320.7738</td>
<td>378.0613</td>
</tr>
<tr>
<td>390.1238</td>
<td>411.8790</td>
<td>388.5940</td>
<td>371.8518</td>
<td>347.8140</td>
<td>341.7192</td>
</tr>
<tr>
<td>314.9642</td>
<td>280.8143</td>
<td>202.6643</td>
<td>142.9342</td>
<td>113.4640</td>
<td>93.5520</td>
</tr>
<tr>
<td>62.4640</td>
<td>80.0320</td>
<td>84.2321</td>
<td>68.9600</td>
<td>88.5599</td>
<td>59.7520</td>
</tr>
<tr>
<td>150.6887</td>
<td>242.5062</td>
<td>370.7015</td>
<td>412.5563</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
### Appendix D

**GRAVMAG MODEL LISTINGS**

**Model Title:** KRISP 90 EW Line (Improved Model - A)

**Profile Orientation:** .00000 Degrees

**Earth's Magnetic Field**

<table>
<thead>
<tr>
<th>Declination</th>
<th>Inclination</th>
</tr>
</thead>
<tbody>
<tr>
<td>-7.000</td>
<td>68.000</td>
</tr>
</tbody>
</table>

**Gravity Calculation Surface:** .000 Km Above Observations

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>Background</td>
<td>2.7000</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>2.1000</td>
<td>.0000</td>
<td>-7.000</td>
<td>68.000</td>
<td>10.00</td>
</tr>
<tr>
<td>2</td>
<td>2.1000</td>
<td>.0000</td>
<td>-7.000</td>
<td>68.000</td>
<td>10.00</td>
</tr>
<tr>
<td>3</td>
<td>2.3000</td>
<td>.0000</td>
<td>-7.000</td>
<td>68.000</td>
<td>10.00</td>
</tr>
<tr>
<td>4</td>
<td>2.4600</td>
<td>.0000</td>
<td>-7.000</td>
<td>68.000</td>
<td>100.00</td>
</tr>
<tr>
<td>5</td>
<td>2.5000</td>
<td>.0000</td>
<td>-7.000</td>
<td>68.000</td>
<td>100.00</td>
</tr>
<tr>
<td>6</td>
<td>2.6000</td>
<td>.0000</td>
<td>-7.000</td>
<td>68.000</td>
<td>100.00</td>
</tr>
<tr>
<td>7</td>
<td>2.8000</td>
<td>.0000</td>
<td>-7.000</td>
<td>68.000</td>
<td>100.00</td>
</tr>
<tr>
<td>8</td>
<td>2.8000</td>
<td>.0000</td>
<td>-7.000</td>
<td>68.000</td>
<td>100.00</td>
</tr>
<tr>
<td>9</td>
<td>2.8600</td>
<td>.0000</td>
<td>-7.000</td>
<td>68.000</td>
<td>10000.00</td>
</tr>
<tr>
<td>10</td>
<td>2.9200</td>
<td>.0000</td>
<td>-7.000</td>
<td>68.000</td>
<td>10000.00</td>
</tr>
<tr>
<td>11</td>
<td>3.2500</td>
<td>.0000</td>
<td>-7.000</td>
<td>68.000</td>
<td>10000.00</td>
</tr>
<tr>
<td>12</td>
<td>0.9400</td>
<td>.0000</td>
<td>-7.000</td>
<td>68.000</td>
<td>10000.00</td>
</tr>
<tr>
<td>13</td>
<td>3.2000</td>
<td>.0000</td>
<td>-7.000</td>
<td>68.000</td>
<td>10000.00</td>
</tr>
<tr>
<td>14</td>
<td>2.8200</td>
<td>.0000</td>
<td>-7.000</td>
<td>68.000</td>
<td>10000.00</td>
</tr>
<tr>
<td>15</td>
<td>2.5500</td>
<td>.0000</td>
<td>-7.000</td>
<td>68.000</td>
<td>10000.00</td>
</tr>
<tr>
<td>16</td>
<td>2.7500</td>
<td>.0000</td>
<td>-7.000</td>
<td>68.000</td>
<td>10000.00</td>
</tr>
<tr>
<td>17</td>
<td>2.7500</td>
<td>.0000</td>
<td>-7.000</td>
<td>68.000</td>
<td>10000.00</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Polygon 2 Outline</th>
</tr>
</thead>
<tbody>
<tr>
<td>X</td>
</tr>
<tr>
<td>214.1000</td>
</tr>
<tr>
<td>215.8600</td>
</tr>
<tr>
<td>227.6500</td>
</tr>
<tr>
<td>240.5740</td>
</tr>
<tr>
<td>234.1000</td>
</tr>
<tr>
<td>218.9000</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Polygon 3 Outline</th>
</tr>
</thead>
<tbody>
<tr>
<td>X</td>
</tr>
<tr>
<td>264.3000</td>
</tr>
<tr>
<td>259.4800</td>
</tr>
<tr>
<td>265.0612</td>
</tr>
</tbody>
</table>
### Appendix D

#### GRAVMAG MODEL LISTINGS

**Polygon 4 Outline**

<table>
<thead>
<tr>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>181.2000</td>
<td>2.0900</td>
<td>180.6000</td>
<td>-0.7700</td>
<td>184.5000</td>
<td>-0.8000</td>
</tr>
<tr>
<td>185.6835</td>
<td>-1.1945</td>
<td>186.0500</td>
<td>-1.2150</td>
<td>187.6300</td>
<td>-1.3410</td>
</tr>
<tr>
<td>188.9800</td>
<td>-1.4510</td>
<td>191.8000</td>
<td>-1.6680</td>
<td>194.5700</td>
<td>-1.8950</td>
</tr>
<tr>
<td>197.2100</td>
<td>-1.9460</td>
<td>199.7100</td>
<td>-1.8850</td>
<td>202.9400</td>
<td>-1.8800</td>
</tr>
<tr>
<td>205.4200</td>
<td>-1.5960</td>
<td>206.5900</td>
<td>-1.3720</td>
<td>208.2500</td>
<td>-1.3180</td>
</tr>
<tr>
<td>210.5700</td>
<td>-1.2620</td>
<td>212.6400</td>
<td>-1.2250</td>
<td>213.3701</td>
<td>-1.2123</td>
</tr>
<tr>
<td>214.1000</td>
<td>-0.6900</td>
<td>234.1000</td>
<td>-1.8200</td>
<td>234.6000</td>
<td>-1.8490</td>
</tr>
<tr>
<td>238.5000</td>
<td>-0.5300</td>
<td>240.5740</td>
<td>-1.0263</td>
<td>241.5800</td>
<td>-1.0720</td>
</tr>
<tr>
<td>244.6200</td>
<td>-1.0140</td>
<td>251.0000</td>
<td>-1.0520</td>
<td>254.2700</td>
<td>-1.0590</td>
</tr>
<tr>
<td>256.6476</td>
<td>-1.1010</td>
<td>256.8000</td>
<td>-1.1000</td>
<td>264.3000</td>
<td>-0.8700</td>
</tr>
<tr>
<td>265.0612</td>
<td>-1.2337</td>
<td>266.7800</td>
<td>-1.3580</td>
<td>269.7600</td>
<td>-1.5900</td>
</tr>
<tr>
<td>274.3600</td>
<td>-1.7010</td>
<td>278.3600</td>
<td>-1.8500</td>
<td>280.8700</td>
<td>-1.8510</td>
</tr>
<tr>
<td>282.8800</td>
<td>-1.8450</td>
<td>285.8200</td>
<td>-1.9470</td>
<td>288.8900</td>
<td>-1.9470</td>
</tr>
<tr>
<td>296.9000</td>
<td>-1.8200</td>
<td>301.0300</td>
<td>-1.8370</td>
<td>307.3500</td>
<td>-1.7840</td>
</tr>
<tr>
<td>312.4600</td>
<td>-1.7910</td>
<td>320.4300</td>
<td>-1.7600</td>
<td>322.8600</td>
<td>-1.7230</td>
</tr>
<tr>
<td>330.5600</td>
<td>-1.7090</td>
<td>333.4000</td>
<td>-1.6794</td>
<td>333.4000</td>
<td>-1.6000</td>
</tr>
<tr>
<td>312.7000</td>
<td>-1.3400</td>
<td>311.7000</td>
<td>-1.0500</td>
<td>283.6000</td>
<td>-0.4600</td>
</tr>
<tr>
<td>273.3000</td>
<td>-0.1700</td>
<td>269.1000</td>
<td>-0.4500</td>
<td>267.1000</td>
<td>1.3100</td>
</tr>
<tr>
<td>312.3000</td>
<td>1.4700</td>
<td>312.7000</td>
<td>2.2500</td>
<td>209.0000</td>
<td>1.5700</td>
</tr>
</tbody>
</table>

**Polygon 5 Outline**

<table>
<thead>
<tr>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>181.2000</td>
<td>2.0900</td>
<td>186.3000</td>
<td>1.4700</td>
<td>195.7000</td>
<td>1.3100</td>
</tr>
<tr>
<td>209.0000</td>
<td>1.5700</td>
<td>212.9000</td>
<td>2.2500</td>
<td>224.0000</td>
<td>2.4500</td>
</tr>
<tr>
<td>240.1000</td>
<td>2.1500</td>
<td>240.1000</td>
<td>2.1500</td>
<td>224.0000</td>
<td>2.4500</td>
</tr>
<tr>
<td>212.9000</td>
<td>2.2500</td>
<td>224.0000</td>
<td>2.4500</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**Polygon 6 Outline**

<table>
<thead>
<tr>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>136.2000</td>
<td>-1.9300</td>
<td>132.2150</td>
<td>-1.9884</td>
<td>134.6700</td>
<td>-2.0700</td>
</tr>
<tr>
<td>139.0500</td>
<td>-2.1790</td>
<td>144.3300</td>
<td>-2.1910</td>
<td>150.7200</td>
<td>-2.2020</td>
</tr>
<tr>
<td>164.6600</td>
<td>-2.3870</td>
<td>167.0000</td>
<td>-2.4260</td>
<td>169.2000</td>
<td>-2.4860</td>
</tr>
<tr>
<td>174.7400</td>
<td>-2.6470</td>
<td>176.0800</td>
<td>-2.6760</td>
<td>177.6196</td>
<td>-2.1664</td>
</tr>
<tr>
<td>159.0000</td>
<td>-1.0900</td>
<td>154.9000</td>
<td>-0.5400</td>
<td>150.6000</td>
<td>-0.8400</td>
</tr>
<tr>
<td>146.6000</td>
<td>-1.8000</td>
<td>151.3570</td>
<td>16.4671</td>
<td>131.1732</td>
<td>1.7700</td>
</tr>
</tbody>
</table>

**Polygon 7 Outline**

<table>
<thead>
<tr>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>330.0000</td>
<td>1.4900</td>
<td>352.9000</td>
<td>4.4700</td>
<td>365.6000</td>
<td>5.3900</td>
</tr>
<tr>
<td>319.1000</td>
<td>5.3900</td>
<td>319.1000</td>
<td>5.3900</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**Polygon 8 Outline**

<table>
<thead>
<tr>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>151.3570</td>
<td>16.4671</td>
<td>131.1732</td>
<td>1.7700</td>
<td>110.9483</td>
<td>-1.8576</td>
</tr>
<tr>
<td>112.1500</td>
<td>-1.8740</td>
<td>115.9400</td>
<td>-1.9700</td>
<td>119.5000</td>
<td>-2.0030</td>
</tr>
<tr>
<td>124.5100</td>
<td>-1.9390</td>
<td>131.1500</td>
<td>-1.9530</td>
<td>131.4880</td>
<td>-1.9642</td>
</tr>
<tr>
<td>136.0000</td>
<td>-1.9300</td>
<td>146.6000</td>
<td>-1.8000</td>
<td>150.6000</td>
<td>-0.8400</td>
</tr>
</tbody>
</table>
Appendix D

GRAVMAG MODEL LISTINGS

<table>
<thead>
<tr>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
<th>X</th>
<th>Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>154.9000</td>
<td>-0.5400</td>
<td>159.0000</td>
<td>-1.0900</td>
<td>177.6196</td>
<td>-2.1664</td>
</tr>
<tr>
<td>180.3100</td>
<td>-1.2760</td>
<td>180.3573</td>
<td>-1.2735</td>
<td>180.6000</td>
<td>-0.7700</td>
</tr>
</tbody>
</table>

Polygon 9 Outline

<table>
<thead>
<tr>
<th>X Z X Z X Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>-10000.0000 25.8700 0.0000 25.9000 18.7000 25.2000</td>
</tr>
<tr>
<td>51.1000 24.6000 87.7000 25.6000 100.0000 27.9000</td>
</tr>
<tr>
<td>176.2000 27.7000 197.5000 27.8600 239.3000 23.7000</td>
</tr>
<tr>
<td>263.9000 23.6500 281.3000 26.8800 351.5000 29.4000</td>
</tr>
<tr>
<td>398.3000 29.4000 453.5000 29.5500 503.0913 29.5673</td>
</tr>
<tr>
<td>10000.0000 28.5000 10000.0000 30.0000 498.6625 33.5958</td>
</tr>
<tr>
<td>453.5000 33.5600 422.2569 37.1789 360.9360 35.3254</td>
</tr>
<tr>
<td>342.0287 35.0588 316.3000 33.8100 296.2000 34.0000</td>
</tr>
<tr>
<td>272.4000 33.4300 261.3000 31.7100 255.3000 30.3800</td>
</tr>
<tr>
<td>242.6000 29.7500 221.3000 29.7500 208.5000 30.3800</td>
</tr>
<tr>
<td>189.8000 31.1400 179.6000 32.4900 169.4000 35.4300</td>
</tr>
<tr>
<td>158.3000 36.8000 131.1000 37.6300 93.6000 37.6300</td>
</tr>
<tr>
<td>53.2000 35.2300 32.1000 34.1500 0.0000 34.1500</td>
</tr>
<tr>
<td>-10000.0000 34.1500</td>
</tr>
</tbody>
</table>

Polygon 10 Outline

<table>
<thead>
<tr>
<th>X Z X Z X Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>-10000.0000 34.1500 0.0000 34.1500 32.1000 34.1500</td>
</tr>
<tr>
<td>53.2000 35.2300 93.6000 37.6300 131.1000 37.6300</td>
</tr>
<tr>
<td>158.3000 36.8000 169.4000 35.4300 179.6000 32.4900</td>
</tr>
<tr>
<td>189.8000 31.1400 208.5000 30.3800 221.3000 29.7500</td>
</tr>
<tr>
<td>242.6000 29.7500 261.3000 31.7100 255.3000 30.3800</td>
</tr>
<tr>
<td>272.4000 33.4300 221.3000 34.0000 316.3000 33.8100</td>
</tr>
<tr>
<td>342.0287 35.0588 360.9360 35.3254 422.2569 37.1789</td>
</tr>
<tr>
<td>453.5000 33.5600 498.6625 33.5958 10000.0000 30.0000</td>
</tr>
<tr>
<td>10000.0000 60.0000 -10000.0000 60.0000</td>
</tr>
</tbody>
</table>

Polygon 11 Outline

<table>
<thead>
<tr>
<th>X Z X Z X Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>-10000.0000 60.0000 10000.0000 60.0000 10000.0000 70.0000</td>
</tr>
<tr>
<td>-10000.0000 70.0000</td>
</tr>
</tbody>
</table>
### Appendix D

<table>
<thead>
<tr>
<th>Polygon 13 Outline</th>
<th>GRAVMAG MODEL LISTINGS</th>
</tr>
</thead>
<tbody>
<tr>
<td>X Z X Z X Z X Z X Z</td>
<td>Calculated Gravity Anomaly (gravity units)</td>
</tr>
<tr>
<td>169.4000 35.4300</td>
<td>188.9123 186.1282 184.3715 181.7555 173.7419 165.8717</td>
</tr>
<tr>
<td>208.3000 30.3800</td>
<td>158.9470 150.3773 143.5016 136.1891 122.8512 109.5138</td>
</tr>
<tr>
<td>255.3000 30.3800</td>
<td>99.4510 83.4202 73.5821 55.4212 36.6803 16.0461</td>
</tr>
<tr>
<td>296.2000 34.0000</td>
<td>-4.7284 -44.0905 -91.9445 -122.4748 -144.6974 -161.0973</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Polygon 14 Outline</th>
<th>X Z X Z X Z X Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>193.2000 8.8000</td>
<td>267.1000 0.8200 269.1000 -0.0500 273.5000 -1.0500</td>
</tr>
<tr>
<td>208.4000 3.5300</td>
<td>283.6000 -0.4600 311.7000 -2.9300 311.9000 -1.6000</td>
</tr>
<tr>
<td>236.1000 4.1800</td>
<td>333.4000 -1.6000 326.7000 -2.9300 311.9000 -1.6000</td>
</tr>
<tr>
<td>259.9000 3.6400</td>
<td>310.5000 -0.3300 306.0000 -2.0000 302.3000 0.2400</td>
</tr>
<tr>
<td>270.7000 2.5100</td>
<td>283.3000 0.5700 278.5000 0.7800 275.9000 1.4200</td>
</tr>
<tr>
<td>253.7000 14.2000</td>
<td>275.5000 1.9300 270.7000 2.5100</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Polygon 15 Outline</th>
<th>X Z X Z X Z X Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>208.4000 3.5300</td>
<td>270.7000 2.5100 275.5000 1.9300 275.9000 1.4200</td>
</tr>
<tr>
<td>236.1000 4.1800</td>
<td>278.5000 0.7800 283.3000 0.5700 302.3000 0.2400</td>
</tr>
<tr>
<td>259.9000 3.6400</td>
<td>305.6000 -0.2400 333.4000 -1.6000 336.6803 -1.5807</td>
</tr>
<tr>
<td>270.7000 2.5100</td>
<td>338.5000 -1.4600 342.0500 -1.4280 344.5500 -1.2850</td>
</tr>
<tr>
<td>191.7000 16.5200</td>
<td>360.6400 -1.2790 365.6300 -1.1660 371.8300 -1.2410</td>
</tr>
<tr>
<td>191.7000 16.5200</td>
<td>387.7000 11.4300 381.1000 13.0600 308.8000 13.0600</td>
</tr>
<tr>
<td>298.0000 15.7700</td>
<td>270.7000 2.5100 275.5000 1.9300 275.9000 1.4200</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Polygon 16 Outline</th>
<th>X Z X Z X Z X Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>208.4000 3.5300</td>
<td>240.1000 2.1500 242.4000 1.6200 251.7000 1.4700</td>
</tr>
<tr>
<td>236.1000 4.1800</td>
<td>258.1000 1.4200 260.3000 0.8300 267.1000 0.8200</td>
</tr>
<tr>
<td>259.9000 3.6400</td>
<td>270.7000 2.5100 265.9000 2.6500 263.3000 3.2000</td>
</tr>
</tbody>
</table>
### Appendix D

#### GRAVMAG Model Listings

<p>| | | | | | |</p>
<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>-179.6531</td>
<td>-200.2492</td>
<td>-203.5461</td>
<td>-194.5074</td>
<td>-161.1002</td>
<td>-130.1561</td>
</tr>
<tr>
<td>-84.5485</td>
<td>9.0234</td>
<td>67.6754</td>
<td>133.2498</td>
<td>204.7967</td>
<td>213.6495</td>
</tr>
<tr>
<td>335.8849</td>
<td>348.7892</td>
<td>355.9307</td>
<td>332.1766</td>
<td>311.5305</td>
<td>141.0275</td>
</tr>
<tr>
<td>-227.6986</td>
<td>-189.1313</td>
<td>-133.5321</td>
<td>-77.3015</td>
<td>-23.9071</td>
<td>31.9066</td>
</tr>
<tr>
<td>65.1034</td>
<td>82.6333</td>
<td>100.7138</td>
<td>107.5310</td>
<td>27.5492</td>
<td>26.2272</td>
</tr>
<tr>
<td>65.0517</td>
<td>98.4196</td>
<td>173.5466</td>
<td>196.9664</td>
<td>254.7320</td>
<td>299.2047</td>
</tr>
<tr>
<td>324.0331</td>
<td>325.3175</td>
<td>317.7998</td>
<td>306.4706</td>
<td>292.6163</td>
<td>287.8452</td>
</tr>
<tr>
<td>259.0713</td>
<td>198.8743</td>
<td>126.6878</td>
<td>102.1106</td>
<td>97.8516</td>
<td>91.3275</td>
</tr>
<tr>
<td>72.6868</td>
<td>64.6828</td>
<td>73.3077</td>
<td>80.8481</td>
<td>126.2913</td>
<td>178.2850</td>
</tr>
<tr>
<td>225.0304</td>
<td>240.7814</td>
<td>299.4998</td>
<td>334.3759</td>
<td>334.6735</td>
<td>332.1891</td>
</tr>
<tr>
<td>329.2498</td>
<td>321.9079</td>
<td>312.7416</td>
<td>299.7560</td>
<td>282.5500</td>
<td>263.2021</td>
</tr>
<tr>
<td>250.0410</td>
<td>226.7969</td>
<td>155.0550</td>
<td>120.4370</td>
<td>97.9367</td>
<td>90.1193</td>
</tr>
<tr>
<td>85.3613</td>
<td>86.1087</td>
<td>89.0115</td>
<td>92.8544</td>
<td>108.2416</td>
<td>141.4989</td>
</tr>
<tr>
<td>197.5869</td>
<td>255.3848</td>
<td>308.7442</td>
<td>349.1072</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

#### Observed Gravity Anomaly (gravity units)

| 414.4050 | 415.5210 | 466.6060 | 483.6945 | 501.8478 | 276.9558 |
| 282.5298 | 268.6792 | 228.1145 | 263.5272 | 271.6047 | 241.7880 |
| 229.1805 | 206.3600 | 204.8733 | 66.9985  | 29.1903  | -28.6095 |
| -63.0000 | -131.4351 | -132.1450 | -144.2400 | -107.1050 | -109.5601 |
| -113.3850 | -201.6600 | -154.8600 | -154.8361 | -71.0490  | -60.0750  |
| 18.6665  | 27.0225  | 77.0305  | 129.3075 | 178.8195 | 203.1000 |
| 340.6750  | 342.7500 | 325.0000 | 325.0750 | 322.4000 | 187.1125 |
| -1.4500  | -142.7750 | -229.2625 | -265.0750 | -281.0625 | -275.7250 |
| -248.0150 | -205.6548 | -125.0768 | -84.2643 | -52.6746 | 49.1715  |
| 101.2118 | 148.4313 | 156.5453 | 141.0780 | 96.2808  | 60.6844  |
| 10.5155  | 32.0265  | 156.9363 | 272.0480 | 316.9965 | 318.2610 |
| 325.0291  | 334.9500 | 315.2465 | 305.5659 | 299.2476 | 157.4550 |
| 108.6010  | 74.4919  | 88.3180  | 128.5181 | 128.3685 | 137.4440 |
| 148.2410  | 134.7195 | 102.0950 | 84.9766  | 96.4924  | 142.1736 |
| 226.0568  | 229.3156 | 247.9460 | 282.9034 | 320.7738 | 378.0613 |
| 390.1258  | 411.8790 | 388.5940 | 371.8518 | 347.8140 | 341.7192 |
| 314.9642  | 280.8143 | 202.6643 | 142.9342 | 113.4640 | 93.5520  |
| 62.4640   | 80.0320  | 84.2321  | 68.9600  | 88.5599  | 59.7520  |
| 150.6887  | 242.5662 | 370.7015 | 412.5563 |
On the Upper Mantle Beneath the Kenya Rift

R. Masotti¹, P.K.H. Maguire¹ and J. Mechie²*

1 Department of Geology, University of Leicester, England
2 Geophysikalisches Institut, Universität Karlsruhe, Germany

March 13, 1995

Abstract

During the Kenya Rift International Seismic Project (KRISP 90) a 450 km long E-W seismic refraction/wide angle reflection profile involving the deployment of 250 instruments was shot across the Kenya Rift. A reflected phase recorded between distances of 260 and 350 km from a 1000 kg shot at the western end of the line, in Lake Victoria has been interpreted as originating from about 60 km beneath the rift's western margin.

Detailed processing of this phase has resulted in defining its polarity in relation to the first arrival diving wave at the same range. Extensive kinematic and dynamic modelling suggests there is a high velocity zone at depth below 60 km under the western flank of the rift. Moreover, we can not exclude the presence of a layered alternating high-low velocity structure as found in the upper mantle beneath the northern part of the N-S seismic profile along the rift axis.

Constraints from xenolith studies suggest anisotropy may be required to explain the high velocity found beneath the reflecting horizon (> 8.40 km/s). Petrological modelling suggests the anisotropy should be due to the preferred orientation of olivine crystals, with either a transverse isotropic structure in which the 'a' and 'c' axes are randomly orientated in the horizontal plane or an orthorhombic structure in which the fast 'a' axis is orientated along the direction of the E-W seismic line being possible. The reflection could also be caused by a pre-rift structure associated with the Proterozoic collisional orogen involving the Mozambique Orogenic Belt and the Archaean Nyasza Craton, whose contact subparallels and lies c.70 km to the west of the Tertiary Rift. The evidence presented here may delimit the lateral extent of the upper mantle region of anomalously low velocity material which seems to be confined to occur below the surface expression of the rift itself at depths below 60 km.

Key words: Kenya Rift, Upper mantle reflection, Modelling.

¹now at GeoForschungszentrum Potsdam, Germany
1 Introduction

One of the objectives of KRISP 90 was to detail the upper mantle structure beneath the Kenya Rift Valley, where the rifting process is still active (Fig. 1).

It is generally accepted that rifting episodes are associated with magmatic activity whose origin resides in the lithospheric upper mantle or in the asthenosphere. Previous geophysical studies in Kenya have demonstrated large gravity and heat flow anomalies indicating the presence of anomalous mantle material or partial melt beneath the rift axis. A structural high has been modelled in the upper mantle, with the Moho rising from 38 km underneath the rift flanks to 30 km beneath the rift axis (Maguire et al., 1994). A close relationship has been suggested between asthenospheric upwelling and penetrative magmatism as driving forces for lithospheric rifting (Bhattacharji et al., 1987).

This paper describes and interprets a phase 'd1', reflected from a discontinuity below the Moho recorded during KRISP 90 on the cross-rift line (Fig. 1) on the section from the Lake Victoria shot point (VIC) at the western end of the profile. Discontinuities within the upper mantle in rift zones have also been identified elsewhere, for example in the Afar depression and the Red Sea (Makris and Ginzburg, 1987; Mooney et al., 1985; Prodehl, 1985; Mechie et al., 1986).

The $d_1$ phase has been the subject of further processing and analysis to define its polarity, in order to determine whether a low velocity zone (LVZ) exists in the upper mantle beneath the reflector. Study of lithosphere structure elsewhere (Hirn et al., 1971; Kind, 1974; Cassell and Fuchs, 1979; Burmakov et al., 1987; Ansoige et al., 1992) has revealed thin layers of both high and low velocity explained either via changes in composition (Blundell, 1992) or physical state. The interpretation of the KRISP 90 axial line has revealed a low velocity layer between two high velocity layers at a depth between 50 to 65 km beneath the northern-central part of the profile (Keller et al., 1994) and for this reason it was important to study the polarity of phase $d_1$.

If we recognise that this phase has the same polarity as the first 'd' arrival, being the diving wave beneath the Moho, we can interpret such polarity as resulting from an increase in velocity across the reflector. The results may be compared with those from a number of long profiles within rifts which have provided information from reflectors below the Moho (Keller et al., 1994).

The seismic velocities derived from the present analysis have been compared with values determined from petrological modelling. The elastic properties of individual minerals can be used to define the seismic velocities at a certain pressure and temperature for a given
Appendix E

assemblage of these minerals. Mineralogical compositions are derived from xenolith studies (Henjes-Kunst and Altherr, 1992), and a possible geotherm for the continental crust is based on a conductive model after Pollack and Chapman (1977).

2 The seismic data

The data studied have been recorded along the cross-rift line (Fig. 1), a 450 km profile between the equator and 1°N, from the 1000 kg underwater Lake Victoria (VIC) shot at the western end of the line. A detailed description of the interpretation of the cross-rift line can be found in Maguire et al. (1994) which also contains a complete record section for the shot point VIC and from which the present crustal model is obtained. The upper mantle reflection \( d_1 \) is shown in two 'windows' extracted from the VIC record section. The seismograms are trace normalised, band-pass filtered (1-20 Hz) and plotted using RSEC90 (Luetgert, 1988) (Fig. 2 and 3).

3 Methods

The investigated phase is observed at a distance of 260 to 350 km from Lake Victoria. The data quality required detailed processing to be undertaken before enabling comparison of the polarity of the reflection to the sub-Moho diving wave at the same offset. Across the reflector a decrease in velocity would produce a reflection of opposite polarity to the first arrival. With a velocity increase then the polarity should be the same as that of the first arrival as long as the reflection is still subcritical. Beyond the critical point the phase of the reflection starts to change and eventually at large distances it will again have opposite polarity to the first arrival. The distance over which the phase change occurs depends on the velocity contrast. For large velocity contrasts it changes over a fairly short range, whereas for mantle reflected events involving small velocity contrasts it usually occurs over a large distance.

4 Procedure

4.1 Processing

RSEC90 (Luetgert, 1988) enables band-pass frequency filtering to improve the quality of the recorded seismograms. This was designed to enhance the crustal phases. The \( d_1 \) phase was first detected on these trace normalised band-pass filtered sections and together with the
4.2 CANC filtering

CANC was first developed by Hattingh in 1986 (Hattingh, 1988) as an extension of Adaptive Noise Cancelling (Widrow and Hoff, 1960; Widrow et al., 1975, 1976). CANC was initially designed to remove interference noise, random noise, and linear or non-linear drift from a time series using adaptive filters instead of fixed filters.

Fig. 4 (modified from Hattingh, 1988) illustrates the CANC procedure. The primary input \( D \) contains the signal \( S_o \) and the noise \( n_o \). The reference input \( X \) contains the same signal \( S_i \) and noise \( n_i \), which is assumed uncorrelated with \( n_o \). The reference input is adaptively filtered to produce the filter output \( Y \), which is a least-squares estimate of \( D = S_o + n_o \). The filter output is subtracted from the primary input to give the system output or error \( E \). The error is then fed back into the adaptive filter and used to adjust the filter weights. Since \( n_o \) is uncorrelated with the filter input \( X = S_i + n_i \), the filter output is a least-squares estimate of the signal \( S \). (Kirk, 1989).

The filter requires various input parameters. Following Hattingh (1988) and Kirk (1989):

- \( L \), the filter length. This should be set to twice the ratio of the signal bandwidth to frequency resolution of the filter. Values from 25 to 38 samples were used following spectral analysis of the signal.
- \( \mu \), the convergence rate. Long filters with small convergence rate provide the greatest signal resolution. Values were set between 0.001 and 0.0001.
- \( I \), the number of iterations. The CPU time required increases with increase in \( I \). Between 20 and 40 iterations were sufficient.

CANC filtering of wide angle seismic data is ideal for enhancing separate signals from one shot point recorded at one recording site. In the present case with signals recorded at a large distance from a single shot we are assuming correlation between the recorded phase at two adjacent recorders (spacing 0.3 - 3.8 km) but with uncorrelated noise.

CANC filtering can perform satisfactorily if there is a small phase shift between reference and input signal (Kirk, 1989). As Kirk (1989) suggests from synthetic data tests, the noisier signal should be on the primary input. Referring to Fig. 4 it is possible to see that the reference input is adaptively filtered to become similar to the primary input. The system output power will be high if the noise power on the primary input signal is high.

Where possible, the signal closest to the shot was used as the reference signal. For very
noisy signals, the reference and input signal were reversed to improve the output quality.

We analysed the \( d \) phase within a 1 second (200 samples) time window starting at 6 sec reduced time (for a reducing velocity of 8.0 km/s). The \( d_1 \) phase was analysed within a window of 2 seconds length.

The results are presented for selected traces for ranges between 260 and 350 km from the VIC shot point.

The derived cross-correlation function between the time windowed CANC filtered traces provides an estimate of the necessary time shift to achieve the greatest similarity. The accepted time shift was always less than a wavelet cycle. Each CANC filtered pair of traces has been cross-correlated with the input traces and only those with a cross-correlation > 60% were accepted.

Following CANC filtering and the subsequent time shifting, the best data sets were stacked. An example of the filtered and resultant stacked seismogram is shown for location numbers 31 to 42 for the \( d_1 \) phase in Fig. 5a and for the same range of location numbers for the \( d \) phase in Fig. 5b. The stacked seismogram for each group of traces is shown at the right hand side of each figure.

For every location number group the final output shows a \( d_1 \) phase with the same polarity as the first arrival \( d \). In order to quantify the similarity in each group between the \( d \) and the \( d_1 \) phases, cross-correlation has been done between \( d \) and \( d_1 \) as well as between \( d \) and an inverted \( d_1 \) phase for each group of location numbers. An example (location numbers 27 to 42 for \( d \) and 27 to 45 for \( d_1 \)) is shown in Fig. 6.

The measures of similarity suggest that the optimum result is achieved when cross-correlating the \( d \) phase with the \( d_1 \) phase rather than the inverted \( d_1 \) phase i.e. the polarity of the two phases may be taken to be the same, and the \( d_1 \) phase results from an increase in acoustic impedance in the uppermost mantle.

5 Interpretation

5.1 Ray tracing

Prior to the polarity analysis, Maguire et al. (1994) modelled the reflector \( d_1 \) via ray-tracing. They required a reflector at a depth of 55 km beneath the western margin of the rift (between about 140 to 160 km from VIC) to generate the high amplitude phase \( d_1 \) immediately after phase \( d \) on the VIC record section. A reflector at a depth of about 50 km is also identified beneath the flank line (Prodehl et al., 1994).
Maguire et al. (1994) arbitrarily suggested the reflection resulted from a positive impedance contrast, consistent with the present flank profile model. Beneath the axis of the rift the anomalously low velocity material immediately beneath the Moho was modelled extending to a depth of at least 60 km. This is consistent with the axial profile model which identified a boundary with a small impedance contrast at this depth south of Lake Baringo (Keller et al., 1994). It is also consistent with the recent model derived from teleseismic data (Green et al., 1991). On further detailed examination of the VIC data set we are not prepared to reject the hypothesis that the reflector identified beneath the flank of the rift does not continue beneath the rift. This reflector would be equivalent to the one modelled at about 60 km depth along the axial line (Keller et al., 1994).

The present new evidence increases confidence in the picking of both $d$ and $d_1$. $d_1$ appears to occur at about 7.5 sec reduced travel time, ($V_R = 8.0$ km/s) between 280 and at least 350 km from VIC (Fig. 2b). The $d_1$ picks obtained in the present study are somewhat later than those obtained in Maguire et al. (1994) which has necessitated the $d_1$ reflector being placed at greater depth.

The revised picked phases $d$ and $d_1$ were again modelled using the MacRay package (Luegert, 1992) based on the ray-tracing algorithm of Červený et al. (1977). The velocity structure above 55 km depth has not been changed from the final model of Maguire et al. (1994). Perturbation of the models has shown that once the phase correlation have been fixed, resolution of the relevant velocity and depth to interface may be accepted as better than 1% and 5% respectively.

Two different models are kinematically acceptable; one with and one without the presence of a LVZ (Fig. 7 a, b). In the former case the reflection arises from the base of the LVZ.

At about 350 km distance the travel-time difference between the $d$ and $d_1$ phases is about 0.5 sec and the two phases have very similar apparent velocities. For this reason and in order to move the theoretical critical point to around 280 km distance, a LVZ was introduced. In this case the observed travel times and amplitudes can be matched using a velocity of 8.4 km/s under the reflector together with a 4 km thick LVZ between 55 and 59 km depth and an average velocity in the LVZ of 7.7 km/s. The velocity in the LVZ is taken from the axial line model.

However, the observed travel times and amplitudes can also be fitted within the error limits without the necessity of a LVZ. In this case a velocity of 8.65 km/s beneath the reflector which should now be at 63 km depth, is required. The resultant theoretical critical point is at 304 km distance from VIC. Correcting for the sphericity of the earth (Mereu,
1967) the velocity below the reflector decreases to 8.56 km/s.

Both models fit the travel times equally well. In both cases the ray coverage on the reflector is limited to the distance range 100-180 km from VIC i.e. to the west of the surface expression of the rift. In the next section we will use the amplitude ratios between the $d$ and $d_1$ phases to try to discriminate between the two models.

5.2 Synthetic seismogram modelling

We have computed synthetic seismograms for the two seismic models using two methods: the ray-method (Cerveny et al., 1977) and finite-difference modelling (Kelly et al., 1976; Sandmeier, 1990), only the latter being discussed here. The synthetic record section in Fig. 8 was computed by a finite-difference technique for 2-D heterogeneous elastic media, which enables a complete description of the seismic wavefield. Finite-difference synthetic seismogram modelling enables a complete response of the model to the input source function. It can automatically account for the relative amplitudes of the various arrivals and includes contributions from converted waves, Rayleigh waves, diffractions from fault zones, and head waves. It is of course subject to limitations, e.g. model resolution and grid spacing. It can enable the matching of calculated to observed times in regions where 2-D kinematic ray-tracing breaks down. ‘The ray method, of course, cannot compete in the accuracy of computations with the reflectivity method’ (Červený, 1985), and the finite-difference technique may be considered as a 2-D reflectivity method.

We digitized our ray-traced model with a grid spacing of 40 m for the calculations. The dominant frequency for which we calculated the seismograms was 4 Hz, according to the stability conditions of the finite difference modelling (Kelly et al., 1976) and the analysis of the power spectra for the mantle phases. This analysis showed that relatively high frequencies (up to 10 Hz) are preserved beyond the anomalous upper mantle, which has no strong damping effect on the P-waves.

5.3 Amplitude ratio versus offset calculation

The maximum and minimum values of the observed amplitudes were automatically measured for the chosen phase within a selected time window. Due to considerable scatter of the observed amplitude data, probably caused by varying site response at the recording stations, we preferred, instead of ‘true’ amplitudes, to use amplitude ratios of the two phases ($d_1/d$).

A comparison of the amplitude ratios of the synthetic seismograms using the two velocity functions (with and without the LVZ) shows that the model with the LVZ produces amplitude
ratios a bit higher than the observed ones while the model without the LVZ produces values somewhat too low before 280 km distance. The \( d \) and \( d_1 \) phases apparently merge beyond 330 km from VIC on the synthetic section. It should be noted that these amplitude ratios have been derived using isotropic layered models. If, as is proposed below, the high velocity layer at about 60 km depth is anisotropic with preferred orientation of olivine crystals then the amplitude ratios shown here may be up to 20\% in error in the undercritical range and up to 5\% in error in the overcritical range.

Thus the two models fit the observed amplitude ratios with more or less equal quality (Fig. 9). Considering the difficulty in defining the exact position of the critical point for the \( d_1 \) phase the calculated amplitude ratios lie within the scatter of the observed ratios. For the finite-difference model without the LVZ the critical point is at about 300 km from VIC, for the one with the LVZ it is at about 280 km from VIC.

An eastward dip on the reflector of 5\° would move the critical point by no more than 30 km, still achieving a satisfactory travel time fit. It is apparent that a range of models between the two end members presented can be considered to fit the travel time and amplitude data. Appropriate variations in the thickness of the LVZ, depth of the interfaces, and seismic velocities may achieve the desired fit.

Although it is therefore not possible to exclude the presence of a LVZ, the seismic modelling does constrain the velocity underneath the reflector at about 8.4 ± 0.1 km/s, at a depth of 59.0 ± 2.0 km in the model with the LVZ, and at about 8.6 ± 0.1 km/s, at a depth of 63.0 ± 2.0 km in the model without the LVZ. As mentioned in Section 5.1 it is possible that there is a reflector across the rift but the velocity underneath it along the axis (7.8-7.9 km/s) is less than the one we find. In either case the relatively high velocity underneath the flank of the rift adjacent to lower velocities underneath the rift axis needs to be explained. The possible presence of a LVZ has also to be discussed.

### 5.4 Petrological modelling

A technique that allows the evaluation of the seismic velocity of a rock at a known pressure and temperature from estimated elastic properties of individual minerals is described in detail in Fuchs (1983) and Mechie et al. (1994). Knowing the mineral composition of a rock and in situ temperature and pressure then P- and S-wave velocities can be calculated. We have applied this method to model the seismic velocity in the uppermost mantle underneath the western flank of the rift valley.

The temperature-depth values are estimated from Pollack and Chapman (1977) for a
continental geotherm. Normal heat flow values of about 50 mW m\(^{-2}\), are found on the rift flanks in Kenya with high but very variable values up to around 100 mW m\(^{-2}\), being limited to the rift valley region (Morgan, 1982; Nyblade et al., 1990; Wheildon et al., 1994). An average value of 40 mW m\(^{-2}\) was found on the western shoulder of the rift valley, north of the Nyanza Rift (Wheildon et al., 1994). The pressure can be estimated (Mechie et al., 1994) for a velocity-depth function derived from the seismic model beneath the western rift flank as 17.1 kbar at 59 km depth or 18.4 kbar at 63 km depth. Errors caused by uncertainty in the pressure are thought unlikely to be significant as a change of 2 kbar (equivalent to about 6 km change in depth) produces only a small change of about 0.02 km/s in P-wave velocity.

In contrast a 100° C change in temperature produces a change of about 0.05 km/s in P-wave velocity. The average mantle composition, P1 (Fig. 10) has been derived from xenoliths brought up by Quaternary volcanoes (Henjes-Kunst and Altherr, 1992) and corresponds to a peridotite with 50% olivine, 30% orthopyroxene, 18% clinopyroxene and 2% spinel. At 63 km depth some garnet may be present at the expense of spinel, and taking into account that the spinel-garnet transition is likely to occur over a wide interval, a garnet peridotite, P3 (Fig. 11) including 55% olivine, 23% orthopyroxene, 14% clinopyroxene and 8% garnet has also been modelled.

Combining a surface heat flow of 40-50 mW m\(^{-2}\) with the Pollack and Chapman (1977) geotherms leads to temperatures of around 600° C at about 60 km depth. Even for such a low temperature, isotropic pure olivine only has a velocity of 8.05 and 8.15 km/s for P1 and P3 respectively (Fig. 10a and 11a). However the velocity of either 8.4 km/s at 59 km depth or 8.6 km/s at 63 km depth beneath the western rift flank can be explained if preferred mineral orientation (anisotropy) is invoked. Invoking anisotropy, either the transverse isotropic olivine model or the orthorhombic olivine model is possible. In the case of the transverse isotropic model the faster 'a' and 'c' axes of the olivine should be randomly orientated in the horizontal plane. Although at temperatures as low as 600° C at around 60 km depth both P1 and P3 have velocities of 8.3 - 8.4 km/s compositions richer in olivine can have velocities up to 8.6 km/s even at temperatures up to 800 – 800° C (Figs. 10b-c and 11b-c). In this respect it is interesting to note that Kimberlites of uncertain age near the western end of the cross-rift line have brought up ultramafic xenoliths including olivine-rich harzburgites and dunites (Ito, 1986). In the case of the orthorhombic model, the fast 'a' axis of the olivine should be orientated along the direction of the cross-rift line. In the case of the model without a LVZ for both P1 and P3 at temperatures between 600° and 800° C at 63 km depth, more than 75% of the olivine should be preferentially orientated in order to
attain a velocity of 8.6 km/s. For the model with a LVZ more than 50% of the olivine should be preferentially orientated in order to attain a velocity of 8.4 km/s for both P1 and P3 at temperatures between 600° and 800° C at 59 km depth. For compositions richer in olivine a smaller amount of the mineral is required to be preferentially orientated.

Keher et al. (1994) and Mechie et al. (1994) have also invoked anisotropy to explain high velocity layers at 40-45 km and 60-65 km depth in the mantle below the axis of the northern part of the Kenya Rift. Their high velocities could be explained by either a transverse isotropic model similar to that utilized here or by an orthorhombic model in which the fast ‘a’ axis of the olivine crystals is orientated along the rift axis. If the layer identified at 60-65 km depth beneath the rift axis and the layer identified here have a common origin, then either a transverse isotropic model or an orthorhombic model with the fast ‘a’ axis of olivine orientated radially away from the apex of the Kenya dome could account for the observed seismic velocities. In either case the cause of the anisotropy could be shear stress along more or less horizontal planes perpendicular to the slow ‘b’ axis of olivine (Nicolas et al., 1971) and possibly directed radially away from the apex of the Kenya dome.

6 Discussion

The upper mantle seismic velocity distribution beneath the western flank of the rift derived here is supported by both kinematic and dynamic modelling of the seismic wavefield and petrological modelling. Our proposed best-fitting model is shown in Fig. 12a. An alternative model can be envisaged with a LVZ between 55 and 59 km, and 8.4 km/s velocity underneath the LVZ (Fig. 7a). The implications are similar. For comparison the axial line model is also shown in Fig. 12b. The reflector modelled at 63 km depth marks an increase in velocity to 8.6 km/s. This velocity is consistent with temperatures of 600 – 800°C, a pressure of 18.4 kbar and a petrological model involving the presence of garnet and preferred orientation (anisotropy) of the olivine crystals either in a transverse isotropic or orthorhombic structure.

There are four points of note:

(A) An upper mantle Low Velocity Zone.

We cannot exclude the presence of a Low Velocity Zone in the uppermost mantle underneath the western flank of the rift. Low velocity layers have been identified in the upper mantle beneath the axial line. The velocity inside the LVZ beneath the cross rift line cannot be constrained because there is no reflection (with inverted polarity) seen from the top of the LVZ. The finite-difference synthetic seismogram modelling provides similar amplitude
ratios for the model with and without a LVZ. Petrological modelling requires relatively low temperatures that do not suggest the presence of melt at depth, but cannot exclude its presence laterally penetrating from the axial zone of the rift. Alternating high and low velocity layers have been found beneath the axis of the rift, where other geophysical results suggest the presence of anomalous mantle and partial melt. The low velocity layers are present below the northern part of the rift and do not extend as far south as the cross-rift line. Elsewhere beneath rifts or tectonically active areas alternating high and low velocity layers have also been found (Prodehl, 1981; Prodehl et al., 1992; Nagumo et al., 1981).

The following considerations are valid assuming that the reflector seen under the western flank of the rift does not extend under the rift axis, and therefore the high velocity is limited to occur under the western flank of the rift:

- if the model with the LVZ is accepted the anomalous mantle material occurring beneath the rift may extend laterally into this LVZ under the rift shoulder.
- if the model without the LVZ is accepted, the anomalous mantle material can be laterally constrained to exist beneath the surface expression of the rift.

We are inclined to prefer the second possibility because there is direct evidence for the presence of the high velocity below the reflector, whereas there is no positive evidence for the LVZ above it. Also there is scant evidence for the LVZ beneath Lake Baringo from the axial line data.

**B) Teleseismic results.**

The results from the KRISP 90 teleseismic survey show (Achauer et al., 1994), in an east-west section through Lake Baringo, a narrow wedge of low-velocity material underneath the rift axis broadening at about 100 km depth. The width of this zone at Moho depth is about 20 km, extending to 30-40 km at 68 km depth, and it has velocities of between 7.5 to 7.8 km/s in the uppermost mantle. Such a pattern would fit the width of the anomalous mantle found here. It does not show: 1) a thin layer of LVZ beneath the margin of the rift, and 2) high velocity values under the rift flank. However, firstly the depth resolution of the teleseismic results is not capable of resolving the thickness of the LVZ layer modelled here, and secondly the proposed anisotropy giving the high velocity values is in the horizontal plane, so it could not be seen by the rays inverted in the teleseismic analysis, supposed to travel along near vertical paths.

**C) High velocity value / Anisotropy.**

Examples of high velocities in upper mantle material have been identified in regions surrounding other rifting areas. In the Jordan-Dead Sea Rift (Ginzburg et al., 1979) a
discontinuity at a depth of 55 km within the upper mantle marks an increase in velocity to 8.6 km/s, a result obtained from seismic data recorded along the rift axis. In the Afar region anisotropy in the mantle under the highlands surrounding the depression has been suggested to explain velocities higher than beneath the depression itself (Gehlen et al., 1975). In the Baikal region (Burmakov et al., 1987) high velocities in the mantle have been explained by anisotropy in garnet pyrohite, from the interpretation of a seismic line across the Baikal rift axis.

Anisotropy in the mantle has also been suggested to exist in other geodynamic regions, where alternating high and low velocity layers have been found. In Fennoscandia (Cassell and Fuchs, 1979) it has been suggested that shear stresses may create preferred orientation of olivine crystals which in turn would cause orientation of the maximum velocity direction.

(D) Rifting or orogenic signature?

There is another intriguing possibility. The seismic line crosses the contact between the Archaean Nyanza Craton (ANC) and the Proterozoic Mozambique Orogenic Belt (MOB) approximately 70 km to the west of the rift's western bounding fault, in the vicinity of the Nandi fault. It is considered that the contact results from a Tibetan style collisional orogen (Shackleton and Ries, 1984; Berhe, 1990).

It is possible the upper mantle discontinuity identified here is a pre-rift structure, resulting from that orogen; for example:

- an intra-plate shear zone, such as a décollement surface,
- a relict oceanic or continental crustal eclogitic layer.

One of the clearest images of a sub-horizontal reflector within the mantle at depths equivalent to that discussed here is seen beneath the DRUM deep reflection line across the Caledonides offshore NW Scotland. This line identified the ‘W’ reflector at a depth of 45-50 km (McGeary and Warner, 1985). The reflector was interpreted initially as a major fault or shear zone within the mantle resulting from the Caledonian orogeny possibly reactivated during Mesozoic extension (Warner and McGeary, 1987); subsequently it was interpreted as a ‘relict oceanic and eclogitic component of a pre-Caledonian subduction zone within the lithospheric mantle’ (Morgan et al., 1994).

The correspondence between this and the Kenya Rift geological environment is striking.
Appendix E

7 Conclusion

In conclusion, an interface within the upper mantle beneath the western flank of the rift has been identified. Detailed analyses of the kinematic and dynamic characteristics of the seismic wavefield confirm the presence of a velocity of at least 8.4 km/s underneath the western flank of the rift. The petrological modelling suggests:

- relatively cool temperatures adjacent to the rift, beneath which partial melt material has been predicted,

- the possibility of anisotropy in which the olivine crystal orientation is suggested as an explanation for the observed velocity structure.

- a further possibility is that the reflector is a pre-rift structure, resulting from the ANC-MOB contact.

The temperature distribution modelled from the velocity values could be used to give an indication of the age of anomalously hot mantle material beneath the rift axis. Turcotte and Shubert (1982) model the temperature effect of an intrusion on the surrounding material at different times after the initial intrusion to provide a way to constrain its age. However, even assuming a large initial temperature difference of 500°C between the thermal anomaly and the surrounding lithosphere at about 35-45 km depth, then even 20 Ma after the first appearance of the thermal anomaly, the rise in temperature 20-30 km away from the anomaly would be only 200 - 300°C. Such a temperature rise would cause a P-wave velocity decrease of only about 0.1-0.15 km/s. A better constrained model using also results for example from possible future deep seismic reflection profiling could allow this type of modelling to be applied to the upper mantle beneath the Kenya Rift.

All the evidence together suggests that the upper mantle region of anomalously low velocity material is confined to being below the surface expression of the rift itself, to a depth below 60 km. There is the possibility of a thin lateral intrusion immediately above the high velocity layer which has been shown to exist beneath the western flank at about this depth.

Acknowledgements

We would like to acknowledge the assistance of the Kenya Government during the course of this project. KRISP was funded by the DFG (Germany), the NSF (USA), the EC and the NERC (UK). We thank Dr. J. Luetgert of the USGS, Menlo Park, California for setting up the USGS wide-angle software on the Geology Department Microvax at Leicester University and C. Abbott of Leicester University for writing the CANCE filtering routine. Also,
we thank Dr. I. Hill of Leicester University for his useful suggestions during the processing and interpretation of the data, Dr. A. Saunders of Leicester University for his constructive comments on the petrological modelling during the production of the manuscript. The synthetic seismogram calculations have been performed on the Siemens S600 supercomputer of the Computer Centre at Karlsruhe University. One of us (R. Masotti) was funded by the European Commission grant (Ref. No. 000009) first, and later by the British Council fellowship (Ref. ITA/2281/225/A).

References


Appendix E


Appendix E


Appendix E


List of Figures

Figure 1 Location map of KRISP 90 explosion seismic profiles.

Figure 2. Record section for the Lake Victoria (VIC) shot point. Reducing velocity = 8.0 km/s. Band pass filter 1-20 Hz.

Figure 3. Record section for the Lake Victoria (VIC) shot point emphasizing phases d and d1. Reducing velocity = 8.0 km/s. Band pass filter 1-20 Hz.

Figure 4. Correlated Adaptive Noise Cancelling Concept (CANC); modified from Hattingh (1988).

Figure 5. a: CANC output seismograms for location numbers 27 to 42 and their stack for the d phase.

b: CANC output seismograms for location numbers 27 to 45 and their stack for the d1 phase.

It should be noted that the filter length may be different and in some cases the traces used are not the same for d and d1.

Figure 6. Example of CANC output stacked seismograms d cross-correlated with a time shifted d1 (location number 27 to 45) top: d with a normal polarity d1; bottom: d with an inverted polarity d1.

Figure 7a. Ray-trace diagram for the intra-mantle reflection d1 for a model with a LVZ (bottom) and corresponding travel time diagram (top).

Figure 7b. Ray-trace diagram for the intra-mantle reflection d1 for a model without a LVZ (bottom) and corresponding travel time diagram (top).

Figure 8. Synthetic seismograms for d and d1 phases for the model with (top) and without the LVZ (bottom) calculated by using a finite-difference technique (Sandmeier, 1990).

Figure 9. Comparison of observed and theoretical amplitude ratios for the model with and without the LVZ.
Figure 10a. Ternary diagram showing the relationship between P-wave velocity and mineralogical composition P1 for a pure isotropic model at 63 km depth.

Figure 10b. Ternary diagram showing the relationship between P-wave velocity and mineralogical composition P1 for a transverse isotropic model at 63 km depth.

Figure 10c. Ternary diagram showing the relationship between P-wave velocity and mineralogical composition P1 for a transverse isotropic model at 59 km depth. Note the similarity in the velocity for P1 between Fig. 10b and this figure.

Figure 11a. Ternary diagram showing the relationship between P-wave velocity and mineralogical composition P3 for a pure isotropic model.

Figure 11b. Ternary diagram showing the relationship between P-wave velocity and mineralogical composition P3 for a transverse isotropic model.

Figure 11c. Ternary diagram showing the relationship between P-wave velocity and mineralogical composition P3 for a transverse isotropic model at 59 km depth. Note the similarity in the velocity for P3 between Fig. 11b and this figure.

Figure 12. Velocity model of the cross-rift (top) and axial (bottom - distance scale halved) profiles derived from ray-trace and synthetic seismograms modelling. P-wave velocities are indicated in km/s. The reflector at a depth of c.60 km beneath the axis of the rift on the cross-rift line is obtained from the axial line model.
normal 'd₁' max = 0.870

inverted 'd₁' max = 0.616
Distance (km)

velocities: km/s
OL

P-wave vels.

\[ Z = 63 \text{ Km} \]
\[ P = 18.4 \text{ kbar} \]
\[ T = 600 \text{ deg C} \]

8.30

OPX

CPX+GA

8.00

OL

P-wave vels.

\[ Z = 63 \text{ Km} \]
\[ P = 18.4 \text{ kbar} \]
\[ T = 800 \text{ deg C} \]

8.45

OPX

CPX+GA

8.00

OL

P-wave vels.

\[ Z = 59 \text{ Km} \]
\[ P = 17.1 \text{ kbar} \]
\[ T = 600 \text{ deg C} \]

8.45

OPX

CPX+GA

8.00

© P3

FIG11
KRISP 90 - Cross Rift profile

- sediments and volcanics (Mioc - Rec)
- lower crust
- upper crust
- upper mantle

KRISP 90 - Axial profile
A crustal and uppermost mantle cross-sectional model of the Kenya Rift derived from seismic and gravity data

P.K.H. Maguire *, C.J. Swain b, R. Masotti *, M.A. Khan *

* Department of Geology, University of Leicester, University Road, Leicester LE1 7RH, UK
b British Antarctic Survey, Cambridge, Cambridge, UK

Received 2 July 1992; accepted 11 February 1993
A crustal and uppermost mantle cross-sectional model of the Kenya Rift derived from seismic and gravity data

P.K.H. Maguire a, C.J. Swain b, R. Masotti a, M.A. Khan a

a Department of Geology, University of Leicester, University Road, Leicester LE1 7RH, UK
b British Antarctic Survey, Cambridge, Cambridge, UK

Received 2 July 1992; accepted 11 February 1993

Abstract

The most significant implication of a combined seismic and gravity model derived from the data along the KRISP 90 cross-rift line on the Equator concerns the compensation mechanism for the uplifted East African plateau. The seismic model shows that despite an elevation difference of 400 m the thickness of the crust at both ends of the profile is approximately the same at about 34 km b.s.l. A regional gradient needs to be added to the gravity field calculated from the seismic model to agree with the observed. This regional is consistent with Ebinger et al.'s (1989) proposal that the plateau is being dynamically supported by convective processes in the mantle. The principal features revealed by the seismic modelling are: (1) the presence of the thick graben infill, and in particular a very deep (8 km b.s.l.) low-velocity body beneath the Kerio Valley adjoining the rift's western boundary, the Elgeyo fault; (2) the correlation between seismic velocities in the upper crust and different Precambrian (Archaean and Pan-African Mozambique orogenic belt) crustal packages identified along the profile; (3) a thickening of the crust beneath both margins of the rift; (4) the apparent symmetry in crustal thinning mirroring the surface distribution of Tertiary volcanics and sediments, suggesting a pure shear mechanism of crustal extension; (5) the presence of a wide, anomalously low-velocity (7.6–7.8 km s⁻¹) body at the base of the crust identified along the axial line and which it has been argued must include partial melt. This may be continuous with a deep low-velocity zone beneath the rift previously identified from the study of teleseismic residuals; and (6) the presence of an intra-mantle reflector at a depth of approximately 55 km beneath the western margin of the rift.

1. Introduction

It has been proposed that the East African plateau, comprising the 1000 m elevated Archaean Nyanza craton between the Eastern (Kenyan) and Western rifts of the East African Rift System, is dynamically supported by convective processes in the mantle (Ebinger et al., 1989). The craton's eastern margin abuts the Proterozoic Mozambique orogenic belt (Fig. 1). Close to and parallel to the margin is the Tertiary Kenya Rift extending north–south across the Kenya dome and underlain by an anomalous low-veloc-
Fig. 1. Location map of KRISP 90 explosion seismic profiles.
ity upper mantle zone penetrating to depths of at least 200 km (Achauer et al., 1992). This, it is suggested, results from melt penetrating the lithosphere in the form of diapirs rising beneath the rift, raising the temperature and introducing 3–6% partial melt into the sub-crustal mantle which in turn invades the overlying crust. The rift is a late Tertiary to Recent extensional feature associated with extensive volcanic activity. It is likely that the Kenya Rift lies above a local culmination of the proposed deeper thermal plume beneath the Nyanza craton.

The effects of lithospheric extension over zones of raised mantle temperatures have recently been described in a series of elegant papers simply relating mantle potential temperatures, stretch factors and volumes of generated basaltic magmas (White and McKenzie, 1989). The features of the Kenya Rift seem to be consistent with these ideas. However, it is difficult to compare it with other zones of lithospheric extension because of: (1) its location close to the junction of an Archaean craton and a Proterozoic orogenic belt with associated probable differences in lithospheric structure and composition; (2) its possible offset from the main upwelling of the mantle plume; and in particular (3) its complex orogenic history, resulting in interference of the Tertiary rift with a previous Jurassic rift, the Anza graben, in northern Lake Turkana.

The comparisons can better be done when an improved estimate of the amount of crustal thinning and the volumes of both crustal and sub-crustal intrusive as well as extrusive magmas can be obtained.

The rift has been defined in terms of asymmetric near-surface structures. It is not yet known whether such asymmetry penetrates to deep crustal and even upper mantle depths. It has also been suggested that extension may occur along crustal detachments with thinning occurring offset from the development of the surface half-graben (Bosworth et al., 1986).

To develop these ideas via the production of a crustal and upper mantle seismic-velocity model a series of experiments have been designed and undertaken in Kenya under the acronym KRISP, the Kenya Rift International Seismic Project.

2. KRISP 90

Following a pilot study in 1985 (Henry et al., 1990) KRISP 90 was undertaken to study the crustal and uppermost mantle structure, composition and physical state beneath the Kenya Rift to answer fundamental questions concerning the process of continental rifting (Prodehl et al., 1994b). Three lines were shot (Fig. 1), one along the axis, one on the northeast flank of the rift, and the one reported here, across the rift between the Equator and 1°N.

3. The cross-rift profile

2-D cross-sectional crustal models of the Kenya Rift have been presented in a number of publications. Not one of them has been derived directly from refraction/wide-angle reflection data. The crucial questions to be answered by the KRISP 90 cross-rift profile include:

Does the Archaean Nyanzian craton have a different crustal signature from that of the Mozambique orogenic belt?

Is there an abrupt or transitional change in structure from beneath the rift proper to beneath its flanks?

Is the graben’s asymmetry continued to mid- and lower-crustal depths and even into the upper mantle?

To what extent is there alteration of, or intrusion into, the crust beneath the rift by the upper mantle?

What is the lateral extent of the anomalous low-velocity upper mantle material immediately beneath the Moho?

The exact location of the seismic profile was constrained by a number of factors:

(1) The line should cross the central part of a postulated half-graben.

(2) The line should pass close to Kaptagat (KAP in Fig. 1) on the western flank of the rift to constrain a previous, poorly resolved crustal thickness estimate (Maguire and Long, 1976).

(3) Lake Baringo should be used as a shotpoint having been shown to provide good seismic wave transmission during the 1985 experiment.
(4) Lake Victoria should be used as an efficient shotpoint at the western end of the line.

(5) The line length should enable the $P_p$ (Moho reflection) phase to be seen from the end shotpoints before the seismic wave energy enters the rift proper.

(6) The line should make use of previously recorded data where possible, in particular along a short section between Lake Baringo and the Kerio Valley (Swain et al., 1981).

(7) The line must lie completely along good, driveable roads.

Given these criteria, a 450-km-long cross-rift profile was sited as deployment D (Fig. 1).

4. Geology

A simplified cross-section of the surface geology along the seismic line (Fig. 2) shows the rift and its infill of Tertiary to Recent sediments and volcanics within the Mozambique orogenic belt. However, there are a number of other geological units that are relevant to the final seismic and gravity model. Lake Victoria lies on the metasediments, metavolcanics and intrusives of the Nyanza craton. From the Victoria shotpoint to 55 km, the line crosses an Archaean greenstone belt. Multi-channel seismic profiling results suggest it is possible that the shotpoint itself may be underlain by thin sedimentary cover (Rach, 1985). From 55 to 110 km the line crosses a granitoid belt before intersecting the Precambrian Mozambique orogenic rocks in the vicinity of the Nandi fault. The Nandi fault, which is located at about 110 km from VIC, is thought to represent a steep thrust ramp marking the western limit of the Archaean-Proterozoic collisional margin. Further west and up to Lake Baringo the margin of the craton is marked by a series of structurally imbricated thrust slices of 2.5 Ga Archaean and post-900 Ma Mozambique lithologies with a uniform westward vergence. The remainder of the line, to the east of Baringo, lies within the Proterozoic mobile belt and comprises typical Mozambique lithologies with ages younger than 600 Ma. Locally within the mobile belt, areas of high-grade rocks yielding ages greater than 1 Ga may represent older crustal remnants and appear to have acted as resistant buttresses to large-scale ductile deformation (P.N. Mosley, pers. commun., 1993).

5. Set-up

The 450-km-long cross-rift line was completed as KRISP 90 deployment D between Lake Victo-
ria in the west and Chanlers Falls in the east. The location of the recording sites was along the crooked line shown on Fig. 1. The topographic variation along the profile is extremely large, rising from about 1 km at Lake Victoria to 3 km at the Elgeyo escarpment, which itself has a sur-

![Diagram of geological section and observation scheme]

**KRISP line D - Schematic geological section**

- sediments (Recent)
- volcanics (Upper Mioc.- Rec.)
- Mozambique Orogenic Belt (Precambrian)
- Nyanzian Metavolcanics (Precambrian)
- Nyanzian Intrusives (Precambrian)
- geological boundary
- fault

**KRISP line D - Observation scheme**

- clearly identified data
- poorly identified data
- complete range

Fig. 2. Simplified geological cross-section and observation scheme for the KRISP cross-rift line D.
face throw of nearly 2 km. A climb to 2 km over
the Kamasia Hills is followed by a drop to 1 km
in Lake Baringo and then a more gentle rise over
the eastern margin of the rift to a maximum of 2
km at about 280 km from the Lake Victoria
shotpoint. Thereafter the topography gently falls
towards the eastern end of the line to 0.7 km at
Chantiers Falls. Shots were fired at Lake Victoria
(VIC), Kaptagat (KAP), Lake Baringo where
there was one large shot (BAR) and three smaller
shots (BAX1, BAX2, BAX3), Tangulbei (TAN)
and Barsalinga (BAS). A total of about 250
recording stations were set out at intervals rang­
ing from about 1 km in the rift itself to about 5
km at both ends of the line. For logistic reasons,
no major shot was recorded from the eastern end
of the line. However, the easternmost 50 km was
occupied during the deployment of the flank line,
deployment E (Prodehl et al., 1994a) when the
shot at Chantiers Falls (CHF) was fired. VIC,
BAR and the three BAX shots were fired in
lakes. KAP, TAN, BAS and CHF were fired in
boreholes. Other details of the shotpoint and
recording parameters may be obtained elsewhere
(Prodehl et al., 1994b).

6. Data

The observation scheme (Fig. 2) indicates that
the two lake shotpoints provided good data along
the complete profile. The greatest distance at
which clear borehole-shot first arrivals could be
identified was about 200 km from BAS to the
west. The shortest distance reached by good
first-arrival data (excluding the CHF shot) was
about 170 km from TAN to the west. However,
even for this shot, clear second-arrival phases
could be observed to 220 km. It may be noted
that clear seismic phases are difficult to identify
within the rift for those shots fired on the flanks.
The four shots from Lake Baringo separated
by about 250 m along the profile and of different
charge sizes (Prodehl et al., 1994b), provided a
similar signal to noise out to the full recording
distance. A combined plot of the data proved
useful in emphasizing the secondary phases to the
west of the shotpoint between about 180 and 80
km and 1 and 4 s reduced time (see Sect. 9.3.1).

7. The record sections

Identification of the principal seismic phases
on the record section implies their interpretation
in terms of a simple model type. In order to
demonstrate the phases on the record section
diagrams the final theoretical ray-traced arrival
times have also been included. Due to the limita­
tions of 2-D ray-tracing (see Section 8), there are
instances where rays defining a particular phase
do not manage to reach the surface, whereas such
energy is observed in practice.

The following principal phases can be seen:
(1) $P_1$ (phase 'a'), the first arrival, a diving wave
through the upper crust. It is clearly seen on all
the sections but in most cases increased gain is
required, even more than in Fig. 3a, indicating
small velocity gradients in the upper crust.
(2) $P_1$ (phase 'b'), a diving wave from below a
mid-crustal boundary is observed on the record
sections from BAR and TAN.
(3) $P_1$, $P_2$ (phases 'b1', 'b2'), at least one
and sometimes two intracrustal reflections are
seen on most record sections.
(4) $P_3$ (phase 'c'), the Moho reflection can be
seen more or less clearly from all shots.
(5) $P_4$ (phase 'd'), the first-arrival refracted
wave from the uppermost mantle can be observed
from all shotpoints but is weak on some sections.
(6) An upper mantle reflection (phase 'd1'),
from below the Moho is seen on the section from
VIC. A phase labelled 'x', seen on some other
sections, may be a pre-critical phase of the same
origin, but could alternatively be a converted
phase or multiple.

Discussion of the individual arrivals from each
shot is best accomplished in relation to the mod­
elled phases produced after complete interpreta­
tion of the whole data set.

8. Interpretation procedure

Interpretation followed a slightly different pro­
cedure to that used for the axial (Mechie et al.,
1994b) and flank lines (Prodehl et al., 1994a).
Initially, phases were identified from the avail­
able record sections and picked using the USGS
SEGY wide-angle processing package BSECS90,
(written by J. Luetgert, U.S. Geological Survey, Menlo Park, Calif.). The picking accuracy was in general better than ±0.1 s. When the phase identification resulted in this accuracy being estimated worse than about ±0.3 s, no pick was produced.

Because of the very complex structure beneath the rift introducing large "static" delays to deep crustal and upper mantle phases, considerable time was spent in attempting to refine the upper crustal structure, involving 1-D modelling, delay time ("Plus-Minus", Hagedorn, 1959; "Time-Term" analysis, Willmore and Bancroft, 1960, and Bamford, 1972), and finally forward modelling of traveltimes and amplitudes. The model was constrained by surface geological observations and in one particular case, namely the presence of two discrete infill layers above the crystalline basement and beneath the lake sediments in the rift itself, by results from a previous local seismic experiment (Swain et al., 1981) which involved a closer shot spacing than the present experiment.

A "top down" approach was applied to derive the velocities and depths to interfaces in the lower crust and uppermost mantle using MACRAY, a newly developed version of the ray-tracing program RAY84, for a Macintosh IICi computer (written by J. Luetgert, U.S. Geological Survey, Menlo Park, Calif.). The program uses a bilinear velocity interpolation procedure. In addition, synthetic seismograms were produced for shots VIC and BAR using the finite-difference method (Kelly et al., 1976; Sandmeier, 1990) enabling a complete description of the seismic wavefield.

The 2-D ray-tracing method is notoriously suspect at tracing rays near boundary and/or velocity grid nodes. Finite-difference synthetic seismogram modelling enables a complete response of the model to the input source function. It is of course subject to limitations, for example model resolution and grid spacing. It can, as well as providing more reliable amplitude information, enable the matching of calculated to observed times in regions where 2-D kinematic ray-tracing breaks down.

The resolution and accuracy of the final model is dependent on a large number of factors, but primarily the correct identification of the various phases and the number of ray paths intersecting a particular volume of the model. Perturbation of the models has shown that resolution of the velocity and depth to interface may be accepted as better than 3% and 7%, respectively, dependent on the uniformity of structure and the ray-path coverage.

The cross-rift line lay normal to the main structural trend and therefore provided the least well-constrained model of the three explosion profiles. The velocity–depth functions beneath BAR and CHF derived from the axial line (Meehan et al., 1994b) and flank line interpretation (Prodehl et al., 1994a), respectively, were used to constrain the model beneath these shotpoints. This proved straightforward in the case of the flank line, but a marginally different section was required beneath BAR.

The record sections were plotted according to their true distance from shot to receiver; however, the relative positions of shots and recording stations used in the ray-tracing were derived by averaging the distances between all shotpoints and all stations. This will have undoubtedly introduced discrepancies between calculated and observed times, but it is considered that these will result in errors which are less than the resolution estimates stated above, and which decrease with increasing shot to station distance.

9. Detailed phase analysis

9.1. Shotpoint VIC

Three versions of the record sections are plotted in Fig. 3, emphasizing different phases. The ray-trace model is shown in Fig. 4, while two finite difference synthetic plots are shown in Fig. 5.

The first-arrival phase 'a' is delayed by about 0.4 s between 0 and 30 km (for a reduction velocity of 6 km s⁻¹) and has an apparent velocity of near 6.1 km s⁻¹ out to 150 km. The upper crustal velocity–depth function has a weak gradi-
Fig. 3. (a) Trace-normalized band-pass filtered (1–25 Hz) record section for shotpoint VIC recorded to the east along the cross-rift line D emphasizing phase 'a'. Reduction velocity 6 km s\(^{-1}\). Note the display gain increased by four and the time scale by two with respect to (b). Phase notation for (a), (b), (c): \(a = P_n\), \(b_1 = P_{11}P\), \(b_2 = P_{12}P\), \(c = P_{n2}P\), \(d = P_m\), \(d_j = \) mantle reflection. (b) Trace-normalized band-pass filtered (1–25 Hz) record section for shotpoint VIC recorded to the east along the cross-rift line D. Reduction velocity 6 km s\(^{-1}\). (c) Trace-normalized band-pass filtered (1–25 Hz) record section for shotpoint VIC recorded to the east along the cross-rift line D. Reduction velocity 8 km s\(^{-1}\).
ent, the ray reaching a maximum depth of about 2.0 km b.s.l.

The modelled critical point of phase 'b1' is at about 75 km. This is a high-amplitude reflection from the top of a layer of velocity 6.5 km s⁻¹ at a depth of about 12 km. Pre-critical energy is seen as close as 40 km.

A second, weaker reflection 'b2' with a critical point at about 135 km is from a lower crustal boundary at a depth of about 25 km. The velocity increase across this boundary is modelled as 0.3 km s⁻¹.

The strongest later arrival seen across the section is phase 'c', a low-frequency reverberative phase with a critical point at 95 km. It can be ray-traced to about 210 km (Fig. 4), at which distance the observed phase emerges slowly from precursory energy. Beyond the cross-over distance with phase 'd' at about 150 km, the crustal phases result in an almost monotonic reverberation across the section from a reduced time of about 0.5 s (Fig. 3b).

Phase 'd', the diving wave into the mantle, is rather weak but recorded out to the maximum distance and well-modelled except for an advance between about 240 and 280 km (Fig. 3c). However, in this region it is refracted along the Moho which is upwarped and therefore arrives early. Ray-tracing (Fig. 4) cannot model these refracted arrivals without local distortion of the Moho structure or velocity gradient beneath the western margin of the rift.

A mantle reflection phase 'd1' is identified immediately after phase 'd' with a modelled critical point at about 270 km. This is interpreted as arising from an intra-mantle interface at a depth of about 55 km.

The finite-difference synthetic sections for the shot from VIC (Fig. 5) have been produced using, for the crustal phases, a wavelet dominant frequency of 2.8 Hz, and for the mantle phase a frequency of 2.25 Hz. Comparison with the appropriate record sections (Fig. 3b,c) shows:

(a) good correlation for the amplitude distribu-
tion of all phases, 'b', possibly extending to a greater distance with higher amplitude than is observed;

(b) consistency with the observed arrival times of phases not ray-traced due to the complex model, for example the advance in arrival time of phase 'a' at about 180 km, and an indication of the advance in phase 'd' between 240 and 280 km;

(c) the modelled rift structure, including three infill layers satisfactorily produces the reverberations observed on the record section;

(d) the modelled Moho reflection appears to transmit energy to at least 330 km as is observed on the record section, which could not be modelled via ray-tracing;

(e) reasonable correlation for the intra-mantle reflection (phase 'd'), although the observed phase does not appear to extend to quite such a great distance as the modelled phase.

9.2. Shotpoint KAP

Record sections from KAP to the west and to the east are shown in Fig. 6. The ray-trace model is shown in Fig. 7.

9.2.1. KAP to the west

The KAP shotpoint was sited on the Uasin Gishu phonolites which cause a small near-source delay for phase 'a'. The modelled arrival occurs approximately 0.25 s before the observed arrival between 30 to 70 km west of KAP.

The large-amplitude phase 'b', is modelled with a critical point at about 80 km, advancing to a reduced time of about 0.1 s at 140 km from KAP.

Phase 'c' can be identified on the most distant few seismograms, but the separation of the traces and its relatively low amplitude make positive identification of its onset tenuous. It is modelled with a critical point at about 105 km.

9.2.2. KAP to the east

The 0.4-s delay on the first-arrival phase 'a' decreases towards the rift margin before a large delay is introduced at about 30 km where the energy enters the rift proper. Within the rift the energy is reverberative and of low signal-to-noise ratio, making identification of arrivals difficult. From the ray-diagram (Fig. 7) the first arrival appears to result from a refraction across the Elgeyo fault, subsequently being overtaken by a
Fig. 5. (a) Finite-difference synthetic seismogram section for shotpoint VIC recorded to the east along the cross-rift line D. Grid spacing 40 m. Dominant frequency 2.8 Hz. Reduction velocity 6 km s⁻¹. Correlated phases as for Fig. 3b. (b) Finite-difference synthetic seismogram section for shotpoint VIC recorded to the east along the cross-rift line D. Grid spacing 50 m. Dominant frequency 2.25 Hz. Reduction velocity 8 km s⁻¹. Correlated phases as for Fig. 3c.
Fig. 6. (a) Trace-normalized band-pass filtered (5–15 Hz) record section for shotpoint KAP recorded to the west along the cross-rift line D. Reduction velocity 6 km s⁻¹. Phase notation for (a) and (b): a = direct and refracted phases through rift infill, b = Pn, c = PmP, d = Pn. (b) Trace-normalized band-pass filtered (5–15 Hz) record section for shotpoint KAP recorded to the east along the cross-rift line D. Reduction velocity 6 km s⁻¹.
diving wave from within a layer of velocity 5.6 km s⁻¹. This in turn is overtaken by the conventional phase ‘a’, the diving wave from within the crystalline basement.

Phase ‘c’ is almost impossible to identify near the modelled critical point at about 100 km, but is consistent with the diffuse energy observed beyond about 140 km from the shotpoint.

Phase ‘d’, although weak, can be clearly identified to a distance of about 220 km.

9.3. Shotpoint BAR

The record sections from BAR are shown in Fig. 8. The ray-trace model is demonstrated in Fig. 9, while Fig. 10 shows the results of the finite-difference synthetic seismogram modelling.

9.3.1. BAR to the west

There is a delay of some 1.5 seconds at the shotpoint, indicative of a substantial thickness of low-velocity sediments and volcanics above the crystalline basement. The first arrival out to about 40 km originates in sequence from the surface layer (2.2 km s⁻¹), and the two underlying layers (3.8 and 5.8 km s⁻¹) before being overtaken by phase ‘a’.

A delay in phase ‘a’ at approximately 55 km occurs over the Kerio basin and is followed by an advance over the basement outcrop at the base of the Elgeyo escarpment. There is a further model misfit, phase ‘a’ being modelled approximately 0.15 s early between 80 and 120 km from BAR. There is a cross-over to a higher-velocity phase ‘b’ at about 160 km.

It is difficult to identify exact onsets of later phases on the reverberative traces. However, phase ‘b₁’ correlates with energy concentrated at about 100 km from the shotpoint, and asymptotically tends towards phase ‘a’ at long offsets.

The combined plot of the four Lake Baringo shots (Fig. 8b) successfully enhances the second-arrival energy identified as phase ‘c’. The critical point is modelled at about 105 km.

Phase ‘d’ is poorly defined but can be positively identified on at least two traces (Fig. 8c). Ray-tracing (Fig. 9) does not produce any diving rays arriving at the surface from BAR and those observed must be refracted along the Moho or diffracted. In this case the observed times are
indicated in Figs. 8a–8c by a dashed line. In order to simulate them the structure/velocity model of the Moho beneath the rift western margin was temporarily perturbed enough to enable rays to emerge at the surface. This suggested that the modelled times are reasonable being only about 0.2–0.3 s early.

A further high-amplitude phase 'x' can be seen on the combined plot (Fig. 8b), apparently following phase 'c'. It can be modelled quite successfully as a reflection from a deep interface at about 55 km, but it has to be pre-critical energy. This is slightly surprising, since the equivalent phase 'd1' from VIC reached its maximum ampli-
Fig. 8 (continued).

Fig. 9. Shotpoint BAR ray-traced final model. For velocity distribution see Fig. 16. Number of rays for each phase limited to a maximum of 20 for display.
Fig. 10. (a) Finite-difference synthetic seismogram section for shotpoint BAR recorded to the west along the cross-rift line D. Grid spacing 40 m. Dominant frequency 2.8 Hz. Reduction velocity 6 km s\(^{-1}\). Correlated phases as for Fig. 8. (b) Finite-difference synthetic seismogram section for shotpoint BAR recorded to the east along the cross-rift line D. Grid spacing 40 m. Dominant frequency 2.8 Hz. Reduction velocity 6 km s\(^{-1}\). Correlated phases as for Fig. 8.
tude at the critical distance, and no significant pre-critical energy was observed. Alternatively, it can be modelled as multiple energy from the upper crustal layer and the further possibility of it being a converted phase should not be ruled out.

9.3.2. BAR to the east

Once again the large delay at the source is indicative of the thick, low-velocity rift infill, the first arrival out to about 65 km emerging from the surface sediments and two underlying layers above the basement.

Phase ‘a’ can be modelled to 110 km at which point it intersects a high-velocity block in the upper crust, and beyond which the observed arrival decreases in amplitude.

The diving wave from the lower crust produces the weak first arrival occurring between about 100 and 200 km. Phase ‘b’, itself is a clear, but weak reflection from the base of the upper crust with a critical point at about 100 km from BAR. It has a complex ray path when passing through the high-velocity block.

Phase ‘c’, with a critical point at about 120 km, extends back to within about 95 km of BAR, at which point it must be pre-critical. It originates from the Moho at a depth of about 34 km beneath the eastern margin of the rift, and becomes diffuse beyond about 150 km from the shotpoint.

Phase ‘d’ seen as a clear first arrival (Fig. 8e), occurs beyond about 190 km from BAR.

Once again there is a later phase ‘x’ following ‘c’, similar to that identified on the section from BAR to the west. The same statements apply concerning its origin.

The synthetic sections for the shot from BAR (Fig. 10) have been produced using a wavelet
dominant frequency of 2.8 Hz. Comparison with the appropriate record sections (Figs. 8a, 8d) shows: (1) a clearly seen delay over the Kerio basin and advance over base of the Elgeyo escarpment between 50 and 60 km from the shot; (2) a Moho diving phase ‘d’ that is consistent with the equivalent observed phase both to the west and east of BAR, but which was not produced by ray-tracing; (3) a Moho reflection phase ‘c’ that appears stronger than observed, possibly suggesting a more complex interface than modelled in the reflection zone.

9.4. Shotpoint TAN

The record sections are shown in Fig. 11, while the ray-trace model is presented in Fig. 12.

9.4.1. TAN to the east

The first arrival to 150 km represents diving waves from layers within the rift and the crystalline basement, phase ‘a’. Thereafter it can be seen continuing towards 200 km from the shot-point. There is good correlation between the modelled and observed arrivals across the complex structure of the Kerio Valley and Elgeyo escarpment.

Phase ‘b’, can be observed with a critical point at about 90 km. No other intra-crustal phases can be confidently identified in the reverberative coda.

Similarly, phase ‘c’ is poorly defined but is modelled with a critical point at about 105 km.

There is no well defined mantle diving wave, phase ‘d’. Given the reciprocal time from VIC to TAN, this should occur at about –4.8 s reduced time at the farthest offset recording on the section. Calculated times for phase ‘d’ can only be obtained by ray-tracing for rays emerging at about 160 km.

9.4.2. AN to the east

To the east of the Tangukei shotpoint, the first arrival once again comprises phases from the
rift infill and the crystalline basement to a distance of about 80 km. The model does not enable this phase to continue beyond the high-velocity upper crustal block, and it can be seen that the amplitude of the first arrival is apparently attenuated at about this distance.

The reflection from the base of the upper crust 'b_r' has a critical point at about 75 km. No rays are traced to the surface until approximately 150 km emerging from underneath the high-velocity block (Fig. 12).

A diving phase 'b' from the top of the mid-crustal layer is identified from about 150 to 190 km.

Phase 'c' is diffuse, but identifiable with a critical point at about 110 km.

Later phases are masked by the strongly reverberative signal and low signal-to-noise ratio, but phase 'c' may be identified and is modelled with a critical distance of 130 km.

The severe limitations in kinematic ray-tracing discussed earlier (see Sect. 8) result in no diving waves emerging between about 90 and 150 km from BAS due to the very complex structure between BAS and the rift. Temporary perturbations of velocity gradient and "corner" points in the rift interfaces suggest a reasonable correlation between the upper and mid-crustal diving phases and the observed first arrivals between about 90 and 150 km from the shotpoint. An important phase 'd' can be identified at a distance of 200 km, and is reasonably modelled by waves passing through the sub-rift anomalous up-

9.5. Shotpoint BAS

The record sections and modelled rays for this shot are shown in Figs. 13 and 14.

9.5.1. BAS to the west

Phase 'a' is modelled to 42 km from BAS. Beyond this distance it is overtaken by a diving wave entering the high-velocity block beneath the shotpoint, required to accommodate the advance in first arrivals from about this distance. It can only be modelled to a distance of about 90 km.

Phase 'b_r' is a strong reflection with a critical point at about 65 km, a pre-critical phase being identified back to about 45 km.

Later phases are masked by the strongly reverberative signal and low signal-to-noise ratio, but phase 'c' may be identified and is modelled with a critical distance of 130 km.

Fig. 12. Shotpoint TAN ray-traced final model. For velocity distribution see Fig. 16. Number of rays for each phase limited to a maximum of 20 for display.
Fig. 13. (a) Trace-normalized band-pass filtered (1–20 Hz) record section for shotpoint BAS recorded to the west along the cross-rift line D. Reduction velocity 6 km s\(^{-1}\). Phase notation for (a) and (b): \(a = P_\text{w} \), \(b = P_\text{d} P \), \(c = P_\text{d} P \), \(d = P_\text{w} \). Note: dashed line identifies observed but not ray-traced phase 'a' (see text). (b) Trace-normalized band-pass filtered (1–20 Hz) record section for shotpoint BAS recorded to the east along the cross-rift line D. Reduction velocity 6 km s\(^{-1}\).
per mantle zone having a velocity of about 7.6 km s⁻¹.

There is no evidence for a mantle reflection originating from below the Moho.

9.5.2. BAS to the east

Phase 'a' is clearly observed to the east of BAS. Its amplitude on the normalized section decreases when the large-amplitude phase 'b₁' appears.

Phase 'b₁' arises as a reflection from the top of the mid-crustal layer and is consistent with this interface rising towards CHF and the modelled depth on the flank line (Prodehl et al., 1994a).

It is probable that the high-amplitude phase occurring at about 2.5 s reduced time on the two traces at the end of the section is phase 'c' modelled as the pre-critical reflection from the Moho.

9.6. Shotpoint CHF

The CHF record section is shown in Fig. 15. The few stations recording the CHF shot detonated during occupation of the flank line, identify a clear phase 'a'. The reflected phase 'b₁' from the base of the upper crust is also seen.

10. Model description

The lowest velocities (about 2.2 km s⁻¹) in the final model are associated with the recent lake sediments in the vicinity of Lake Baringo. These reach their thickest development in the southwest corner of the lake. A thin veneer is also modelled immediately to the west of the Tanguilbei shotpoint, TAN.

There is no evidence of significant sedimentary cover at the western end of the profile. There is however a shotpoint delay at VIC which has been modelled by a 1400-m-thick layer of lake sediments beneath the shotpoint, not inconsistent with Rach's (1985) observations.

Within the rift proper, the principal near-surface layer has a velocity of about 3.8 km s⁻¹. A similar velocity was previously associated with the Miocene–Early Pliocene phonolites, trachytes and sediments (Swain et al., 1981) which infill the rift. Beneath the Kamasia Hills the upper layer velocity in the model increases to about 4.4 km s⁻¹. It is possible that an alternative model of a thinner lower-velocity (3.8 km s⁻¹) layer above the 5.7 km s⁻¹ material would also satisfy the data. This would be more consistent with the results of Swain et al.’s (1981) experiment, which

![Fig. 14. Shotpoint BAS ray-traced final model. For velocity distribution see Fig. 16. Number of rays for each phase limited to a maximum of 20 for display.](image-url)
showed a 2.5-km-thick top layer of velocity 3.7 km s\(^{-1}\) beneath the Kamasia Hills, in place of the 3.5-km-thick layer modelled here. Such a discrepancy may have arisen due to the very great complexity in ray-path geometry through this zone.

The deepest of the layers infilling the graben (i.e. that with a velocity greater than 5.0 km s\(^{-1}\) is problematical. Ray-tracing indicates that this layer has a velocity that varies from 5.6 km s\(^{-1}\) beneath the Kerio Valley, through 5.8 beneath Lake Baringo to 5.1 km s\(^{-1}\) in the east of the graben. Swain et al. (1981), identifying an equivalent velocity of 5.5-5.8 km s\(^{-1}\) between the Kerio Valley and Lake Baringo, suggested that it originated from the crystalline basement. However, a higher velocity of more than 6.0 km s\(^{-1}\) is obtained elsewhere for the upper crustal layer in the present model. Elsewhere in the rift, the highest-velocity infill has been measured at 5.1-5.2 km s\(^{-1}\) beneath Lake Naivasha (Mechie et al., 1994b). The layer reaches its maximum depth in a very deep (8.7 km) "trough" beneath the Kerio Valley. The existence of this feature is necessary from analysis of rays originating principally from KAP, BAR and TAN (Figs. 7, 9, 12). The first arrival from KAP, refracted across the Elgeyo fault, defines the western boundary of this low-velocity block to within the station spacing. Its depth is poorly defined, because no rays penetrate beneath it to emerge as upper crust first arrivals.

The near-source delay at the KAP shotpoint (Fig. 8) is interpreted as due to the thick sequence of volcanics on the Uasin Gishu plateau. This unit has a velocity of about 4.3 km s\(^{-1}\). Here its depth is constrained by outcropping Mozambique belt rocks in the escarpment. Its western extent is defined by surface outcrop. Ray-tracing suggests the sequence is synformal.

There are significant lateral variations in velocity throughout the upper crust. The values can generally be measured to better than about 0.1-0.2 km s\(^{-1}\), although this may be exceeded in those sections where there is great structural complexity. At the western end of the profile the increased apparent velocity of phase 'a' in the first 40 km of the VIC section (Fig. 3b) has been interpreted as a true velocity of 6.4 km s\(^{-1}\). The location of this high-velocity zone coincides with the Archaean greenstone belt bordering Lake Victoria. The high-amplitude mid-crustal reflection, 'b', requires a velocity contrast of about 6.2 above 6.5 km s\(^{-1}\) in order to reduce the critical point to 75 km (Fig. 4), thus necessitating "normal" upper crustal velocities to exist beneath the high-velocity zone.

To the east, coinciding with the outcropping Archaean granitoid belt, the velocity is about 6.1 km s\(^{-1}\). Beneath Kaptagat to the Elgeyo escarpment there is an increase in velocity to about 6.3 km s\(^{-1}\). It is primarily defined by data from Kaptagat to the east. There is little control on the depth extent of this zone (Fig. 7). Beneath the rift, the velocity continues at a high upper crustal velocity of about 6.2 km s\(^{-1}\) towards the east before returning to values closer to 6.0 km s\(^{-1}\).
from the rift margin towards the end of the line at Chanlers Falls. There is one exception. Beneath Barsalinga, a high-velocity upper crustal block appears necessary to advance the modelled arrival times from BAS to the west. Shots from the west appear to result in reduced-amplitude first arrivals above this zone, which is consistent with the modelled rays failing to penetrate across the low- to high-velocity sloping interface (Fig. 12). The eastern end of this block is very poorly defined, as is its base. However, the mid-crustal reflections from both BAR and TAN are consistent with the zone being underlain by low-velocity material.

The depth to the mid-crustal boundary has been constrained primarily by the upper crustal reflection 'b,' and then found to be consistent with the diving wave from immediately below this horizon, where this is identified. The accuracy in depth to this boundary may be taken to be 1–2 km where reflections or refractions have originated from it. Reflections from VIC and KAP (Figs. 4, 7) define the step down between about 80 and 100 km. Beneath the rift the depth is constrained by results from the axial profile (Mechie et al., 1994b), and is consistent with the very weak reflected phase from TAN to the west. The rise in the level of this boundary to the east is constrained by reflections from BAR, TAN, and BAS (Figs. 9, 12, 14) together with the flank line model results (Prodehl et al., 1994a). The lower crustal reflection 'b,' is only positively identified from VIC, but has also been included on the axial line model (Mechie et al., 1994b). Elsewhere, there is little evidence for it, in most instances due to the low signal-to-noise level resulting from the reverberative data within the rift itself. In the flank line interpretation (Prodehl et al., 1994a) the boundary is modelled as a second-order discontinuity, and has been treated as such where identified by a dashed line in Fig. 16. The thickness and velocity variations of the lower crustal layer across the model result primarily from linking the well defined sections (beneath the Archaean terrane at the western end, beneath the rift and beneath the flank line) assuming no dramatic lateral changes in velocity or layer thickness. The accuracy in velocity of this layer is generally poor, estimated to be between about 0.2–0.3 km s\(^{-1}\). The accuracy in the depth to its top surface is about 1–2 km where defined by a reflection.

The Moho depth and velocity of the uppermost mantle have been deduced from analysis of phases 'c' and 'd,' the accuracy in depth being about 2–3 km. The accuracy in velocity values are not easy to obtain, but perturbation of the values suggests that, assuming the general features of the model are accepted, they lie between about 0.1 and 0.3 km s\(^{-1}\). Fig. 17 shows the total ray coverage for both phases 'c' and 'd,' demonstrating the poorly constrained sections of the Moho where there are few or no rays. Between Lake Victoria and the Elgeyo escarpment, the Moho depth varies from about 34 to 38 km with an apparent modest thickening immediately to the west of the rift. The sub-Moho velocity is relatively high at about 8.2 km s\(^{-1}\), but not inconsistent with values obtained elsewhere beneath Archaean crust (for a review see Durrheim and Mooney, 1991). Beneath the Elgeyo escarpment the crust thins to 30 km below the rift axis at Lake Baringo and the velocity drops sharply to about 7.6 km s\(^{-1}\). This value is defined from the shallow sub-Moho material beneath the rift. On the eastern margin of the rift, the depth to Moho is well defined at 34 km from phase 'c' between TAN and BAS, and then deepens to the east, reaching a depth of 38 km before rising to 34 km beneath CHF defined from the flank profile (Mechie et al., 1994b). That section of the Moho between 350 and 430 km from shotpoint VIC is defined primarily by the late arrival of phase 'c.' The lateness of this phase had to be modelled by introducing this deep Moho as well as the low-velocity anomalous mantle material extending well to the east of the rift. The wide zone of low velocities (7.6–7.8 km s\(^{-1}\)) extends over a distance range almost identical to the distribution of Tertiary volcanics at the surface. The velocity values immediately beneath the
Moho then increase to be consistent with the value of 8.1 km s$^{-1}$ identified beneath the flank profile. The immediate sub-Moho velocity gradients are small.

A reflector is required at a depth of 55 km beneath the western margin of the rift (between about 140 and 160 km from VIC) to generate the high-amplitude phase ‘d,’ immediately after phase ‘d’ on the VIC record section (Fig. 3c). A reflector at a depth of about 50 km is also identified beneath the flank line (Prodehl et al., 1994a). Phase ‘x’ from shots BAR and TAN to the east, if interpreted as pre-critical, could originate from the same reflector. No analysis of the phase of the reflection from this horizon has yet been undertaken, and thus the velocity beneath is un-

---

Fig. 16. Final 2-D ray-trace model for KRISP 90 cross-rift line D. Dashed lines indicate transitions where there is a change in velocity gradient but no significant discontinuity. (Velocities in km s$^{-1}$.)
constrained. It has arbitrarily been taken as providing a positive impedance contrast, consistent with the present flank profile model. There is no direct evidence for the presence of this reflector beneath the axis of the rift and here the anomalously low-velocity material immediately beneath the Moho has been modelled as extending to a depth of at least 60 km. This is consistent with the axial profile model which identifies a boundary with a small impedance contrast at this depth beneath Lake Baringo (Mechie et al., 1994b) and the recent model derived from teleseismic data (Green et al., 1991). The resulting distribution of low-velocity upper mantle material beneath the rift takes the form of a V-shape, but there is no constraint whatsoever on what happens at its base.

11. Gravity data

A Bouguer gravity profile was constructed from data in the catalogue by Swain and Khan (1977). Generally, the gravity stations follow the same roads as the KRISP 90 seismic stations though they are further apart: 112 gravity stations were used leaving only insignificant gaps. The lines diverge just west of the Elgeyo escarpment by 5 km where different roads were used. The data in the catalogue are corrected for topography and earth curvature to a distance of 167 km using the conventional Bouguer density of 2.67 g cm\(^{-3}\). The station positions were projected onto a straight line with a heading of N\(68^\circ\)E. This is a different procedure for creating a profile than was used for the seismic section, but the differences in distance are almost all less than 1 km except at the easternmost end of the line where they reach 5 km and these are considered negligible at the plotting-scale used.

11.1. Initial density model

Gravity and seismic data tend to be complementary due to the general relationship between seismic velocity and density (Ludwig et al., 1970). The seismic-velocity section (Fig. 16) was converted into an initial density section (Fig. 18) using the velocity/density relation given in Table 1. For the crust and upper mantle rocks (densities 2.70–3.25 g cm\(^{-3}\)) this represents a linear approximation to the Nafe–Drake relation, except

![Fig. 17. P\(_m\)P and P\(_n\) phases ray-traced final model. For velocity distribution see Fig. 16. Number of rays for each phase limited to a maximum of 20 for display.](image-url)
for the upper mantle LVZ which is thought to be caused by partial melting (e.g. Achauer, 1992) and therefore requires a smaller density difference than is indicated by the velocity difference. Volcanics and sediments do not fall on this line and densities similar to those used by Swain et al. (1981) and Swain (1992) have been used.

In converting the velocity section, the subhorizontal interfaces have generally been assumed to separate homogeneous layers, because in most cases the lateral velocity variations are small. The exceptions are: (a) the lower infill layer beneath the rift of velocity greater than 5.0 km s$^{-1}$ which has a lower velocity (5.1 km s$^{-1}$) to the east than elsewhere (5.8 km s$^{-1}$); (b) the basement, which shows a higher velocity (6.2–6.3 km s$^{-1}$) between 130 and 270 km than elsewhere (5.9–6.1 km s$^{-1}$)—no attempt has been made to

---

**Fig. 18.** The observed and calculated gravity profiles along the cross-rift line D showing the initial density model derived directly from the final ray-trace seismic model. The gravity effect of the deep compensation as discussed in the text is also shown. (Densities in g cm$^{-3}$.)
Table 1  
Velocity versus density values used in modelling the gravity data

<table>
<thead>
<tr>
<th>Unit</th>
<th>( V_p ) (km s(^{-1}))</th>
<th>( \rho ) (initial model) (g cm(^{-3}))</th>
<th>( \rho ) (improved model) (g cm(^{-3}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sediments</td>
<td>2.0</td>
<td>2.10</td>
<td>2.10</td>
</tr>
<tr>
<td>Sediments</td>
<td>3.0</td>
<td>2.30</td>
<td>2.30</td>
</tr>
<tr>
<td>Upper volcanics</td>
<td>3.9</td>
<td>2.46</td>
<td>2.46</td>
</tr>
<tr>
<td>Lower volcanics</td>
<td>5.2</td>
<td>2.55</td>
<td>2.55</td>
</tr>
<tr>
<td>Lower volcanics</td>
<td>5.8</td>
<td>2.63</td>
<td>2.50/2.75</td>
</tr>
<tr>
<td>Basement</td>
<td>6.1</td>
<td>2.70</td>
<td>2.70/2.75</td>
</tr>
<tr>
<td>Basement</td>
<td>6.3</td>
<td>2.76</td>
<td>2.80/2.82</td>
</tr>
<tr>
<td>Basement</td>
<td>6.4</td>
<td>2.80</td>
<td>2.80</td>
</tr>
<tr>
<td>Upper crust</td>
<td>6.6</td>
<td>2.84</td>
<td>2.86</td>
</tr>
<tr>
<td>Lower crust</td>
<td>6.85</td>
<td>2.90</td>
<td>2.92</td>
</tr>
<tr>
<td>Upper mantle LVZ</td>
<td>7.6</td>
<td>3.18</td>
<td>3.20</td>
</tr>
<tr>
<td>Upper mantle</td>
<td>8.1</td>
<td>3.25</td>
<td>3.25</td>
</tr>
</tbody>
</table>

model the high-velocity near-surface section at the western end of the profile; (e) the upper mantle LVZ under the rift.

In these cases interfaces have been inserted at the positions of the lateral velocity changes.

11.2. Isostasy

The gravity anomaly due to the density section shown in Fig. 18 was calculated for an essentially two-dimensional model using GRAYMAG (Busby, 1987), and an arbitrary constant subtracted for comparison with the observed anomaly. The two profiles show a general similarity in shape, but there is a major difference in overall gradient along the line. The observed gravity falls by more than 100 mGal from east to west, whereas the calculated gravity shows little overall gradient. A substantial gradient in the Bouguer anomaly is expected if the topography, which is rising steadily from the low-lying plains of eastern Kenya towards the East African plateau, is isostatically compensated, as has been shown elsewhere (Bullard, 1936; Khan and Mansfield, 1971; Banks and Swain, 1978; Bechtel et al., 1987).

Thus, according to the seismic-velocity section, the source of the compensation does not lie in the crust or uppermost mantle: the Moho depth is about the same at both ends of the profile and the velocity just below it is, if anything, higher at the west end, although it is not very well constrained by the data. It appears unlikely that the compensation would be provided by crustal-density variations. Thus it seems that the source of the compensation of the plateau lies deeper within the upper mantle. A similar conclusion was reached by Ebinger et al. (1989) from a study of the admittance function between gravity and topography. Since the plateau appears to be over-compensated, they modelled the admittance in terms of simple convection mechanisms and obtained satisfactory fits with sources at depths between 160 and 400 km. We think the fact that the Moho does not appear to provide the compensation mechanism along the KRISP cross-rift line is highly significant and provides strong support for this hypothesis.

Since we wished to model the Bouguer gravity using the KRISP seismic model as the basis, we decided to remove a regional field representing the gravity effect of the deep compensating masses. The regional (Fig. 18) was calculated by filtering a 2500-km-long topographic profile, including the KRISP line, with a function \( W \), mGal m\(^{-1}\) approximating the observed admittance (Ebinger et al., 1989), but with:

\[
W = 0 \quad \lambda < 600 \text{ km}
\]

\[
W = -0.135 \quad \lambda > 1800 \text{ km}
\]

This follows from the linear filter representation of the isostatic mechanism (Dorman and Lewis, 1970):

\[
G(k) = W(k)H(k) + N(k)
\]

where capitals imply Fourier transforms and \( G \) is the Bouguer gravity, \( H \) is topography, \( W \) is admittance, \( N \) is isostatic anomaly and \( k \) is the wavenumber.

11.3. Improved density model

Barton (1986) has shown that the Nafe–Drake relation actually involves considerable uncertainty in converting from velocity to density (up to about ±0.2 g cm\(^{-3}\)). Bearing this in mind, we attempted to improve the fit to the gravity data by
adjusting the densities of the initial model, but not the interfaces, which the seismic data usually delineate quite well. This does not in general lead to a unique density model (Barton, 1986), but as usual, "geologic reasonableness" does allow certain useful inferences to be made.

An improved model is shown in Fig. 19. Significant changes from the initial model are:

(a) the density of the upper crustal layer between 130 and 270 km has been increased in order to reproduce the large positive anomalies west of the Elgeyo escarpment and within the rift;
(b) the latter was assisted by increasing the density of the "lower volcanics layer" between 215 and 270 km;
(c) introducing an increased upper crustal density (2.75 g cm$^{-3}$) between 270 and 375 km;
(d) reducing the density in the lower infill layer in the rift west of 215 km to fit the large (60 mgal) anomaly across the Elgeyo fault; and
(e) reducing the size of the Moho step at about 350 km where the seismic model is poorly constrained.

It should be noted that the misfit between 340 and 430 km could be further improved by removing the dip in the Moho under this section of the line.

12. Discussion

The final "improved density model" results from a combined interpretation of the seismic and gravity data.

12.1. Upper crustal composition

The variation in seismic velocity and density throughout the upper crust across the section is consistent with variations in composition, both within the Archaean craton, across its refoliated margin and in the Proterozoic Mozambique orogenic belt, as discussed by Smith and Mosley (1993).

The improved density model (Fig. 19) involves increased densities within the uppermost crust from 60 km west of the Elgeyo escarpment to about 100 km east of the rift. Generally the high densities occur throughout the upper crustal layer, unlike previous gravity models (e.g. Swain et al., 1981). The highest density occurs beneath the rift floor and is probably due to a combination of relatively dense metamorphic rocks (hornblende–biotite gneiss outcrops in the Saimo escarpment 20 km to the north (EAGRU, 1973)) and dyke intrusion. The high densities on the western Rift shoulder correlate with hornblende–biotite gneiss which outcrops in the Elgeyo escarpment. The lower eastward dipping interface of the block intersects the surface at the Nandi thrust marking the boundary between the Archaean craton and the Mozambique orogenic belt. The presence to the east of the rift of the high-velocity/high-density block in the upper crust, whose base and eastern margin are poorly constrained by the seismic data is consistent with the location of older crustal rocks within the Mozambique belt (P.N. Mosley, pers. commun., 1993).

12.2. The rift infill

The lower infill layer in the rift (with a seismic velocity of greater than 5.0 km s$^{-1}$) is somewhat enigmatic. In Fig. 19 it has been divided into three sections partly on the basis of its velocity variations and partly to help explain the sharp negative anomaly in the Kerio Valley and the positive anomalies in the rift. Previous interpretations of the Kerio anomaly (e.g. Swain et al., 1981) have inferred the presence of a low-density "basement block", in addition to thicker rift infill, and the seismic profile has independently confirmed it, although its geological nature is still open to speculation. It could simply be an extremely thick sequence of sediments and volcanics. Its density could be higher than 2.5 g cm$^{-3}$ since the seismic data do not constrain its base very well, which could be deeper. The 2.75 g cm$^{-3}$ section of this layer is presumably Samburu basalt, the oldest volcanic unit which outcrops on the east flank of the rift (EAGRU, 1976), but not on the west. The 2.55 g cm$^{-3}$ section should also be Samburu basalt: the difference in density has been invoked to reproduce the steep gravity gradient here, and because its velocity is lower. It is also possible that this anomaly is partly caused by dyke intrusion.

12.3. Lateral variation in lower crustal structure across the rift

While there seems to be a continuous boundary across the complete section at a depth of about 10 to 15 km b.s.l., the lower crustal interface at a depth between 25 and 29 km b.s.l. is not so well defined. It is only positively identified on the cross-rift profile to a distance of about 90 km from VIC, a region which lies entirely within the Nyanga craton. Elsewhere, within the Mozambique orogenic belt it is not seen, except beneath the axial profile (Mechie et al., 1994b). Beneath
the flank line, the top of this lowest crustal layer is identified as a second-order discontinuity or a change in the velocity gradient with depth. Thus there appears to be a consistent difference in the lower crustal structure beneath the craton and orogenic belt. The presence of the strong reflector beneath the rift itself may be related to intrusive material penetrating the crust to this deep level, the reflector identifying the top of a sill-like body. The velocity variation across the section within the lower crust is strictly too small to be significant. However, it is noticeable that beneath the rift the values are slightly lower than beneath the margins, a decrease which may occur in response to raised crustal temperatures.

12.4. Moho topography

Whereas in the initial interpretation of the cross-rift data (KRISP Working Party, 1991) there appeared to be an asymmetry in Moho depth across the rift, in the present more refined interpretation this has all but disappeared. It has gone for two reasons: (1) the near-source data from VIC, not available for the previous interpretation, requires the presence of a small thickness of sediments beneath the shotpoint at VIC, and this has resulted in the raising of all the interfaces beneath this end of the profile; (2) the P data from BAR to the east, not identified in the earlier interpretation, requires the presence of the “down-dropped” section of Moho between 350 to 420 km. The present model suggests a more symmetrical crustal thickness variation across the rift, there being a thickening beneath the two flanks. Were such a thickening to have existed prior to the initiation of rifting, crustal fracture seems to have preferred that part of the crust that is thickest. This is exactly in accord with models of the rifting process examined in relation to lithospheric strength (e.g. Kusznir and Park, 1987).

12.5. A pure-shear mechanism

The crustal thinning occurring beneath the rift appears to mirror the lateral distribution of Tertiary volcanics and sediments. Also, the lateral extent of the anomalous upper mantle material immediately beneath the Moho extends over the same width. It has been suggested that the rifting process in Kenya takes the form of simple shear failure of the lithosphere (Bosworth et al., 1986), necessitating significant offset between the surface half-graben and the upraised Moho and thinned lithosphere. The present results suggest that this is not the case, the thinned crust lying immediately beneath the surface rift consistent with the pure-shear mechanism of crustal failure as proposed by McKenzie (1978) and subsequently developed in many publications elsewhere.

12.6. The anomalous upper mantle body beneath the rift

The anomalous low-velocity/low-density body beneath the rift has been suggested elsewhere (KRISP Working Party, 1991; Mechie et al., 1994a) to represent uppermost mantle material at a raised temperature and including 3–6% partial melt. The shape of this body in the final model (Fig. 19), reflecting the distribution of low velocities in the seismic model, is poorly constrained. However, it is not inconsistent with the concept of melt rising from great depth in the form of diapirs as suggested by Achauer et al. (1992) and “ponding” at the base of the stretched crust. The base of the anomalous material is not identified. It is almost certainly continuous with the low-velocity column of material identified beneath the rift from analysis of teleseismic residuals (Green et al., 1991).

12.7. Intra-mantle boundary

There is a well defined reflection from a depth of about 55 km beneath the western margin of the rift. This horizon is not positively identified elsewhere on the cross-rift profile. However, an equivalent horizon is observed both on the flank (Prodohl et al., 1994a) and the axial profiles (Mechie et al., 1994b), in the latter case there being a small velocity contrast across a boundary at 60 km depth beneath Lake Baringo. A full discussion of this reflector can be found elsewhere (Keller et al., 1994); however, it is impor-
tant to emphasize that there is no significant evidence from the present study for its presence beneath the axis of the rift beneath Lake Baringo, implying a sharp upper mantle transition from normal to anomalous mantle immediately beneath the rift's western margin. There is no clear evidence for or against such a transition beneath the rift's eastern margin.

12.8. Plateau compensation

One of the most important implications of the joint seismic/gravity model has already been mentioned. It is that the source of the isostatic compensation of the East African plateau lies deep (>60 km) within the upper mantle, and is consistent with it being supported by convective motion within the asthenosphere. It is possible that the magmatic and tectonic activity in the Kenya Rift results directly from the deep thermal process occurring beneath the centre of the uplifted Nyanza craton which forms the plateau.

Acknowledgements

We would like to acknowledge the assistance of the Kenya Government during the course of this project. KRISP was funded by the DFG (Germany), the NSF (USA), the EC and the NERC (UK). We thank Dr. J. Luetgert of the USGS, Menlo Park, California for setting up the USGS wide-angle software on the Geology Department Microvax at Leicester University. Also, we thank Dr. J. Mechie and Dr. K.-J. Sandmeier of the Geophysical Institute at Karlsruhe for performing the synthetic seismogram calculations on the Siemens S600 supercomputer of the Computer Centre at Karlsruhe University. We are particularly grateful to Dr. J. Mechie for processing the seismic data, and for his constructive comments during all stages of the interpretation and production of the manuscript. One of us (R. Masotti) was funded by the European Commission grant (Ref. No. 900009).

References


Physics and Evolution of the Earth’s Interior

Series now complete!

Constitution of the Earth’s Interior
Edited by J. Leilwa-Kopystynski and R. Teisseyre
Physics and Evolution of the Earth’s Interior Volume 1
1984 xli + 366 pages
Dfl. 267.00 (US $ 152.50)

Seismic Wave Propagation
In the Earth
By A. Hanyga
Physics and Evolution of the Earth’s Interior Volume 2
1985 xvi + 476 pages
Dfl. 318.00 (US $ 181.75)
ISBN 0-444-99611-7

Continuum Theories in
Solid Earth Physics
Edited by R. Teisseyre
Physics and Evolution of the Earth’s Interior Volume 3
1986 xlv + 566 pages
Dfl. 318.00 (US $ 181.75)

Gravity and Low Frequency Geodynamics
Edited by R. Teisseyre
Physics and Evolution of the Earth’s Interior Volume 4
1989 xli + 476 pages
Dfl. 313.00 (US $ 178.75)
ISBN 0-444-98908-0

Evolution of the Earth and Other Planetary Bodies
Edited by R. Teisseyre, J. Leilwa-Kopystynski and B. Lang
Physics and Evolution of the Earth’s Interior Volume 5
"This volume is a competently constructed up-to-date and detailed summary of planetary evolution. It is for the planetary scientist above other fields; in this category, the book deserves a wide readership simply for its breadth of coverage. Researchers in other fields will also find this a book worth dipping into, and whole lecture courses could be based around its contents. It appears that the initial wish to discuss planetary evolution across the solar system has resulted in an intelligent advanced level treatise that will become widely referenced itself."

Dynamics of the Earth’s Evolution
Edited by R. Teisseyre, L. Czechowski and J. Leilwa-Kopystynski
Physics and Evolution of the Earth’s Interior Volume 6
This sixth volume in the monograph series Physics and Evolution of the Earth’s Interior presents the problems of the mature evolution of the Earth’s Interior. It provides comprehensive coverage of the present state of the mantle convection theory. The relations between palaeomagnetism, plate tectonics and mantle convection theory are discussed. A more general view of the evolution based on the thermodynamics of irreversible processes is also given.

"The Dutch Guilder (Dfl.) prices quoted apply worldwide. US $ prices quoted may be subject to exchange rate fluctuations. Customers in the European Community should add the appropriate VAT rate applicable in their country to this price."