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GEOPHYSICAL AND ISOTOPIC CONSTRAINTS ON THE
LITHOSPHERE OF THE EAST AFRICAN RIFT SYSTEM

by

Kian Fadaie, B.Sc., M.Sc.

A thesis submitted to the Faculty of Graduate Studies
in partial fulfillment of the requirements of the degree
of Doctor of Philosophy

Carleton University
Ottawa, Ontario
December, 1987

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ABSTRACT

The present work combines interpretation of geophysical and geochemical data in an attempt to model the configuration and dynamics of the asthenosphere/lithosphere system in the east African rift, and to evaluate the genesis of associated alkaline rocks.

A comparison of five major continental rift systems and a detailed study of the east African rift indicate that they all have similar geomorphic, magmatic, and geophysical features. This suggests a similarity in the dynamics of their formation.

The existence of an anomalous low-velocity, low-density zone in the upper mantle beneath the east African rift is deduced on the basis of gravitational, seismological and thermal data. This zone is considered to be the result of the presence of hot asthenospheric material under the rift.

Gravity modelling of different regions of the rift shows that in the eastern rift the lithosphere is thinnest at approximately the latitude of the Equator, whereas maximum thinning occurs in the western rift around 1.7°S . The depth to the lithosphere/asthenosphere boundary changes both along and across the rift. This is consistent with seismological data.

The rift system is associated with a region of high heat flow. The corrected heat flow values for the eastern rift, western rift, and the Precambrian shield are 111 ± 53 , 90 ± 34 , and 54 ± 28 mW/m², respectively. The rheological profiles inferred from the geotherms show that the topmost brittle layer and the lithosphere as a whole are, on the average, thinner under the eastern than the western rift. The

thicknesses of the lithosphere estimated from gravity, seismic, thermal, and rheological data are in relatively good agreement.

Total lithospheric strengths using rheological profiles are 27, 62, and 188 MPa for the eastern rift, western rift, and the Precambrian shield respectively. Taking into account the total strength and the possible stress amplification in the upper lithosphere, a combination of stress-producing processes appears to be required to initiate rifting. The principal of these processes is the plateau uplift.

Analyses of 43 samples from the Birunga and Toro-Ankole volcanic fields in the western branch of the east African rift indicate that there is a wide range in major and trace element abundances as well as Sr and Nd isotopic compositions.

The initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios range from 0.70447 to 0.70971 and initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratios from 0.51209 to 0.51269. There is an excellent negative correlation of the data from the Birunga volcanics on the plot of $^{87}\text{Sr}/^{86}\text{Sr}$ vs $^{143}\text{Nd}/^{144}\text{Nd}$. The data extend from the depleted quadrant close to bulk Earth well into the enriched quadrant. The depleted quadrant contains the data from the nephelinites and nepheline leucitites of Niragongo; the data from the remaining Birunga samples fall into the enriched quadrant.

Quantitative modelling of crystal fractionation, mixing, and assimilation combined with crystal fractionation shows that the genesis of the Birunga rocks cannot be completely explained by only one of these processes. Metasomatism and/or mixing of a nephelinitic magma with continental crust can explain most of the chemical characteristics of the Birunga volcanics. These processes are consistent with the upwelling of the asthenospheric material beneath the east African rift that induces metasomatism and partial melting.

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TABLE OF CONTENTS

Abstract.....	11
Acknowledgements.....	iv

CHAPTER 1: CONTINENTAL RIFT SYSTEMS: A GENERAL REVIEW

1.1	Introduction.....	1
1.2	Volcanism.....	7
1.3	Seismicity.....	9
1.4	Body wave and surface wave data.....	13
1.5	Gravity.....	16
1.6	Heat flow and temperature.....	19
1.7	Tectonic hypothesis on rift development.....	22
1.7.1	Regional tectonic tension.....	24
1.7.2	Membrane tectonics.....	24
1.7.3	Stresses associated with uplifted continental regions.....	25
1.7.4	Generalized mantle convection.....	27
1.7.5	Mantle plumes.....	28
1.8	Discussion.....	30

CHAPTER 2: THE EAST AFRICAN RIFT SYSTEMS

2.1	General geology.....	34
2.2	The pattern of rift faulting in relation to Precambrian structures.....	37
2.3	Tectonic history of EARS.....	43
2.4	Major faulting and the rift structures.....	46
2.4.1	The eastern rift valley.....	47
2.4.2	The western rift valley.....	50
2.5	Volcanicity.....	52
2.5.1	The eastern rift.....	52
2.5.2	The western rift.....	55

CHAPTER 3: GRAVITY

3.1	Introduction.....	59
3.2	Description of the anomalies.....	62
3.2.1	The eastern rift.....	62
3.2.2	The western rift.....	63
3.3	Relationship between gravity anomalies and topography.....	65
3.4	Gravity profiles.....	73
3.5	Quantitative interpretation of Bouguer anomalies.....	85

CHAPTER 4: SEISMICITY

4.1	Introduction.....	97
4.2	Macroseismic data.....	97
4.3	Spatial distribution of seismicity: epicentral maps.....	100
4.3.1	Western rift.....	100
4.3.2	Eastern rift.....	103
4.4	Microseismic data.....	103
4.4.1	Western rift.....	103
4.4.2	Eastern rift.....	104
4.5	Magnitude-frequency relations.....	107
4.6	Focal depths.....	110
4.7	Focal mechanism.....	117
4.7.1	Western rift.....	117
4.7.2	Eastern rift.....	119
4.8	Travel-time anomalies.....	119
4.9	Surface wave data.....	122
4.10	Lithospheric structure in the rift zone.....	125
4.11	Discussion.....	131

CHAPTER 5: HEAT FLOW AND REGIONAL GEOTHERMS

5.1	Introduction.....	135
5.2	The western rift.....	136
5.2.1	Lake Kivu.....	136
5.2.2	Lake Tanganyika.....	141
5.2.3	Lake Malawi.....	141
5.3	The eastern rift.....	142
5.4	Heat flow corrections.....	147
5.5	Regional geotherms.....	152
5.6	Thermal thickness of the lithosphere.....	155
5.7	Electrical conductivity.....	159
5.8	Discussion.....	160

CHAPTER 6: THE RHEOLOGY OF THE LITHOSPHERE

6.1	Introduction.....	164
6.2	Laws of brittle and ductile deformation.....	165
6.2.1	Brittle deformation.....	165
6.2.2	Ductile deformation.....	166
6.3	Rheological profiles.....	167
6.4	The total strength of the lithosphere.....	183

CHAPTER 7: GENERAL INTRODUCTION TO THE BIRUNGA FIELD

7.1	Introduction.....	188
7.2	The Birunga field.....	190
7.2.1	The spatial evolution and the age of the volcanoes of the Birunga field.....	191
7.2.2	Volcanic cycles in the Birunga field.....	192
7.3	Association of Volcanoes with geological features.....	193

7.4	Rock samples.....	194
7.5	Classification and nomenclature.....	194
CHAPTER 8: PETROCHEMISTRY		
8.1	Introduction.....	201
8.2	Major and trace element chemistry.....	201
8.2.1	Rare earth element chemistry.....	206
8.3	Incompatible element patterns.....	207
8.4	Primary and derivative magmas in the Birunga and Toro-Ankole fields.....	217
CHAPTER 9: ISOTOPIC SYSTEMATICS		
Sr ISOTOPIC SYSTEMATICS		
9.1	Introduction.....	220
9.2	Sr isotopic analysis.....	220
9.3	Results.....	221
Nd ISOTOPIC SYSTEMATICS		
9.4	Nd isotopic composition.....	229
9.5	Results.....	231
Sr, Nd AND Pb ISOTOPIC SYSTEMATICS		
9.6	Anti-correlation plot (ACP).....	231
9.7	Pb isotopic systematics.....	238
CHAPTER 10: HYPOTHESES FOR THE GENESIS OF ALKALINE ROCKS		
10.1	Introduction.....	241
10.2	The origin of alkalic rocks of the Birunga field	242
10.2.1	Crystal fractionation.....	242
10.2.2	Quantitative simple binary mixing.....	244
10.2.3	Assimilation and fractional crystallization model.....	247
10.2.4	Metasomatism.....	249
10.2.5	Discussion.....	253
CHAPTER 11: CONCLUSION		
11.1	Gravity.....	254
11.2	Seismicity and seismic wave propagation.....	254
11.3	Heat flow.....	255
11.4	Rheology.....	255
11.5	Major and trace element data.....	256
11.6	Isotopic systematics.....	257
11.7	Modelling.....	258
References.....		259

LIST OF TABLES

1.1	Volcanic succession of the rift system.....	10
1.2	Heat flow data from four rift systems of the world.....	21
1.3	Common characteristic features of major continental rift systems.....	23
2.1	Name and age of the orogenies associated with the Precambrian structures in EARS.....	39
2.2	The volcanic history of the western rift.....	39
2.3	Structural evolution of the Kenya rift valley.....	48
2.4	Structural history of the western rift valley.....	51
2.5	Generalized volcanic sequence and corresponding volumes in the Kenya rift valley.....	53
3.1	Depth to the base of the lithosphere in both branches of EARS based on the quantitative gravity modelling.....	95
4.1	Seismograph stations in EARS.....	99
4.2	The number of seismic events in the rift zone.....	109
4.3	The ranges in 'b' values for the two branches of EARS.....	109
4.4	List of earthquakes in the western and the eastern rifts for which the fault plane solutions are determined.....	114
4.5	Seismic delay times for three seismic stations in east Africa and their corresponding Bouguer anomaly values.....	121
4.6	Lithospheric model of the eastern and the western branch of EARS.....	129
5.1	Heat flow measurements in Lake Kivu.....	137
5.2	Heat flow measurements in Lake Tanganyika.....	137
5.3	Heat flow measurements in Lake Malawi.....	139
5.4	Average heat flow values for the three sections of Lake Malawi.....	139
5.5	Heat flow measurements in the Kenya rift.....	143

5.6	Average heat flow values for the floor and flanks of the Kenya rift.....	146
5.7	The alternating high and low heat flow regions in the western rift from north to south.....	146
6.1	Creep parameters for lithospheric materials.....	173
6.2	Temperature structures in the eastern rift, western rift, and the shield region.....	173
6.3	Viscosity of the crust and upper mantle in the rift zone and the Precambrian shield of east Africa.....	178
6.4	Uncertainties in flow-law parameters.....	178
7.1	Sample number, rock type, volcano and the volcanic field for all the samples analysed.....	195
7.2	Petrographic summary of the rocks of the Birunga and Toro-Ankole fields.....	199
8.1	Whole-rock major and trace element analyses.....	202
8.2	The average of some elements for the three groups of volcanoes of the Birunga field.....	205
8.3	The rare earth element and Th analyses.....	205
9.1	Sr isotopic composition of the Birunga and Toro-Ankole volcanics.....	222
9.2	The Nd and Sm concentrations of four samples using the isotope dilution method.....	230
9.3	Reproducibility of Nd isotopic ratios.....	230
9.4	The isotopic analyses of the Birunga and Toro-Ankole fields.....	232
9.5	Pb isotopic composition for some of the Birunga samples.....	240
10.1	Nb and U concentrations for several Birunga samples.....	252

LIST OF FIGURES

1.1	Simplified tectonic maps of Afar, Baikal, Ethiopian and Rio Grande rifts.....	5
1.2	Seismic velocity profiles for Baikal, Rhine and Rio-Grande rifts.....	14
1.3	Schematic cross-section of an idealized continental rift.....	17
1.4	One possible sequence of rift development.....	29
2.1	The east African rift system.....	35
2.2	Structural sketch map of the east African rift system.....	36
2.3	Map of the east African rift system showing Precambrian structures.....	38
2.4	Structural map of east Africa based on mapping from ERTS-1 imagery.....	41
2.5	Broad shape of the uplift of the Kenya dome.....	45
2.6	Pattern of faulting in the eastern branch of EARS.....	49
2.7	Age and distribution of the main volcanic associations of the eastern rift.....	54
2.8	Volcanic fields of the western branch of EARS.....	56
3.1	Bouguer anomaly map of Africa.....	60
3.2	The axis of the linear positive anomaly of the eastern rift..	64
3.3	Contoured map of the Free Air anomaly.....	67
3.4	Contoured map of the Bouguer anomaly.....	68
3.5	Contoured map of the topography.....	69
3.6	Graph of $1^\circ \times 1^\circ$ mean elevation vs $1^\circ \times 1^\circ$ Free Air anomaly..	70
3.7	Graph of $1^\circ \times 1^\circ$ mean elevation vs $1^\circ \times 1^\circ$ Bouguer anomaly...	71
3.8	Histograms of $1^\circ \times 1^\circ$ Free Air, Bouguer anomaly and elevation.....	72
3.9	Profiles of the gravity anomaly.....	78
3.10	Profiles for the amplitude and the extent of the positive Bouguer anomaly over the eastern rift.....	83
3.11	Model of Gumper and Poméroy (1970) for standard African lithosphere.....	87

3.12	Models of the structure of the lithosphere satisfying the smoothed Bouguer anomaly.....	93
4.1	Major seismic stations in the east African rift zone.....	98
4.2	Epicentral map of the earthquakes in EARS.....	101
4.3	Epicentral locations of microearthquakes in the northern Ruwenzori region.....	105
4.4	Plot of body-wave magnitude vs Log N for the earthquakes listed in Appendix 4.1.....	108
4.5	Magnitude-frequency relation for the two branches of EARS....	111
4.6	Histograms of earthquakes hypocentres in the eastern rift, the western rift and the Precambrian shield.....	113
4.7	Fault plane solutions in the Ruwenzori region.....	116
4.8	Fault plane solutions in the eastern rift.....	118
4.9	Plot of delay time vs Bouguer anomaly in EARS.....	124
4.10	A composite true-scale cross-section of the Kenya dome based on delay times of the stations along and across the rift.....	126
4.11	Models resulting from simultaneous inversion of teleseismic data.....	130
4.12	Schematic cross-section illustrating asymmetrical half-graben and the reversal of the main boundary fault along the axis of the rift.....	133
5.1	Heat flow measurements in Lakes Kivu and Tanganyika.....	138
5.2	Heat flow values in Lake Malawi.....	140
5.3	Heat flow measurements in the Kenya rift.....	145
5.4	Geotherm families of continental regions.....	154
5.5	The geotherms for eastern rift, western rift and the Precambrian shield, and corresponding lithospheric thickness.....	154
5.6	Extreme geotherms and corresponding range of lithospheric thickness in the eastern and western rifts.....	156
5.7	A composite true-scale cross section of, the east African rift system incorporating gravity, seismic and heat flow data model thickness of the lithosphere.....	163
6.1	Petrological models based on the lithospheric model of Nolet and Mueller (1982).....	168

6.2	Creep strength of several lithospheric materials in the eastern rift.....	169
6.3	Creep strength of several lithospheric materials in the western rift.....	170
6.4	Creep strength of several lithospheric materials in the shield area.....	171
6.5	Rheological models for eastern rift, western rift and the Precambrian shield.....	176
6.6	The upper and lower boundary of rheological profiles using the uncertainties in the flow-law parameters.....	181
6.7	Field of significant tensional lithosphere deformation compared with estimated stress levels for possible sources of tensional stress.....	185
7.1	The location of the eight major volcanoes of the Birunga field.....	189
7.2	The locations of the samples from the Bufumbira field.....	196
7.3	Sample locations of the Birunga field.....	197
8.1	The chondrite-normalized plots of the rare earth elements....	208
8.2	Spidergrams for different rock types of the Birunga field....	214
8.3	Spidergram for different rock types of the Toro-Ankole field and an average metasomatized mantle nodule.....	215
8.4	The nephelinite-normalized plot of incompatible elements.....	216
8.5	The plot of Na_2O vs Nb for all Birunga samples.....	218
9.1	Histograms of Sr and Nd isotopic ratios for Birunga and Toro-Ankole fields.....	223
9.2	$^{87}\text{Sr}/^{86}\text{Sr}$ vs $^{87}\text{Rb}/^{86}\text{Sr}$ for the Birunga volcanics.....	225
9.3	$^{87}\text{Sr}/^{86}\text{Sr}$ vs Sr for the Birunga samples.....	226
9.4	$^{87}\text{Sr}/^{86}\text{Sr}$ vs SiO_2 for the Birunga rocks.....	228
9.5	$^{143}\text{Nd}/^{144}\text{Nd}$ vs $^{147}\text{Sm}/^{144}\text{Nd}$ for some of the Birunga samples.....	233
9.6	Anti-correlation plot (ACP).....	237
9.7	Comparison of the isotopic ratios from Birunga and Italian volcanic provinces.....	238
10.1	The binary mixing model.....	246

LIST OF APPENDICES

3.1	Average 1° x 1° Free Air anomaly and altitude.....	287
4.1	Time, location, depth and magnitude of all the earthquakes in EARS from 1912 to 1983.....	291
8.1	Major element chemistry (XRF method).....	299
8.2	CIPW normative mineral composition.....	300
9.1	Procedures for isotopic studies.....	303
10.1	Correlation coefficients for the Birunga samples.....	307

CHAPTER 1

CONTINENTAL RIFT SYSTEMS: A GENERAL REVIEW

1.1 INTRODUCTION

Continental rift systems are associated with a variety of geophysical and geochemical processes such as domal uplift, crustal extension, vertical movement, lithospheric thinning, upwelling of mantle material and igneous activity. The study of rift systems is important in order to relate these processes to their underlying tectonic causes and to understand the interaction of asthenosphere and lithosphere. Although rift systems have been extensively studied, their origin and the genesis of the alkaline magmatism associated with them are still the subject of controversy.

The East African Rift System (EARS) is the world's largest continental rift system. The whole sequence of rift processes, from incipient rifting in southern Africa, to major rift valley development in Kenya, to the development of sea floor and continental splitting in the Gulf of Aden, is present (Girdler, 1983).

This thesis is an attempt to explain the geodynamic and magmatic evolution of the east African rift zone. Geophysical data highlight characteristic features of the deep-seated structures and provide constraints for modelling studies of the lithosphere/asthenosphere system under the rift. Geochemical and isotopic data emphasize differences and similarities with other rift systems and other magmatic provinces, and help to identify physical constraints for the formation of the alkaline magmas associated with the rift system.

The classic morphologic picture of a rift or rift valley is that

of a symmetric-parallel-sided graben typically tens of kilometers wide and a few hundred kilometers long. However, several recent studies (for example Sander et al., 1986 and Rosendahl, 1987) show that the cross-sectional morphology is often that of an asymmetric half-graben, with one steep side representing the major bounding fault and the opposing side being characterized by a series of step-like homoclinal fault blocks, monoclines or flexures. Rifts occur in places where the entire thickness of the brittle lithosphere has ruptured under deviatoric tension (Burke, 1977). These extensional features are characterized by abnormally thin lithosphere.

The majority of the continental rifts of the world tends to form in a roughly north-south direction (Ranalli and Tanczyk, 1975). Included in this majority are the African, Rhine and Rio Grande rifts, but the Baikal rift system is a notable exception even after account is taken of rotation due to continental drift. The near-meridional directions of the rifts indicate that these extensional features imitate the trend of the major plate boundaries (Mohr, 1982); a phenomenon which is related either to the proximity of the poles of rotation of the plates to that of the Earth's rotational axis (Burke and Whiteman, 1973) or to the properties of the stress field of the Earth. The common orientation also suggests that the mechanisms of formation of continental rift systems may be similar to one another (Ranalli and Tanczyk, 1975); if this proves to be the case it would imply that their geological and geophysical properties would also be similar. This possibility will be examined briefly below by reference to five major continental rifts, namely, Afar, Baikal, Ethiopian, Rhine, and Rio Grande rifts (Fig. 1.1), but the East African Rift

(A)

(B)

(C)

(D)

(E)

Figure 1.1: Simplified tectonic maps of a) Afar, b) Baikal, c) Ethiopian, d) Rhine and e) Rio Grande rifts. (from various sources).

System (EARS) will be reserved for detailed study in the following chapters.

The Afar rift is one of the few places where the transition from continental rifting to continental breakup can be observed. Barberi and Varet (1975, 1977, 1978) have presented a two-stage model for the formation of this rift, involving initiation in Early Miocene followed by continental breakup and formation of proto-oceanic structure beginning at about 4 Ma.

The Baikal rift system extends for about 2500 km mostly in a southwest-northeast direction. It extends for a distance of 1500 km along the southeastern edge of the Precambrian Siberian platform and about 1000 km in the Sayan-Baikals fold belt. It represents a vertical displacement of up to 6 - 8 km and is the world's deepest intra-continental rift valley (Logatchev and Florensov, 1978). By the study of the thick Cenozoic sedimentary rift formations, two stages of development are recognized for this rift. The first and longer stage is dated from Eocene to Early Pliocene and the second from Pliocene to Recent (Logatchev and Florensov, 1978).

The main Ethiopian rift system begins at around 6°N latitude and extends up to approximately 11°N latitude where it widens into the Afar triangle. The rift valley is tectonically characterized by en-echelon swarms of north-northeast trending faults. The rift has a well defined linear graben to the north of the latitude 9°N whereas to the south of this latitude no clear axial graben is observed (DePaolo, 1972). The tectonic stages for this rift are poorly known because of the widespread Plio-Quaternary cover over the whole area. However, the oldest date relating to the outpouring of the rhyolites on the rift margins dates back to Middle Miocene (Zanettin et al., 1979).

7

The Rhine graben is a coherent structural feature sharply bounded by master faults for a distance of 300 km from Frankfurt to Basel (Illies, 1970). The graben is superposed on a complex system of faults and fractures extending from the Netherlands to southern France. The stratigraphic sequence shows that this rift developed in two stages, the first lasting from Middle Eocene to Lower Miocene (Wagner, 1953) and the second from Middle Pliocene to the Present. The interval between these stages was a period of tectonic quiescence (Barberi et al., 1982).

The Rio Grande rift comprises a series of basin-type structures extending from southern Colorado through central New Mexico into northern Chihuahua, Mexico (Chapin, 1979). According to Ramberg and Larsen (1978), the area of the rift was a shallow sedimentary basin in which rifting began at 32-27 Ma. The uplift associated with the rift started at 7-4 Ma, and together with volcanism has continued up to Present (Chapin and Seager, 1975). That the rift is active is evidenced by fault scarps cutting alluvial fans (Chapin, 1971) and earthquake activity.

1.2 VOLCANISM

Magmatic activity is associated with all continental rift systems. Depending on the volume of the erupted lavas the rifts can be divided into high (Afar rift, eastern branch of EARS, and the Ethiopian rift) and low (Baikal rift, Rhine graben, and the western branch of EARS) volcanicity rifts. Mohr (1982) refers to these as wet and dry rifts, respectively.

Three types of volcanism can be distinguished in the Afar

region: oceanic, continental, and stratoid (flood basalts and ignimbrites). Volcanism of oceanic nature, which represents the latest stage of tectonic evolution, occurs in the rift axes of the Red Sea and the Gulf of Aden. Fractionation of the primary magma of this series, which has a transitional nature between tholeiites and alkali basalts, has produced intermediate lavas and peralkaline rhyolites. Continental volcanism occurs mainly along the rift margins and is characterized by acid lava domes and flows, ignimbrite sheets, and fissural basalt lava fields. Stratoid volcanism occurs mainly in central and southern Afar. This type, characterized by a very high soda content, consists of thick fissural basalts, mainly transitional basalts and rhyolites of peralkaline composition (Barberi et al., 1972).

In the main Ethiopian rift, the volcanicity is characterized by basaltic and silicic magmas, mostly alkaline and peralkaline rhyolites. Strongly alkaline lavas such as nephelinites, basanites, phonolites, and trachytes are reported in the surrounding Ethiopian plateau, but they are absent in the rift itself.

The Baikal rift is characterized by alternating eruptions of basaltic and alkaline lavas from the Early Paleocene to Holocene and by strongly alkaline lavas which are mostly confined to the Quaternary. Although the greatest volume of volcanics is associated with the first stage of rift development, the strongly alkaline lavas are coincident with the second stage of rift evolution (Barberi et al., 1982).

In the Rhine graben, the strongly alkaline and silica undersaturated basic rocks (basanites, nephelinites and ankaratrites) are characteristic of the whole period of activity of the rift, whereas saturated and weakly saturated rocks are confined to the last 20 Ma

(Wimener, 1974). The volcanic activity shows two periods of maximum intensity in Middle to upper Eocene and Pliocene-Quaternary times (Barberi et al., 1982).

Throughout its development, the Rio Grande rift has undergone extensive volcanic activity. Early volcanism consisted of calc-alkalic basalts and basaltic andesites of alkalic affinities, but in the last 5 Ma basaltic lava flows have become more predominant (Chapin and Seager, 1975). In general alkali basalts, basanites and tholeiites are associated with the Pliocene-Holocene volcanic activity in the rift. In contrast to the tholeiitic variety which seems to be confined only to the rift floor, the alkalic types occur on both the rift floor and its shoulders (Kudo et al., 1971).

Comparison of the volcanicity of the above rifts (Table 1.1) shows that the low volcanicity rifts display a wider variation of the chemistry of the erupted lavas. Furthermore, the variety of basic rocks associated with these rifts indicates that strongly undersaturated alkalic rocks (nephelinites and basanites) are predominant. In general, continental rifts are characterized by alkaline magmatism from transitional to strongly alkaline character. The alkaline rocks are usually enriched in alkalis (both Na and K), incompatible trace elements, rare earth elements (especially light REE), carbon dioxide, and water.

1.3 SEISMICITY

Although the detectability of earthquakes for the Afar region is generally poor, it is known that the seismicity occurs mostly in swarms and appears to be concentrated mainly along the Ethiopian escarpment

Table 1.1: Volcanic succession of the rift systems

Time	AFAR	BAIKAL
	Basaltic fissure eruption : Cinder cones & restricted Trachytes and rare : Lava flows Comendites (N Afar) : Benmorites and trachytes Quaternary : Volcanic eruption of intermediate composition : Alkali olivine basalts and Last basaltic eruption of : Basanites Stratoid series : Beginning of rifting	
Pliocene	Flood basalts (East and South Afar) Eruption of stratoid series (Southern Afar)	Fissure volcanic activity : Mugearites, Benmorites and Trachytes Alkaline and subalkaline Basalts
Miocene	Basaltic & rhyolitic flows : Basaltic and subalkaline Alkaline intrusions (the Ethiopian escarpment) : Basalts Pre-rifting stage	
Eocene	Basaltic volcanism : Lava flows and pyroclastic units of basaltic composition (Afar Depression) First stratoid series Oligocene : Peralkaline series (Plateaus & rift margin)	
Source	Barberi et al., 1972 Mohr, 1982	Kiselev et al., 1978

Table 1.1 (ctd)

Time	Ethiopian	Rhine	Rio-Grande
Quaternary	Trachyte & Alkali Rhyolites (The Rift Axis) Basanites, Phonolites & Trachytes (Plateaus)	Phonolites & Melilite Ankaratrites	Alk. Basalts, Basaltic Flows & Cinder Cones Culminating Rhyolite Stage
Pliocene	Ignimbritic Rhyolites, Scoriaeous Basalts & Pyroclastites Basanites, Melanephelinites (Rift Margin & Plateaus)	Phonolites & Melilite Ankaratrites	Basaltic Lavas, Differentiated alkalic Volcanics Basaltic and high Silica Rhyolitic Eruptions
Miocene	Peralkaline Trachytes and Phonolites (Plateaus)	Tholeiitic Basalts, Melilite Ankaratrites Phonolites (Restricted to Central Volcanoes)	Rhyolitic Ash Flow Tuffs, Alkali Andesites and Basalts Quartz Monzonites and Latites
Eocene - Oligocene	Fissural Basalts, Ignimbrites and Basaltic Flows	Basaltic Volcanic Activity (in the South), Basanites & Olivine Nephelinites (Marginal Zones)	Andesite & Rhyodacite Flows & Breccias Quartz Latites and Rhyolites Ash-Flow Tuffs Trachytes & Phonolites
Source	Meyer et al., 1974 Morbideilli et al., 1974	Lippolt et al., 1972 Barberi et al., 1982	Kudo et al., 1971 Lipman et al., 1972

(Gouin, 1976, 1979). The fault plane solutions for earthquakes of this region show the tensional axis to be normal to the strike of the rift.

Seismicity in the Ethiopian rift is mostly associated with an 8-km-wide zone of intense faulting which follows the axis of the rift. This zone, which is characterized by large earthquakes, coincides with the youngest volcanic centres which occur along the length of the rift (Searle and Gouin, 1972). Microseismicity occurs along the rift floor although at a low level of activity (Molnar et al., 1970).

Seismicity of an extensional nature occurs also in the Baikal rift zone. More than 2000 earthquakes of magnitude larger than 1.5-2 are recorded every year in this region (Fairhead and Stuart, 1982). The epicentres of small earthquakes follow linear belts which usually strike parallel to the main rift structures. The seismicity is associated with volcanic eruptions as well as with vertical fault movements (Solonenko, 1978). The foci of these earthquakes, which occur both as individual shocks and in swarms, seem to be concentrated at shallow depths in the upper crust (Golenetsky and Misharina, 1978).

The Rhine graben, which is characterized by recent tectonic activity, is marked by a patchy distribution of epicenters with maximum activity at a depth of about 10 km (Mueller, 1970). The seismicity is of an extensional nature (Aherner, 1970) and the epicenters seem to be confined to the upper 25 km of the crust (Schneider, 1970).

The Rio Grande rift exhibits earthquake and microearthquake activity (Hoffman, 1975). Historical records since 1849 indicate that the seismicity is concentrated along the rift (Cordell, 1978) and includes earthquakes of up to magnitude of about 6.3 (Hoffman, 1975). However at present the rift seems to be relatively quiet.

Earthquake and microearthquake activities are typical of

continental rift systems. Seismicity, which is extensional in nature with the relative tension axis perpendicular to the strike of the rift, appears to take place mainly along the rift axis, both as individual shocks and as swarms, and to coincide with the young volcanic centres. Earthquake foci are concentrated at shallow depths and are generally confined to the crust beneath the rift zones.

1.4 BODY-WAVE AND SURFACE-WAVE DATA

An anomalous upper mantle with P-wave velocity of 7.2 to 7.6 km s^{-1} * was detected by Searle (1974) under the Afar region. The depth to the top of this anomalous mantle layer ranges from 26 km in southern Afar to 16 km in northern Afar (Ruegg and LePine, 1973). The thickness of this layer is estimated to range from 15 to 40 km (Berckhemer et al., 1974).

P-wave travel-time anomalies in the upper mantle beneath the Ethiopian rift were detected by Makris et al. (1970). They interpreted this to be caused by a volume of low velocity material in the uppermost mantle. This anomalous upper mantle, with a velocity of 7.7 to 7.8 km s^{-1} , is estimated to extend to a depth of 150 km beneath the rift.

Large scale seismic refraction experiments indicate that the uppermost mantle, whose depth varies from 35 to 45 km along strike, underneath the Baikal rift is characterized by an anomalously low velocity of 7.6 to 7.7 km s^{-1} (Olsen, 1983) (Fig. 1.2a). This layer, which has an average thickness of around 17 km spans over the entire

* "Normal" upper mantle velocity (just below the Moho) is about 8.0 km s^{-1} .

Fig. 1.2: Seismic velocity profiles for a) Baikal rift, b) Rhine Graben, and c) Rio Grande rift. a after Puzrev et al. (1978), b and c after Olsen (1983).

region (Puzyrev et al., 1978) and is suggested to be due to the penetration of mantle material into the lower crust (Logatchev et al., 1983) and/or to the increase in temperature and partial melting in the uppermost mantle (Zorin, 1981). In the first case the extent of the penetration has been suggested to be as shallow as 20 km or less (Gornostayev, 1972).

Seismic surface wave studies support the presence of a low velocity layer at a depth of only 10 km beneath the Rhine graben (Mueller, 1970). The existence of this low velocity layer is also inferred from a decrease in P-wave velocity of some 7% and high values of heat flow (Scheelke, 1974). In the southern part of the graben this layer is estimated to commence at a depth of about 25 km and is characterized by a velocity of 7.6 to 7.7 km s⁻¹ (Fig. 1.2 b).

The regional seismic refraction data indicate that the mantle (at a depth of about 40 km) has a lower than normal P-wave velocity of 7.6 km s⁻¹ beneath the Rio-Grande rift (Olsen, 1983) (Fig. 1.2 c). The reflected P and S-wave phases from local micro-earthquakes indicate that this anomalous layer starts at a depth of 19 to 20 km in the Socorro region. This is interpreted to be due to the presence of a substantially molten layer (magma body) with a thickness of only a few kilometers beneath this part of the rift (Rinehart et al., 1979). A similar layer is also detected beneath Santa Fe at a depth of 10 to 17 km and beneath El Paso at a depth of about 21 to 28 km (Pedersen and Hermance, 1978).

Explosion as well as earthquake seismology reveals the common features of the upper mantle under continental rift zones. The anomalous low velocity in the uppermost mantle seems to extend beneath the whole region of rifting. This may indicate that geodynamic

processes as well as the physical state of the upper mantle are similar under different continental rift systems.

1.5 GRAVITY

The gravity field in Afar is affected in the southern and western parts by the Somali and Ethiopian plateaus, respectively. These plateaus are characterized by large negative Bouguer anomalies and their corresponding isogals closely follow the topographic features towards the rift. For the Somalian plateau the anomaly is -160 mgals whereas for the Ethiopian plateau it reaches a level of -220 mgals (Makris et al., 1972). The border of the Ethiopian plateau towards the rift is therefore characterized by a larger negative Bouguer anomaly.

Within the main Ethiopian rift valley a positive Bouguer anomaly of amplitude 60 mgals characterizes the whole rift floor. This anomaly, however, is superimposed on a broad negative anomaly of the Ethiopian plateau of amplitude -220 mgals (Searle and Gouin, 1972).

In the Baikal rift, sharp minima of the gravity field are associated with the rift valley (Artemjev and Artyushkov, 1971). Lake Baikal is characterized by a very pronounced minimum which is evident even after the thickness of the sediments is accounted for. The negative Bouguer gravity anomaly is bounded by relative maxima along the surrounding shoulders (Puzyrev et al., 1978). Although a positive gravity anomaly associated with the floor of the rift is not evident from the Bouguer anomaly data, this gravity high can be seen when correction is made for a 6 km-thick slab of sediments in the rift floor (Zorin, 1981).

In the Rhine graben a steep negative Bouguer gravity anomaly of

amplitude -55 mgals (Goguel, 1957) is observed over the rift valley and is associated with a chain of gravity minima which roughly follows the axis of the rift (Mueller and Peterschmitt, 1966). The axis of the negative Bouguer anomaly also coincides with the region of highest seismic energy release (Mueller and Rybach, 1974).

In the Rio Grande rift a very broad negative anomaly is associated with the uplift over the whole region (Decker and Smithson, 1975). Superimposed on this anomaly is a positive gravity anomaly (of about +30 mgals amplitude) along the axis of the rift (Cordell, 1978). The general feature of the Bouguer anomaly profile in this region is that the uplifts are associated with relatively high gravity values whereas the intervening basins are associated with low values.

The examination of the gravity data from the continental rifts indicates that: (a) the uplifted regions of the rifts are in approximate isostatic equilibrium; (b) these rifts are associated with a broad negative Bouguer anomaly coincident with the areas of regional uplift; and (c) superimposed on the anomaly is sometimes a narrow, short wavelength, positive Bouguer anomaly associated with the rift floors (Fig. 1.3). In most of these rifts the negative anomaly is attributed to the thinning of the lithosphere and to the presence of low density asthenosphere beneath the rift zone. The positive anomaly has been explained as due to the presence of a dense intrusive zone penetrating the upper crust and feeding the volcanic centres of the rifts.

The existence of an anomalous zone of lower than normal P-wave velocity in the uppermost mantle underneath the rift zones supports the hypothesis that the temperature is anomalously high, or even that some

Fig. 1.3: Schematic cross section of an idealized continental rift with the corresponding Bouguer anomaly profile. After Ramberg and Morgan (1984), slightly modified

melt fraction is present under the rift systems, a conclusion substantiated by gravity data.

1.6 HEAT FLOW AND TEMPERATURE

The electrical conductivity is directly related to temperature. Therefore, high conductivities under the rift systems are attributed to high temperatures beneath them.

The presence of a low density, low velocity "pillow" of material in the uppermost mantle beneath the entire Afar rift was detected by Makris et al. (1972). Their study suggested that the resistivity ranges from 200-500 Ωm up to a depth of 15 km where it suddenly drops to 10-50 Ωm . Low density, low velocity and low resistivity are usually associated with high temperature. It appears therefore that this high temperature zone begins at a depth of about 15 km. The area of low resistivity extends towards the Ethiopian rift indicating an anomalously high temperature (800-1000 $^{\circ}\text{C}$ at a depth of about 15 km, i.e. a temperature gradient of 60 $^{\circ}\text{C}/\text{km}$, Berkthold et al., 1975). Such a temperature is compatible with partial melting of material at this depth. From this Makris et al. (1972) concluded that the whole width of the main Ethiopian rift is underlain by a shallow partially molten intrusion. The heat flow of this region is considered to be fairly similar to that of Iceland which is about 210-250 mW/m^2 (Berkthold et al., 1975).

Lubimova et al. (1972) measured a high heat flow value of 84-143 mW/m^2 in the region of the Baikal rift. The zone of high heat flow is confined to a narrow band about 60 km wide along the axis of the rift. The high temperature zone which starts at a depth of 12-15 km and has a

thickness of less than 30 km, displays high electrical conductivity which was detected by Hermance (1982) through magnetotelluric deep sounding. It is also characterized by a low velocity layer at 15-17 km (Berdichevsky et al., 1980). It has been argued that the electrical anomaly may be associated with the accumulation of a melt fraction derived either by partial melting in situ or by segregation and upward migration of melt from deeper zones. All these characteristics are compatible with the presence of a pillow of low velocity material beneath the rift (Zorin and Osokima, 1984).

High values of heat flow (average $109 \pm 35 \text{ mW/m}^2$) in the Rhine graben, detected by Meincke et al. (1967) and Mueller and Rybach (1974), coincide with the extent of a low velocity layer at a depth of 10 km beneath the graben (Mueller, 1970). This finding is consistent with the magnetotelluric and geomagnetic studies of Fuchs (1974). Furthermore, Haak et al. (1970) determined a depth of 25 km for an anomalously conducting layer beneath the axis of the rift. Losecke (1970) and Meissner et al. (1970) suggest that this layer is associated with a pillow of low velocity material in the upper mantle.

The Rio Grande rift is also characterized by a belt of unusually high heat flow ($108 \pm 27 \text{ mW/m}^2$) (Reiter et al., 1979; Sass et al., 1976). The existence of the zone of lower than normal P-wave velocity beneath the rift (Brown et al., 1980) is consistent with the existence of a conducting layer at a depth of 21 km found by Jiracek et al. (1983) using magnetotelluric measurements. This conducting layer seems to extend to a depth larger than 100 km.

From the above review it is evident that continental rift systems are distinguished by high values of heat flow (Table 1.2).

Table 1.2: Heat Flow Data

REGION	HF (mW/m^2)	Source
Afar (very similar to Iceland)	210 - 250	Makris et al., 1972 Berkold et al., 1975
Baikal	84 - 143	Lubimova et al., 1972 Hernandez, 1982
Rhine	109 $\pm$ 35	Meincke et al., 1967 Mueller & Ryback, 1974 Bram, 1979
Rio Grande	108 $\pm$ 27	Decker & Smithson, 1975 Reiter et al., 1979

Furthermore, the presence of a low velocity pillow of mantle material at the base of the thinned crust appears to be typical of continental rifts (Neugebauer, 1983), although its extent varies. In the Baikal rift and Rhine graben this pillow has a thickness of only 15-25 km whereas below the Rio Grande rift it extends to a depth of more than 100 km. This low-velocity layer coincides with a region of high conductivity detected under the rift systems. The most widely accepted explanation for the low P-wave velocity and high conductivity is partial melting, i.e. the presence of a small melt fraction under the rift systems.

In conclusion, the geophysical and geochemical properties of rifts are similar: the volcanism is mildly to strongly alkalic; the seismicity, occurring as both micro and macro-earthquakes, is shallow and of an extensional type with the tensional axis normal to the strike of the rift; the presence of a subcrustal low velocity zone is typical; the gravity is distinguished by a broad negative Bouguer anomaly; and heat flow is high. Table 1.3 notes the common characteristic features of continental rift systems. Given their similar geophysical and geochemical signatures, the rift systems are bound to have developed through a similar sequence of processes. In the next section the tectonic hypotheses of rift formation and evolution are summarized.

1.7 TECTONIC HYPOTHESES ON RIFT DEVELOPMENT

Contradicting the previously widespread notion that rifts are compressional features, Gregory (1921), Willis (1938), and Cloos (1939) were among the first to suggest that the major features of the rifts have been produced by extension of the Earth's crust. It is now

Table 1.3
Common Characteristic Features of Major Continental Rift Systems

1. Structural and Geomorphic Features

- a. Width generally 35 to 60 km
 - Baikal 50 km
 - EARS 60 km
 - Rhine 36 km
- b. Length of the order of 1000 km
 - Baikal 2500 km
 - Afar } 4000 km
 - Ethiopian }
 - EARS }
 - Rio Grande 850 km
 - Rhine graben 300 km, the central segment of a continuous system of rifts.
- c. Topographic relief < 2.5 km
 - Baikal 2200 m
 - Ethiopian 1700 m
 - EARS 2000 m
 - Rio Grande 1100 m
 - Rhine graben 2200 m
- d. Existence of extensional features, normal faulting and dikes. Amount of horizontal stretching:
 - Baikal 10 km max.
 - EARS: Kenya 8-15 km
 - Ethiopian 40 km in the N, 25 km in the S
 - Rio Grande 50-70 km
- e. Thin crust, 35 km or less, and thin lithosphere
- f. Broad early stage sedimentary basins
- g. Broad upwelling (sometimes)

2. Magmatic Features for both wet and dry rifts.

- a. Volume of volcanism (since Eocene times)
 - Baikal (dry) 5000 km³
 - Ethiopian (wet) 345000
 - EARS eastern (wet) 144000 } 500000
 - western (dry) 11000 }
 - Rio Grande (dry) 12000
 - Rhine graben (dry) ~9000
- b. Volcanic rocks of diverse composition, presence of bimodal activity in terms of SiO₂ content.

3. Geophysical Features

- a. P_n velocities less than 7.8 km s⁻¹
- b. Crustal low-velocity layers
- c. Teleseismic P-wave delay
- d. Long wavelength Bouguer gravity low, sometimes associated with short wavelength positive anomaly along the axis of the rift, e.g. Baikal, Ethiopian, eastern branch of EARS and Rio Grande rift
- e. High surface heat flow
- f. Shallow seismicity, aligned with the rift axis
- g. Extensional tectonics inferred from focal mechanisms

established that rifting occurs by normal faulting in response to horizontal deviatoric tension affecting the crust.

1.7.1 Regional Tectonic Tension

This hypothesis (suggested by Vening Meinesz, 1958) implies that, when the lithosphere is subjected to tensile stress, the brittle layer of the upper crust responds by brittle extension while the underlying ductile part of the lithosphere, which includes the lower crust and topmost mantle, deforms by viscoelastic creep (Bott, 1976). The deviatoric stresses in the ductile layer are dissipated by creep, resulting in an increase of tensile stress in the overlying brittle layer (Kusznir and Bott, 1977). Then, when the elastic limit is exceeded, the brittle layer fails in shear, developing a normal fault. The bending stress in the elastic part of the lithosphere subsequently causes the second fault on the down throw side, where bending is at its maximum, or along a pre-existing line of weakness. The tension in the lithosphere can originate from membrane stresses, stresses due to uparching, or stresses associated with other plate tectonic processes.

1.7.2 Membrane Tectonics

Turcotte (1974) showed theoretically that as the plates move towards the Equator, compression will increase at their margin and tension will be created in their central part (the reverse applies to the plates moving away from the Equator). This is because as the plates change latitude their shape have to conform to the varying curvature of the ellipsoid between the pole and the Equator. The change in shape

will create stresses within the plates which may eventually result in their fracturing. The magnitude of the stress created by this process depends on the initial and final position of the plate.

This hypothesis has been applied to the African plate in order to explain the generation of EARS (Oxburgh and Turcotte, 1974). It has been shown that if a plate as large as the African plate undergoes large latitudinal change, stresses of the order of hundreds of MPa will be created within it. This magnitude of stress is of the same order as the strength of the plate and is sufficient to rupture it. In its application to EARS this hypothesis is consistent with the paleomagnetic studies of McElhinny (1973) which demonstrate that between Miocene and Present, the African plate has undergone a northward latitude change of about 10 degrees. The major criticism to membrane tectonics is that membrane stresses are "non-renewable", i.e. although they might have initiated rifting, they decay and dissipate with brittle and ductile deformation and are not renewed.

1.7.3 Stresses Associated with Uplifted Continental Regions

The horizontal deviatoric tension necessary for the formation of rifts may result from the bending stresses in the elastic part of the lithosphere which are caused by uparching. However, the strain associated with such uparching has been shown to be inadequate for the graben subsidence (Artemjev and Artyushkov, 1971) since the amplitude of uplift due to bending alone is not enough to account for more than about 3% of the extensional strain (Mulugeta, 1985).

The other type of stress system which may contribute to rift

formation is the one associated with uplifted continental regions which are in isostatic equilibrium. This stress system is attributable to the uplifted topography and upthrust resulting from the anomalously hot, low density mantle region. The stresses caused by crustal thickness variation between the uplifted regions and non-uplifted terrain were originally analysed by Bott (1971a;b) and Artyushkov (1973; 1981).

According to Bott (1982b), if the whole system behaves as an elastic solid, the horizontal stress difference between the uplifted and non-uplifted regions could reach up to 50 MPa if the uplift is assumed to be 2 km. This stress system, which persists as long as the topography and the low density regions are present, seems to be too small to account for the formation of a graben. However, if the upper 10 to 20 km of the crust behaves elastically and the lower part deforms by creep, the stress differences in the uppermost crust may be significantly increased by relaxation in the lower lithosphere. The finite element analysis of Bott and Kusznir (1984) indicated that, the thinner the elastic layer, the greater the stress differences in it.

Since the temperatures within and below the lithosphere are generally above average underneath the uplifted regions, the elastic layer could have a thickness of only 10 to 15 km. In this case stress differences of about 400 MPa can develop in the upper crust. This difference is more than enough to account for observed graben formation (Bott, 1981). Therefore, it seems that uplifted regions can cause large enough stresses to form the rift systems. Two mechanisms are suggested to generate the uplift: generalized mantle convection and isolated mantle plumes.

1.7.4 Generalized Mantle Convection

In this context, both uplift and volcanism are considered to be related to the thermal state of the upper mantle. In other words, anomalously high temperature in the mantle causes expansion which results in the uplift of the lithosphere and magmatism succeeding it. This model was proposed by Gass (1970) who considered mantle convection to be the driving force. It has been supported by several researchers (Spohn and Schubert, 1983; Yuen and Fleitout, 1985; Fleitout et al., 1986) on the basis of geophysical data including gravimetric and seismic observations which indicate the presence of a low velocity zone under the rifts. This zone has been interpreted by these authors as a subcrustal swell of mantle material.

Bailey (1974) objects to this model on the ground that rifting and volcanic activity have persisted for a long time (for instance since the Cretaceous in east Africa) and that the model does not account for the source of heat in these long, narrow belts. Furthermore, this model appears to require a very complex pattern of convection cells in the mantle. Sowerbutts' (1969) objection to this model is based on gravity modelling and heat flow data. It is argued that mantle convection alone is not enough as the heating agent to account for the high heat flow values in the rift systems. Sowerbutts (1969) and Bailey (1974) contend that a mechanism of heat generation and migration by extension is required and therefore suggest that the volcanism in the rift zones is caused by volatile transport from the lower mantle in which case no complex pattern of mantle convections below the rift zones would be needed. The role of volatiles is to reduce the solidus of the rocks of the upper mantle and provide the

area not only with additional heat (as the pressure is released in the rising material) but also with mobile elements, especially the alkalis.

1.7.5 Mantle Plumes

According to this model, proposed by Morgan (1971, 1972a and b), mantle plumes, which are pipe-like zones of rapidly ascending mantle material, originate at or close to the core/mantle boundary (according to some, e.g. McKenzie, 1984, at the mantle transition zone) and rise to the overlying lithosphere causing warping and magmatism in the rift zones. Burke and Whiteman (1973) suggest that partial melting of mantle material occurs above the rising diapir and forms large volumes of alkali basalt magma at the base of the lithosphere. This large volume of magma causes a mass deficiency whose isostatic response will be the elongated domal uplifts seen in the rift zones. This model can explain the zones of extensive magmatic activity far from the plate margins (Oxburgh, 1974), and can also account for the volcanic zones along which the age of the volcanism and volcanic activity have changed regularly with time. Fig. 1.4 indicates the stages of development of the continental rift system according to this hypothesis.

On the basis of theoretical and experimental fracture studies on glass bottles by a controlled hydraulic technique, Bahat (1984) contends that the fracture systems of the rift valleys were initiated by upward movement of mantle diapirism. Diapiric intrusions result in a stress field which forms deep regional tensional fractures of brittle type through the crust. These fractures propagate and cause major and minor bifurcations. The major bifurcations are responsible for the

Fig. 1.4: One possible sequence of rift development: a) formation of hot spots; b) heating and thinning of the continental lithosphere, development of tensile stress system in the upper crust; c) graben formation starts when the stress is sufficiently large. After Bott (1981).

development of the continental rift systems and the minor bifurcations form the small horst and graben structure. In this model the magmatic activity along the fractures causes vertical uplifts and subsidences which result in block faulting.

One argument against the plume model is that the shear stresses produced by the drag of the plumes on the base of the lithosphere are not likely to be much more than 1 MPa, too low in comparison with the strength of the lithosphere (Parmentier et al., 1975). This argument can be rejected on the basis of the phenomenon of stress concentration in the upper elastic lithosphere and the stresses due to topography. This stress concentration is inversely proportional to the thickness of the elastic layer which in turn depends mainly on the local geothermal gradient (Bott, 1982a). Since the geothermal gradient is high under the rift zones and lithospheric thinning occurs in the area, it may be plausibly argued that high tensile stresses could develop in the upper crust in regions of rifting.

1.8 DISCUSSION

With regard to the previous discussion, several points are worth mentioning. The intersection of the eastern and western branches of EARS (at Mbeya, Tanzania) is characterized by a concentration of carbonatites and alkali intrusions which range in age from Precambrian to Tertiary (Bailey, 1961). The existence of these rocks testifies to the fact that EARS had started to form long before the major Tertiary-Recent volcanic activity. Bahat (1984) concludes that in EARS the Tertiary volcanism occurred along a pre-existing brittle fracture system which was probably formed in the Precambrian due to upward mantle diapirism.

The fact that the continental rift systems, specially EARS, might have had an early origin was first noted by Teale (1936). The majority of continental rift systems formed in Precambrian platforms. The parallelism of the rift faults with Precambrian trends along with the parallelism of the major basins with the rift systems indicate that later faulting has followed the older zones of weakness in the lithosphere. In other words, the Cenozoic rift faults are the latest rejuvenation of the Precambrian faulting patterns (McConnell, 1972). These ancient deep faults, or "perennial deep lineaments" (McConnell, 1974) are indicators of a persistent pattern of zones of weakness in the lithosphere.

When discussing the persistence of activity along these ancient deep lineaments, it must not be forgotten that the activity has taken place within a lithosphere which has moved relative to the underlying mantle. If these deep lineaments are the expressions of older zones of weakness which penetrate the major part of the lithosphere and are not restricted to the thin brittle part of it, they could accomodate lithospheric movement in a variety of directions (Watterson, 1975). These lineaments are periodically reactivated and are localized heat flow channels and/or channels for penetration of underlying molten material into the zone of weakness.

In general, geophysical data from the rift zones substantiate the existence of an asthenospheric high beneath the thinned lithosphere of the rifts. This asthenospheric high induces an arched uplift by thermal expansion and isostasy, together with a concentration of volatile flux, heating, partial melting, and magmatism in the rift zones.

Although from the above arguments it appears that the simple sequence of domal uplift, volcanism, and crustal rifting is generally followed in the development of the rifts, the story is likely to be more complex. Most of the major continental rifts are associated with crustal doming or plateau uplift (for instance, the Baikal, EARS, and the Rhine graben). But doming as the first process in the formation of the rifts is not the general rule, since not all the rift systems are formed in uplifted regions; moreover, crustal attenuation during rifting causes initial subsidence which will be followed by further subsidence due to thermal relaxation of the lithosphere and sediment loading in the post-rifting period (Ramberg and Morgan, 1984).

Study of the timing of the domal uplift and rifting from the thickening of early sedimentary and volcanic formations in EARS convinced Baker (1972) that in the initial stages the rift zone was subsiding rather than uplifting. In addition Williams (1978a) noted that in the Ethiopian rift there has been a very small magnitude of domal uplift if any. This is in accordance with King's (1978) interpretation that in the Kenya rift there has been a broad downwarping followed by domal uplift. A broad shallow basin resulting from the downwarping of the crust as a precursor to uplift and rifting is also noticed for the Rio Grande rift by Chapin (1979). Furthermore, the same kind of broad gentle crustal subsidence has been detected by Illies (1978) in the Rhine graben and by Logatchev (1978) in the Baikal rift.

Although the crustal subsidence has been observed in many major rifts of the world, it does not necessarily follow that all rift systems are similarly characterized by subsidence as a precursor to

rifting (Ramberg and Morgan, 1984).

Experimental and numerical modelling of rifts has provided important information about the dynamics of rifting (see, for example, Bridwell and Potzick, 1981; Ramberg, 1981; Lynch, 1983; Morgan and Baker, 1983; and Mulugeta, 1985). These studies show that the rifting is caused by the stretching of the crust above the rising and laterally spreading low density anomalous body. The underlying cause of this uprising is the lateral difference in density with depth. The upwelling and spreading of the anomalous body causes stresses in the lithosphere which are directly related to the formation of the rift systems.

CHAPTER 2

THE EAST AFRICAN RIFT SYSTEM

2.1. GENERAL GEOLOGY

The east African rift system (including the Ethiopian rift) extends about 4000 km in a south-southwesterly direction down the eastern side of Africa from the Gulf of Aden and Gulf of Zula (Red Sea) to the Zambezi River at 18°S (Gregory, 1921; Dixey, 1956; McConnell, 1967). This rift system is totally intracontinental (except at the northernmost end) and was formed in the ancient African Precambrian platform of about 3 Ga of age (McConnell, 1972). It is characterized by a system of normal faults enclosing trough-like features of 40 to 65 km width (Baker et al., 1972) (Fig. 2.1).

Just north of latitude 10°S EARS separates into two clearly distinguishable branches. The eastern branch starts from the Mbeya junction (junction of the eastern and western branch in southern Tanzania) and extends northward through the Gregory or Kenya and the Ethiopian rifts. The western branch extends also northward from this junction, bends around the Tanzanian (Tanganyika) shield and abruptly terminates against the northwest trending Aswa mylonite zone of northern Uganda which is a narrow, linear zone of cataclasites about 150 km long (Whiteman, 1971) (Fig. 2.2).

The two branches of EARS, the eastern traversing the Kenyan and Ethiopian domes and the western traversing a region of uplifted and locally warped plateaus, surround the Lake Victoria block (Tanzania, Uganda, west Kenya, Rwanda, and Burundi; Hermance, 1982) and join to

Fig. 2.1: The East African Rift System. The stippled areas show the extent of the rift valleys. After McConnell (1972) and Mohr (1982).

Fig. 2.2: Structural sketch map of the east African rift system. After King (1970), slightly modified. The stippled areas indicate the volcanic fields.

form the Malawi rift to the south (Barberi et al., 1982) (Fig. 2.2). The east African plateau is located approximately between 5°N , 10°S and 24°E , 42°E , and has formed during several stages of regional uplift since the Jurassic (Dixey, 1956; Saggerson and Baker, 1965).

The Kenya dome or Kenya arched uplift, between 2°N to 2°S , is an elliptical highland region 100 km long and 350-400 km wide with an elevation of 3.5-4 km. It was initiated in Miocene time prior to the formation of the rift valley (Banks and Swain, 1978).

On the basis of its structural history the Precambrian crust of east Africa has been divided into five zones. They are, from oldest to youngest: Tanzanian shield (> 3000 Ma), Dodoman (2700-2300 Ma), Ubendian (2500-1650 Ma), Karagwe-Ankole (1400-1300 Ma), and the Mozambique belt (1100-450 Ma) (Fig. 2.3 and Table 2.1).

The upfaulted block of the Precambrian crust between the two branches of the rift is a craton, the Tanzanian shield, which is older than 3000 Ma (McConnell, 1972) (Fig. 2.3). This shield has played an important role in the bifurcation of the system into two branches. In other words, the presence of this stable block has caused the deflection of the fracture systems around it giving rise to the formation of the eastern and western rifts (Mohr, 1974). Moreover, it appears that on the regional scale, rift faults are frequently parallel to the Precambrian structures, whose significance will be the subject of the next section.

2.2 THE PATTERN OF RIFT FAULTING IN RELATION TO PRECAMBRIAN STRUCTURES

McConnell (1951) and Dixey (1956) were the first to stress the correlation between fault directions and older Precambrian structural

Fig. 2.3: Map of the east African rift system showing Precambrian structures. Compiled from the International Tectonic Map of Africa (1968) and McConnell (1972).

Table 2.1: Name and Age of the Orogenies Associated with the Precambrian Structures in EARS (McConnell, 1972)

Major Precambrian Assemblages	Name of Orogeny	Age (Ma)
Mozambique Belt	Katangan	500-750
Irumide, Kibara, Burundi, Karagwe-Ankole	Kibaran	800-1300
Ubendian	Ubendian	1600-2000
Dodoman, Nyanzian, Kavirondian	Archean	2300-2700
Tanganyika	Primitive	3000

Table 2.2: The Volcanic History of the Western Rift

Age	Rungwe	South Kivu	Birunga	Toro-Ankole
Recent	Phonolite	K-Quartz	Potassic	Mainly tuffs
	Trachyte	porphyry	intermediate	and potassic
		K-rhyolite	and basic	ultrabasic
			lavas	lavas
Pleistocene		Trachyte	Plateau	
		Basalts	basalts	
	Basalts, Trachyte and Phonolite lavas and tuffs	Basalts		Sources: King and Sutherland (1960); Shackelton (1978); Barberi et al. (1982)

trends (Fig. 2.4). The rift structures and the Cenozoic faulting are related to old tectonic lines which are Precambrian in origin; that is, the pattern of rift faulting appears to be controlled by lines of mechanical weakness which have been present since the Precambrian. Tectonic stresses have given rise, in successive intervals, to the rejuvenation of fractures which had first been established as "primordial lines of weakness" (Dixey, 1956).

The parallelism of the rift and the Precambrian structures can be seen in large parts of the eastern rift in Kenya and in the western rift in the Lake Tanganyika-Lake Rukwa-northern Lake Malawi sector. In this section of the western rift (latitude 9°S , longitude 33.5°E) the rift trends NW-SE. McConnell (1967) noticed that these remarkably linear fault patterns are matched by the parallel NW-SE trending Mbendide structures whose intense deformation is indicated by their cataclastic character.

The relationship between Precambrian and rift structures can be interpreted in two ways: (a) The pattern of rift faults is a manifestation of major Precambrian structures, i.e. a single major tectonic style has continued to operate since Precambrian times (McConnell, 1972); or (b) the rift faults have followed pre-existing tectonic lines of weakness, i.e. the overall pattern of the rift system has developed through different structural events superimposed on the pre-existing ones (King, 1970; Baker et al., 1972).

The major Precambrian structures which mark the zones of mechanical weakness could have remained intermittently active over time and have been the locus of the major faulting of the rift zone in east Africa. As mentioned earlier, (see Chapter 1), if these zones of mechanical weakness affect the major part of the thickness of the

Fig. 2.4: Structural map of east Africa based on mapping from ERTS-1 imagery. After Mohr (1974). The dotted lines indicate the structural lineaments.

lithosphere, they could also persist over time taking into account the lithospheric movements (Watterson, 1975).

In several sectors of the eastern and the western rifts, however, rift structures are not parallel to Precambrian structures. In the eastern rift the detailed ground mapping of Baker (1963) revealed that east of the Gregory rift in north-central Kenya between latitude 2°N and the Equator, the Precambrian trends N-NNW whereas the rift faulting immediately to the west trends NNE. It should, furthermore, be noted that the Kavirondo rift within the Lake Victoria block trends ENE-NE, almost at right angles to the Precambrian structures.

In the western rift, although the central sector of the Tanzanian rift is strongly controlled by the NW-SE trending Ubendide structures, from latitude 6°S northward the rift follows the meridional trends of the younger Precambrian belts. In southwest Uganda the folded belt of the Karagwe-Ankolean system runs NW-SE which is almost at right angles to the rift.

The Precambrian basement of east Africa is characterized by many ancient lines of structural weakness. However, although at different periods the pre-existing fractures were rejuvenated, new fractures were also formed. This phenomenon is shown by the existence of the non-parallel structures in the rift.

The parallelism of the rift and the Precambrian structures does not prove that rifting began in the Precambrian times. It does, however, indicate that the rifting has sometimes taken advantage of and followed zones of weakness established by Precambrian tectonic patterns. This is further supported by the repetition of alkaline magmatism from deeper sources in the rift zone (Bailey, 1983).

2.3 TECTONIC HISTORY OF EARS

Since the final breakup of Gondwana (Late Triassic), the whole of Africa has been subjected to a complex sequence of uplift, rifting and alkaline volcanism. Several workers have tried to unravel the history of the development of EARS using the sedimentary record, erosion surfaces, structure, and isotopic data (see for instance Shackleton, 1978; Barberi et al., 1982; Mohr, 1982).

In the Late Mesozoic, east Africa was a lowland area; it was unevenly uplifted in Late Cretaceous (100-65 Ma) (Saggerson and Baker, 1965). This Late Cretaceous uplift which is evidenced in preserved planar erosion surfaces marks the first stage of the rifting process. In the first part of the Tertiary (65-20 Ma) east Africa was reduced to a peneplain (Shackleton, 1978), and basalts were emplaced in northwest Kenya and southwest Ethiopia at around 30 Ma (Baker et al., 1971).

In Early to Mid-Miocene (20-15 Ma), the subsidence of the Turkana depression, a triangular lowland in Northwestern Kenya 200 km wide in the north narrowing southward (Baker et al., 1972), and the uplift of the Uganda-Kenya border (maximum 300m in central Kenya) were followed by massive eruption of basalts and rhyolites in the Turkana area (Logatchev et al., 1972). In the uplifted areas, the new isolated volcanoes of eastern Uganda and Kavirando rift were formed. During this period the eastern rift (Kenya rift) was magmatically quiescent. The western rift began to form in early Miocene when it was a shallow depression and subsiding (Bishop, 1965). This stage is indicated by the oldest sediments of the Lake Albert (Hopwood and Lepersonne, 1953).

In Late Miocene (15-11 Ma) much of the activity was concentrated in the eastern rift with which were associated a gentle regional uplift

(maximum 700 m in the west and 300 m in the east) and the eruption of flood phonolites and phonolitic trachytes (Baker et al., 1971). Extensive rift faulting followed at the end of Miocene. The Late Miocene-Early Pliocene (11-6 Ma) is characterized by downfaulting, shoulder uplift and sedimentation which were followed by eruption of volcanics.

In Early to Middle Pliocene (6-4 Ma) the volcanic activity was primarily concentrated in the central part of the eastern rift. This volcanic activity underwent a substantial change in composition and became increasingly more variable (trachytes, alkali trachytes and phonolites). The end of this stage was associated with the formation of major boundary faults and the volcanic activity extended southward.

In Late Pliocene to Early Pleistocene (4-1.8 Ma), massive eruptions and sedimentation covered a large part of the eastern rift. Regional uplift (maximum 1500 m in central Kenya) (Baker et al., 1972) and the main faulting which yielded the present configuration of the eastern rift occurred in this period (Mohr, 1982). The major uplift of the Kenya dome occurred at this time. Fig. 2.5 indicates the broad shape and height of the uplift from the isopachs of sub-Miocene erosion surfaces. New faults were initiated within the Turkana depression which caused further deepening of this trough (Baker, 1972). At this time the grabens of the western rift were formed. The beginning of this stage is distinguished by the most reliable stratigraphic marker, the Kaiso sediments of Villafranchian age (3.2-0.9 Ma) (Logatchev et al., 1972). The volcanic activity in the Rungwe area, which is characterized mostly by basaltic, trachytic and phonolitic lavas and tuffs, began in Late Pliocene, marking the first eruption in the western branch of EARS.

Fig. 2.5: Broad shape of the uplift of the Kenya dome based on the sub-Miocene erosion surfaces. Heights are in meters. After Baker et al. (1972).

Early to Middle Pleistocene times (1.8-0.7 Ma) are characterized by weakly alkaline basaltic volcanism which was erupted on the floor of the eastern rift and by the development of new faults (Mohr, 1982). This period of intense faulting gave rise to the swarms of minor faults of the rift floor in the eastern rift valley and local rejuvenation of the marginal faults. Volcanoes were also generated along the axis of the rift (Shackleton, 1978). A major episode of faulting occurred in the western rift valley postdating the erosion surfaces of the rift (Barberi et al., 1982). This major faulting was associated with the eruption of basalts and trachytes in the South Kivu field as well as with plateau basalts in the Birunga field starting in the Middle Pleistocene.

From the Middle Pleistocene to the Present (0.7-0 Ma), grid faulting (swarms of closely spaced sub-parallel minor faults; Fairhead, 1976) on the rift floor of the eastern branch was followed by strongly alkaline-carbonatite magmatism. The grid faulting, which mainly occurred at around 0.5 Ma (Burke and Whiteman, 1973) has been followed by volcanism up to the present. In the western branch, the volcanic activity in the Toro-Ankole field started at this time and, as with the activity of other volcanic fields, has continued up to the Present (King, 1970) (Table 2.2). From the above it appears that the direction of propagation of rifting and volcanicity is from N to S in the eastern rift and from S to N in the western rift.

2.4 MAJOR FAULTING AND RIFT STRUCTURES

Two different systems of faultings dominate EARS, block faulting and rift faulting. Block faulting is major faulting which has produced

differential uplift and depression of blocks of different shapes and sizes. This kind of faulting, which is sometimes associated with tilting or warping, occurs in different directions and has been activated intermittently since Precambrian times. This broad zone of block faulting includes certain zones in which sub-parallel to parallel faults are concentrated. This is called rift faulting and has developed in Late Cenozoic times.

2.4.1 The Eastern Rift Valley

The eastern rift valley is characterized by numerous sub-parallel normal faults and by the small-scale horst and graben faulting of its floor. At the End of Miocene the faulting of the phonolite plateau (formed as a result of eruption in Mid- to Late Miocene) was initiated. This faulting period gave rise to the Turkana basin and a shallow rift valley was formed. The largest faults were formed on the western side of the rift valley but monoclinical flexuring was more important on the eastern side. Although the rift was stable in the Early Pliocene, in the Mid-Pliocene major uplift of central Kenya (about 1500 m) was followed by a second period of rift faulting during which the new faults within the Turkana depression were formed and the movement on the boundary faults on the west side of the rift valley was rejuvenated.

A major episode of faulting occurring in the Early Pleistocene affected the eastern margin of the rift and generated step faulting of the rift floor; this was followed by a period of minor faulting in Post-Mid Pleistocene (Table 2.3). Fig. 2.6 indicates the pattern of faulting in this branch of EARS.

Table 2.3: Structural Evolution of the Kenya Rift Valley

Late Oligocene- Early Miocene	Regional downwarp, minor faulting in the Turkana area
End of Miocene	Faulting of phonolite plateau, largest faults on the W, monoclinical flexuring on the E
Early Pliocene	Comparatively Stable
Mid-Pliocene	Movement on the boundary faults, creation of new faults in the Turkana depression, the deepening of the central graben, formation of the present configuration of the rift
Early Pleistocene	Major episode of faulting (third period), renewed movement on the pre-existing faults, step faulting of the rift floor deepening its central part, breakup of the floor into sub-parallel blocks
Mid-Late Pleistocene	Minor faulting, rejuvenation of the faults of central graben, further deepening of the rift valley

Sources: Baker and Wolhenberg, 1971; Baker et al., 1971, 1977; King and Chapman, 1972; Logatchev et al., 1977; Williams, 1978a, b; Crossley, 1979.

Fig. 2.6: Pattern of faulting in the eastern branch of EARS. After Fairhead (1976), slightly modified.

2.4.2 The Western Rift Valley

The western rift is associated topographically with a series of lakes. Lake Tanganyika occupies a graben and its southern part is characterized by parallel graben and horst structures extending into the rejuvenated Precambrian Rukwa trough. The Lake Malawi trough continues the extension of the rift faults to the south.

In the Lake Malawi trough there exist dinosaur beds which date back to Early Cretaceous. The only major rifting in this area is that of Mid-Pleistocene which deepened the floor of the pre-existing valley. This deepening reached a maximum of 700 m and formed the basin which is now occupied by Lake Malawi.

From the Lake Tanganyika graben as far north as Lake Kivu the major faults trend NW, N, and NE. Further north the rift faults, which range from Miocene to Late Pleistocene in age (Cahen, 1954), trend N and NE. The faults at the northern part of the western rift are mostly en echelon in character. Examples include the Lake Edward- Lake Kivu and the Lake Albert graben.

The oldest sediments of the Lake Albert basin date back to Early Miocene. Therefore the initial downwarp of the basin must have occurred at least as far back as this period. This was followed by sedimentation (up to 900 m in thickness) in Pliocene to Early-Mid Pleistocene. The major faulting occurred in Late Pleistocene, postdating the erosion surfaces of the rift, and was followed by minor faulting. The result of this faulting was the continuous sinking of the rift and the accumulation of the maximum of 2600 m of sediments in the Lake Albert area (evidenced by drilling). In Late Pleistocene minor faulting of the rift floor has continued up to the present (Bishop and Trendall, 1966) (Table 2.4).

Table 2.4: Structural Evolution of the Western Rift Valley

Late Cretaceous- Early Paleocene	Regional uplift
Early-Mid Miocene	Initial Downwarp
Late Miocene	Warping continued and minor faulting initiated
Pliocene	Step faulting
Early-Mid Pleistocene	Sedimentation (maximum 900 m in the Lake Albert valley)
Mid-Late Pleistocene	Major boundary fault movement
Late Pleistocene	Minor faulting
Sources: Hopwood and Lepersonne, 1953; Cohen, 1954; Bishop and Trendall, 1966; King, 1970; Burke and Whiteman, 1973	

2.5 VOLCANICITY

Although EARS displays the greatest range of magma types in composition and age, three types of volcanic rocks can, in general, be distinguished (Gerasimovsky and Polyakove, 1974): (1) the alkaline basaltic rock series which include picritic basalts, basalts, olivine basalts, andesitic basalts, trachy-basalts, trachytes and phonolites; (2) the Na-alkaline rocks such as olivine melilitites, olivine melanephelinites, alkalic picrites, nephelinites and phonolites; and (3) K-alkaline rocks that include katungites, mafurites, ugandites, leucitites and leucite basanites. In general the Na-alkaline volcanic rocks are restricted to the eastern rift whereas the K-alkaline rock series are typical of the western rift (Gerasimovskv, 1977).

2.5.1 The Eastern Rift

In the eastern branch of EARS, which is a high-volcanicity "wet" rift (see Chapter 1), the volcanic activity is not confined to the rift zone, but extends both to the west and east for a distance of more than 200 km (King and Chapman, 1972) (Fig. 2.7). The volcanic activity is, however, centred on the Kenya dome which structurally dominates the eastern rift. The Kenya (Gregory) rift, which is separated from the Ethiopian rift by Lakes Stephanie and Rudolf (Turkana), bisects the Kenya dome (Moore and Davidson, 1978) (See Fig. 2.1). In this rift, which is 750 km long, 25-60 km wide and 1-2 km deep, continuous volcanic activity has persisted from Oligocene-Miocene time up to the present. Thus, a study of the volcanic evolution of the Kenya rift can reveal much about the evolution of the eastern rift in general. Table 2.5 and Fig. 2.7 show the generalized volcanic sequence and its

Table 2.5: Generalized Volcanic Sequence and Corresponding Volume in the Kenya Rift Valley

Age	Main Volcanic Activity	Total Volume (km ³)
Recent & Pleistocene	Minor basaltic cones, cinderite and rhyolite plugs Phonolitic-trachytic calderas	23000
Pliocene	Trachytic Ignimbrites, flood trachytes in the rift floor Basalts and basanites Nephelinite-trachyte-pyroclastic volcanoes of the rift floor	56000
Miocene	Extensive fissure eruptions of phonolite over a wide area Eruption of alkali basalts and tuffs from many small centres Eruption of nephelinites, mel- lites, emplacement of alkaline igneous rocks and carbonatites in the Kavirondo Gulf area	65000
Sources: Williams, 1972; Bailey, 1974

Figure Eruptions or Multicentre Eruptions

	nephelinite phonolite	basalt basanite	basalt-phonolite melanephelinite	phonolite phonolitic trachyte	trachyte phonolitic trachyte	comendite paestellerite	hugawite trachytic rhyolite
Quaternary							
Pliocene							
Miocene							

Major Central Volcanoes

	nephelinite phonolite	basalt-trachyte phonolite	phonolite trachyte	basalt trachyte
Pliocene to Recent				
Miocene				

Fig. 2.7: Age and the distribution of the main volcanic associations of the eastern rift. After Williams (1970).

distribution in the Kenya rift valley. The timing of this volcanic activity along with the timing of the tectonic events reveal that in this sector of EARS the tectonic and volcanic events started earlier in the northern part and migrated southward (King and Chapman, 1972).

2.5.2. The Western Rift

The western branch of EARS, a low volcanicity "dry" rift (see Chapter 1) is characterized by rocks of the potassic series such as katungites, mafurites, ugandites, leucitites and basanites. Four volcanic zones are associated with this rift, namely, from north to south, Toro-Ankole, Birunga (including Bufumbira), South Kivu, and Rungwe volcanic fields (Fig. 2.8).

The Toro-Ankole field which is located to the south and southeast of Ruwenzori is characterized by unsaturated potassic volcanic rocks (Holmes, 1940, 1950). Volcanic activity in this zone started in the Middle Pleistocene (Holmes and Harwood, 1932).

The Birunga field is situated on the northern side of Lake Kivu and is characterized by lavas ranging from leucite basanites, nephelinites and latites to basanitoid trachybasalts. This field has been continuously active during the last 2 to 3 Ma. The majority of the samples analyzed in the second part of this thesis belongs to this field.

The South Kivu field is located in the southern part of Lake Kivu and is composed of two volcanic zones, Mwenga-Kamituga and Tshibinda, situated respectively to the SW of Lake Kivu and on its southwestern border. This field which is characterized by basic lavas like basanites, alkali basalts, olivine basalts (poor in K), trachytes

▨ Tertiary and Quaternary volcanic areas

— Major Faults

(after Guibert, 1978 ; Barberi et al, 1982)

Fig. 2.8: Volcanic fields of the western branch of EARS.
After Guibert (1978) and Barberi et al. (1982).

and quartz-tholeiites (Holmes, 1940) was active from at least Early to the Middle Pleistocene. An eruption was recorded in the last century (Guibert, 1978).

The Rungwe field lies at the junction of the Lake Rukwa and Lake Malawi rifts which join the eastern and the western branches of EARS. In this field, the volcanic activity started in Late Pliocene and stopped at the end of the eighteenth century (Harkin, 1960). The most abundant rock types are alkali basalts, nephelinites, phonolites and trachytes (Barberi et al., 1982). It is therefore evident that in the western branch of EARS volcanicity gets younger to the north (Barberi et al., 1982).

Since the volcanic rocks of the two branches of EARS may be used to infer the characteristics of the upper mantle beneath them, their variation in age, volume, and the type of the volcanic activity may reflect differences related to the upper mantle. Thus, it is instructive to compare the characteristics of the volcanics from the two branches.

In contrast to the eastern branch, the western branch of EARS is a low volcanicity rift, as its volume of volcanics is lower (see Chapter 1). This branch was initiated and developed within the last 5 Ma, i.e. much later (20-25 Ma) than the eastern rift. Its volcanic activity is of shorter duration and gets younger to the north. The reverse, however, holds for the eastern branch whose volcanic activity began in Oligocene in Ethiopia, Miocene in Kenya, and later in Tanzania (Shackleton, 1978).

In the alkalic rocks of the western rift K dominates over Na and leucite is the characteristic mineral. However, in the alkalic rocks of the eastern rift Na dominates over K and nepheline is the dominant

mineral. Furthermore, in the western rift the extreme undersaturated lavas and carbonatites of recent age occur in the north, whereas in the eastern rift they occur only in the south. These facts rule out interpretation of the two branches of EARS as simply cracks that propagated in the same direction.

To understand these differences in morphology, tectonic history and age, volume and chemistry of the volcanic rocks of the two branches of EARS, it is necessary to know the structure of the lithosphere and the nature of the upper mantle beneath the rift system. Gravity, seismicity and heat flow data together provide the best information on these subjects.

GRAVITY

3.1 INTRODUCTION

One of the earliest gravity surveys of east Africa dates back to 1936 (Bullard). The major finding of this survey was the existence of a broad negative Bouguer anomaly associated with the east African plateau, on which narrower anomalies associated with the rift system were superimposed. This observation was later confirmed by the surveys of Girdler et al. (1969), Vogt (1978) and Fairhead and Reeves (1977). Several other results have also been established on the basis of the features of gravity anomalies:

(a) There is an inverse correlation between the surface elevation and Bouguer anomalies in east Africa, indicating that this area is in approximate isostatic equilibrium (Bullard, 1936; Sowerbutts, 1969; Banks and Swain, 1978). Moreover, east-west Free Air anomaly profiles over the rift fluctuate around zero and have wavelengths similar to the changes in the topographic relief. This implies that the region of rifting is indeed in isostatic equilibrium (Darracott et al., 1972; Fairhead, 1976).

(b) The major anomalies have very long wavelengths (Fig. 3.1), extending for more than 1000 km over the region of rifting (encompassing both the eastern and western rifts) (Girdler et al., 1969). These anomalies, therefore, have deep-seated sources and are caused by mass distributions at depth rather than by near-surface features like the presence of low- or high-density volcanics. The fact

Fig. 3.1: The Bouguer anomaly map of Africa at contour intervals of 20 mgal. Redrawn from Slettene et al. (1973).

that the regional anomaly persists in the areas which are not covered by the Cenozoic volcanics (for example South Kenya) also indicates that the anomaly cannot be attributed to the effects of the surface volcanics. Besides, the lateral extent of these volcanics is very small (eastern rift, 70 km) compared to that of the negative Bouguer anomaly, and can only correlate with the short-wavelength anomalies which in turn are superimposed on a much broader negative Bouguer anomaly (Fairhead, 1979).

(c) Since the east African plateau is in approximate isostatic equilibrium, its Cenozoic uplift must somehow be compensated. As evidenced by erosion surfaces, the plateau has been uplifted and peneplaned several times since the Jurassic. Assuming that the plateau was in approximate isostatic equilibrium at the end of each erosion cycle, each succeeding uplift must have been caused by lighter material being added at depth. Therefore, it is suggested that the uplift might have been contemporaneous with either formation or migration of low density subcrustal materials in the uplifted areas (Bailey, 1964; Bott, 1965). Studies of erosion surfaces show that most of the African continent was uplifted a few hundred meters in the Cenozoic, during which most of east Africa was subjected to a much greater uplift (around 1300 m) (King, 1970). This implies that a greater amount of low density material was generated or migrated beneath east Africa. The occurrence of low density material beneath the rift system is evidenced by the presence of currently active volcanoes and the large amounts of volcanics in the eastern branch, and, to a lesser degree, in the western branch of EARS. Furthermore, a trend of small wavelength negative Bouguer anomalies following the volcanic centres, together

with density measurements, suggest that these small wavelength anomalies are caused by low density lava piles associated with these volcanic centres (Darracott et al., 1972).

3.2 DESCRIPTION OF THE ANOMALIES

The Bouguer anomaly map of Africa (Fig. 3.1) shows that the whole of EARS north of 5°S is associated with a negative Bouguer anomaly. This gravity feature, which is clearly related to the rifting (Brown and Girdler, 1980) is about 3500 km long and 1000 km wide with a maximum amplitude of less than -200 mgal*. This amplitude increases southward the morphotectonic expression of the rifting decreases (Darracott et al., 1972). Within the rift itself, superimposed elongated gravity lows are associated with the older sedimentary basins, and sometimes a positive anomaly is superimposed on the negative gravity anomaly.

3.2.1 The Eastern Rift

Although the negative Bouguer anomaly covers the whole region of EARS, the largest anomaly correlates with the eastern rift near the Equator (Girdler, 1975). Here the Bouguer anomaly reaches its maximum amplitude of -240 mgal and is associated with the area of maximum uplift of the Kenya dome. Along a typical profile at 2.5°S, the Bouguer anomaly decreases from +150 mgal along the shores of Lake Victoria in the west to -190 mgal over Lakes Eayasi and Magadi at 36°E and increases again to -50 mgal in the east.

* 1 milligal (mgal) = 10 gu (gravity unit) = 10^{-5} m/s^2

65

In general, the rift shoulders, the volcanic centres in the rift, the lakes, and the sedimentary basins are characterized by very small wavelength elongated gravity minima superimposed on a broader anomaly. The long wavelength Bouguer anomaly appears to die out south of 5°S which is coincident with the change in character of the rift from a complex graben structure to a region of simple block faulting (Darracott et al., 1972).

A much smaller but important feature of the gravity field is the presence of a superimposed linear positive anomaly, about 10 to 70 km wide and parallel to the axis of the rift from Lake Rudolf to Lake Magadi (Fairhead, 1976). This positive anomaly, running for at least 200 km and with a relative amplitude of 30-60 mgal (Girdler et al., 1969; Searle, 1970), was first discovered by McCall (1967) in the region 0.5°N and 0.5°S . It follows the main trend of the grid faulting, the geothermal and microseismic activity and the volcanic centres of the rift floor (Fig. 3.2), and it is interpreted to be due to a relatively dense intrusion zone in the lower crust along the axis of the rift.

3.2.2 The Western Rift

The western branch of EARS, like the eastern branch, is situated on the same large negative anomaly. Its eastern side is associated with an anomaly of -150 mgal which decreases to -200 mgal as it approaches the eastern rift. The western side correlates with an anomaly of -80 to -90 mgal which is in part due to the presence of light sediments in the Congo basin.

A large amplitude negative anomaly (minimum -200 mgal) is associated with the Lake Albert and Lake Edward rift valleys. Since

Fig. 3.2

The axis of the linear positive anomaly of the eastern rift and its correlation with geological features. Sources: McCall, 1967; Johnson, 1969; Searle, 1969, 1970; Tobin et al., 1969; and this study (see Fig. 3.10).

geological data and information from exploratory drillings suggest a maximum thickness of 2000-2500 m of sediments in the Albert rift, part of the negative anomaly of this region may reflect the presence of the sediments that fill the rift valley and have a lower density than the rocks of the floor and the flanks of the rift. The effect of the thickness of the sediments on the gravity anomaly is indicated by the Bouguer contours for both Lake Albert and Lake Edward rift valleys. However, in the Lake George and Lake Kivu rifts, only thin sheets of sediments must be occupying the troughs since their presence does not have an appreciable effect on the Bouguer contours.

3.3 RELATIONSHIP BETWEEN GRAVITY ANOMALIES AND TOPOGRAPHY

In order to determine the regional (smoothed) Bouguer gravity values, the data for the average $1^{\circ} \times 1^{\circ}$ Free Air anomaly and altitude* (Appendix 3.1) are used to calculate the average $1^{\circ} \times 1^{\circ}$ Bouguer anomaly (a detailed Bouguer anomaly map of Africa is not available). This procedure is preferred to the direct use of the Bouguer anomaly because a large data base is used to obtain average values for each $1^{\circ} \times 1^{\circ}$ grid, crustal irregularities are smoothed out in the averaging process, the average values are already keyed into map coordinates, it is more convenient to use the smoothed profile in the quantitative modelling, and the deviation from the observed values to the calculated values is small. The following formula is used:

$$BA = FAA - 2 \pi \rho G h$$

*Data supplied by Bureau Gravimetrique International of France.

where BA is the average Bouguer anomaly, FAA is the average Free Air anomaly, ρ is the density of upper crustal rocks which is assumed to be 2670 kg/m^3 , G is the universal gravitational constant, and h is the average height for each grid cell. Using this formula, the Bouguer gravity means are calculated (Appendix 3.1). The average values of the FAA, the Bouguer anomaly and altitude are then contoured using the Isoval computer program of Davaud (1975) (Figs. 3.3, 3.4, 3.5 respectively). The regional anomaly trends indicate that the Congo basin corresponds to a background Bouguer anomaly value of -65 to -100 mgal, whereas the background value for the coast of the Indian Ocean lies between -30 to -50 mgal.

Fig. 3.4 shows that the regional (smoothed) Bouguer anomaly is oval-shaped with a major axis which trends slightly east of north. The regional anomaly is asymmetrical: it increases very sharply to the east towards the Indian Ocean, as evidenced by the closeness of the contours, but rises only gradually to the west towards the Congo basin.

Figs. 3.6 and 3.7 show $1^\circ \times 1^\circ$ Free Air and Bouguer averages plotted versus $1^\circ \times 1^\circ$ mean altitudes for the area 20°E to 40°E and 5°N to 10°S . No clear correlation is observable in the FAA data, which, however, cluster around zero. A negative correlation can be observed for the topography and the Bouguer anomaly in Fig. 3.7. The histogram of the corresponding means (Fig. 3.8) indicates peaks at -4 to +4 mgal for the Free Air anomaly (average of all means -3 mgal), -80 to -120 mgal for the Bouguer anomaly (average of all means -104 mgal) and 400 to 1200 m for the elevation (average of all means 950 m) of this area. For the whole of African continent corresponding values are 7

Fig. 3.3: Contoured map of the Free Air anomaly at intervals of 10 mgal for the area 5°N to 10°S and 20° to 40°E.

Fig. 3.4: Contoured map of the Bouguer anomaly at intervals of 15 mgal.

Fig. 3.5: Contoured map of the topography at intervals of 200 m.

Fig. 3.6: Graph of 10×10 mean elevation versus 10×10 Free Air anomaly.

Fig. 3.7: Graph of 10×10^6 mean elevation versus 10×10^6 Bouguer anomaly.

Fig. 3.8: Histograms of (a) $1^{\circ}X1^{\circ}$ Free Air anomaly, (b) Bouguer anomaly and (c) elevation.

mgal for the Free Air, -48 mgal for the Bouguer anomaly and 400 m for the topography (Brown and Girdler, 1980).

The histogram of the Free Air anomaly and its corresponding peaks indicates that the values of this anomaly fluctuate around zero and that the peaks of the Free Air anomaly correspond to the peaks of elevation. These results provide further evidence for the validity of the assertion that this area is in approximate isostatic equilibrium. This is further confirmed by the negative correlation of Bouguer anomaly and topography.

3.4 GRAVITY PROFILES

The Bouguer anomaly map of EARS compiled by Shelkove (1973) at the scale of 1:5,000,000 is used to obtain gravity profiles. Five east-west profiles have been constructed at 0.5°N , the Equator, 1°S , 1.7°S and 2.3°S . These profiles extend from the Congo basin (20°E) to the coast of the Indian Ocean (40°E) including both the western and the eastern rifts (Fig. 3.9). On these profiles are also drawn a gravity profile which was smoothed by eye in order to exclude minor crustal effects (the dotted lines) and the regional anomaly calculated from $1^{\circ} \times 1^{\circ}$ Free Air values (dashed lines). In the following section the latter profile is used in the quantitative modelling, although because they are fairly close to each other, either smoothed profile could be used, and the resulting difference between them in the lithosphere/asthenosphere boundary would amount to a maximum of 2 or 3 km.

For each of the Bouguer anomaly profiles the corresponding topography profiles are also given in Fig. 3.9. Comparing the Bouguer anomaly and topography profiles, it is seen that for all the profiles,

17°S

Fig. 3.9: Profiles of the gravity anomaly and topography. Above: solid lines denote the Bouguer anomaly based on the Bouguer map of Shelkove (1973); dotted lines indicate smoothed Bouguer gravity; and dashed lines are the calculated Bouguer anomaly using the Free Air anomaly and altitude data (see section 3.3 for details). Below: topography.

despite a slight shift in some cases, the Bouguer anomaly is almost the mirror image of the topography. This is especially so if the general features of topography and the regional gravity anomaly are taken into account.

The Bouguer anomaly profiles clearly show that both branches of EARS are associated with gravity lows, with the minimum more pronounced for the eastern rift. They also show that the largest Bouguer gravity anomaly over the western rift is situated around 1.7°S (-180 mgal) whereas for the eastern rift it lies near the Equator (-260 mgal). The Bouguer anomaly, asymmetrical in nature, increases systematically towards the shoulders of the rift.

The profile at the Equator shows clearly a positive anomaly associated with the floor of the eastern rift. In order to determine the length and magnitude of such anomaly profiles in quarter-degree intervals have been taken over the eastern rift using the gravity map of Shelkove (Fig. 3.10). These profiles, which start at 0.5°N and end at 2°S , indicate that the positive anomaly starts at 0.25°N , increases in amplitude southward with a maximum at 0.25°S and dies out at 2°S . The positive axial anomaly has a width of 10 to 70 km, an amplitude of 20 to 50 mgal, and a length of more than 200 km with the maximum amplitude (50 mgal) occurring near the Equator.

The regional negative Bouguer anomaly over EARS could be due to lateral changes in the density of the crust, in the density of the upper mantle (i.e. variation in the thickness of the lithosphere), or in the depth of the crust/mantle boundary, individually or in some combination (Fairhead, 1976). As noted earlier, however, the long wavelength of the major anomalies shows that they cannot but have a deep source. This argues against the first of the above factors as a

2

Microlith

Fig. 3.10: Profiles for the amplitude and the extent of the positive Bouguer anomaly over the eastern rift.

possible explanation. The third is ruled out, at least as the major contributor, by seismic studies that have shown that the thickness of the crust in the region does not change sufficiently to explain the broad negative anomalies (Brown and Girdler, 1980; Shudofsky, 1985). There remains the second factor: the negative anomalies arise from changes (i.e. thinning) in the thickness of the lithosphere and the thermal processes in the underlying asthenosphere which result in anomalously low density (Fairhead, 1979; Brown and Girdler, 1980). This conclusion is supported by seismic studies in which the delay times reveal the existence of a substantial low velocity zone in the upper mantle indicating the presence of anomalous low density material beneath EARS. This interpretation is in agreement with that of Sowerbutts (1969), who attributes this negative Bouguer anomaly to a thin body of low density material at the base of the crust, and Girdler et al. (1969) who interpret it as a thick asthenospheric body of very small density contrast which has replaced the lower part of the lithosphere.

The narrow positive anomaly over the eastern rift has been interpreted by Zorin and Lepina (1985) to be due to a dense intrusion zone 10 to 25 km wide and reaching to within 10 km of the surface in some places. This interpretation has been supported by a seismic refraction line along the axis of the rift indicating a seismic velocity of 6.4 km s^{-1} at a depth of even less than 10 km (Griffiths et al., 1971). Therefore, this anomaly can be interpreted as being due to the presence of an intrusive body of a small positive density contrast (with respect to the crust) feeding the volcanic centres of the rift.

3.5 QUANTITATIVE INTERPRETATION OF BOUGUER ANOMALIES

The negative Bouguer anomaly is interpreted to be due to the penetration of low density asthenospheric material (as compared to the lithosphere) beneath the rift system. In the following, only the lithosphere/asthenosphere boundary is modelled, neglecting the penetration of the asthenospheric material into the crust, which produces the superimposed positive anomaly in the eastern rift.

In order to interpret the negative Bouguer anomalies quantitatively the choice of density contrast and the thickness of the lithosphere away from the region of rifting are important, since a small change in the contrast will change the position of the lithosphere/asthenosphere boundary. What is evident is that the density contrast is positive with respect to the sialic crust for the axial positive Bouguer anomaly associated with the eastern rift between 0.25°N and 2°S , whereas for the long wavelength negative anomaly this contrast is negative with respect to the lithospheric upper mantle.

From surface wave dispersion studies over east Africa the thickness of the lithosphere away from the region of rifting is estimated to range from 80 to 100 km (Fairhead, 1976). It is thus reasonable to assume a thickness of 90 ± 10 km for the seismically defined lithosphere away from the rift zone. Since the Free Air anomaly fluctuates around zero, the following relation is approximately valid.

$$\Delta BA = - 2 \pi G \Delta \rho \Delta h$$

where Δh is the difference in the thickness of the standard lithosphere from the lithosphere in the rift zone; $\Delta \rho$ is the contrast in density between the lithosphere and asthenosphere; and ΔBA is the

corresponding difference in Bouguer anomaly between the standard lithosphere and the lithosphere in the rift zone.

If it is assumed that the positive anomaly over the eastern rift is due to the penetration of asthenospheric material into the crust, then it can be concluded that the largest negative Bouguer anomaly indicates the position where the asthenosphere touches the base of the crust. This is because a positive density contrast produces the positive gravity anomaly, and all of the negative anomaly corresponds to the effect of asthenospheric upwelling. Therefore, before estimating the thickness of the lithosphere the thickness of the crust in EARS should be known.

Although seismic data on crustal thickness in east Africa are lacking, the model of Gamper and Pomeroy (1970) on the possible standard African lithosphere seems to be quite useful. In their model which is based on group and phase velocity dispersion data (Fig. 3.11), a thickness of 36.2 km is estimated for the crust. If the standard lithosphere is assumed to have a thickness of 90 km, the lithospheric thickness reduces to 36.2 km at the place where the negative anomaly is at its minimum (e.g. at 0° and 36°E). At 0° and 36°E , the negative Bouguer anomaly reaches a value of 260 ± 10 mgal. In this case $\Delta h = 90 - 36.2 = 53.8$ km and $\Delta BA = -260 - (-48) = -212$ mgal where -48 mgal is the most frequent value for the Bouguer anomaly for the whole African continent and is the best estimate away from the regions of rifting (See histograms of mean Bouguer and Free Air anomaly in Brown and Girdler, 1980). From these the density contrast is calculated to be $\Delta \rho = 100 \pm 5$ kg/m³.

For the quantitative modeling, the two-dimensional MAGRAV computing program of McGrath (1979) was used. The model assumes (a) a

Fig. 3.11

Model of Gumper and Pomeroy (1970)
for standard African lithosphere.

thickness of 90 km for the lithosphere, (b) the crustal structure as proposed by Gumper and Pomeroy (1970), (c) a density contrast of 100 kg/m³ at the lithosphere/ asthenosphere boundary and a constant $\Delta\rho$ throughout the rift zone. In Fig. 3:12 the smoothed Bouguer anomalies (calculated from 1°x1° Free Air data) are plotted along selected profiles across EARS. The lithosphere/ asthenosphere boundary is adjusted, by trial and error, to fit the smoothed gravity values. The models give Bouguer anomalies which fit the smoothed values very closely (within ± 5 mgal). For each model the shape and the depth to the lithosphere/ asthenosphere boundary are also drawn. It should be noted that the depth of the lithosphere/ asthenosphere boundary depends very much on the choice of density contrast. If the magnitude of $\Delta\rho$ is doubled, the Bouguer anomaly will also double and the inferred depth to the asthenosphere/lithosphere boundary will decrease by approximately 50 percent.

The overall shape of the lithosphere indicates that it is thinnest under the eastern rift in the area of the Equator, the area which has been affected by the largest amount of uplift. This agrees with the results of Fairhead (1976) and Zorin and Lepina (1985).

The positive anomaly in the eastern rift is related to the intrusion of the asthenospheric material into the crust. The evidence of the existence of a thinned crust under this branch of EARS also comes from seismic refraction studies along the axis of the rift (Maguire and Long, 1976). Since the positive anomaly is narrow (10 to 70 km wide) and has a low magnitude of 30 to 60 mgal, relative to the broad, negative anomaly of the whole rift, it is not quantitatively modelled here but the causative body has been inferred to be within 10

0.0

ER

WR

1°S

0°E

W R

E R

30

190

1.7 °S

0

36

90

28

30

32

34

36

38

0E

Fig. 3.12: Models of the structure of the lithosphere satisfying the smoothed Bouguer anomaly (see text for details). Top: gravity profiles. Solid lines, observed and dashed lines, computed values. Intermittent dashed and solid lines indicate that the curves are coincident. Bottom: profiles of the fitted lithosphere / asthenosphere boundary. Vertical exaggeration for the bottom profile is 8.85 : 1. Crust not to scale.

km of the surface near the Equator (Zorin and Lepina, 1985).

Table 3.1 indicates the depth which the asthenospheric material should reach in order to explain the negative gravity anomalies observed for each profile (disregarding crustal intrusions where occurring). This Table shows that, in general, the lithosphere is thinner under the eastern rift than under the western rift. The model for the profile at 1.7°S shows that this area is where the lithosphere is thinnest in the western rift. It should be noted that the first two profiles pass through the Toro-Ankole field and the profile at 1.7°S passes through the Birunga volcanic field.

The lithosphere is thinnest in the area of maximum uplift (in the eastern rift) as well as in the areas associated with the active volcanic fields (in the western rift). The larger amount of thinning most probably has resulted in a higher amount of uplift and volcanism in the eastern branch as compared to the western rift.

The gravity data attest to the existence of an asymmetric lithosphere under the rift zone in east Africa. The modelling indicates that this asymmetry varies from profile to profile, normal to the strike of the rift. Similar conclusions have been reached, using the gravity data, for most of the major continental rifts (e.g. Swain et al., 1981; Browne and Fairhead, 1983; Browne et al., 1985).

In conclusion, the broad negative Bouguer anomaly is caused by the presence of asthenospheric material at shallow depth, with a small negative density contrast at the base of the lithosphere. This asthenospheric body extends under the whole east African Plateau and is shallowest beneath the rifts. All the models agree on the existence of a very thin lithosphere underneath the two branches of EARS, especially in the eastern rift beneath the domal uplifted area of

Table 3.1

Depth to the base of the lithosphere in both branches of EARS based on the quantitative gravity modelling of the negative Bouguer anomaly.
(Note: effects above 36 km are neglected.)

Profile	Eastern Rift (km)	Western Rift (km)
0.5 °N	36	59
Equator	36	54
1.0 °S	38	48
1.7 °S	40	45
2.3 °S	41	52

Kenya. The variation in the amount of thinning of the lithosphere might explain the differences in the chemistry and in the volume of the volcanics in the two branches of EARS, since these differences are related to the general state of the lithospheric upper mantle beneath the rift zone.

CHAPTER 4

SEISMICITY

4.1 INTRODUCTION

Seiberg (1904) was the first to describe the seismic activity in the east African rift zone. His macroseismic studies were followed by a more detailed description by Krenkel (1923). The first epicentral maps based on teleseismic records were prepared by Gutenberg and Richter (1949; 1954). Subsequent seismicity studies by Fairhead and Girdler (1972) and Pointing et al. (1985) show that a narrow zone of shallow seismicity, associated with the Gulf of Aden, connects with the shallow seismicity of the rift system in east Africa.

In this chapter seismological data are compiled for an area of east Africa from 5°N to 10°S and 25°E to 40°E , covering both branches of EARS down to the Mbeya junction. The aim is to analyse the seismological similarities and contrasts between the eastern and western branches as well as the structure of the crust and upper mantle beneath them.

4.2 MACROSEISMIC DATA

Earthquake data for EARS have been recorded on temporary as well as permanent stations. In EARS itself, 6 major stations are in operation. Relevant information about these stations is given in Fig. 4.1 and Table 4.1.

Fig. 4.1: Major seismic stations in the East African Rift Zone.
 AAE=Addis Ababa, Ethiopia
 BTC=Butembo, Zaire
 BTR=Butare, Rwanda
 LWI=Lwiro, Zaire
 NAI=Nairobi, Kenya
 RUM=Rumangabo, Zaire

Table 4.1: Seismograph stations in EARS according to the United States Coasts and Geodetic Survey (USGS) list of March 1974.

Code	Station and Country	Coordinates	Elevation (m)
AAE	Addis Ababa, Ethiopia	9.03°N, 38.77°E	2442
BTC	Butembo, Zaire	0.13°N, 29.28°E	1700
BTR	Butare, Rwanda	2.61°S, 29.73°E	1648
LWI	Lwiro, Zaire	2.27°S, 28.80°E	1748
NAI	Nairobi, Kenya	1.27°S, 38.80°E	1692
RUM	Rumangabo, Zaire	1.34°S, 29.36°E	1590

The data for this chapter have been supplied from the data file of the Institut für Geophysik, University of Zurich. They include the time (day, month, year, hour, minute, second), location (latitude and longitude), depth and the magnitude (m_b , body wave magnitude) of all the earthquakes recorded by each station from July 9, 1912 to March 21, 1983 (Appendix 4.1).

4.3 SPATIAL DISTRIBUTION OF SEISMICITY: EPICENTRAL MAPS

All the earthquakes in the data file, as well as earthquakes reported in the literature but not included in the data file, are plotted in Fig. 4.2. The accuracy of epicentre locations provided by the data file is not specified. However, Bram (1972) showed that the accuracy of epicentres is within ± 15 km for the earthquakes in the southwest of EARS by comparing the epicentre locations determined from the 1968 Seismological Tables with the P-phase recorded at different stations. In Fig. 4.2 the tectonic lineaments of the region are added using the International Tectonic Map of Africa (UNESCO, 1968). Inspection of this figure shows that, in general, the western rift is seismically more active than the eastern rift and that the earthquakes are not confined to the rift but occur also in areas in which no surface expression of faulting is evident. The latter has already been noticed by Fairhead and Henderson (1977) and Shudofsky (1985).

4.3.1 Western Rift

The distribution of epicentres in this branch of EARS indicates that almost all the tectonic structures are seismically active. At the

Fig. 4.2: Epicentral map of the earthquakes in EARS.

northern end of the rift (i.e. north of Lake Albert) both faulting and seismicity die out along a series of en echelon faults which terminate against the Precambrian Aswa shear zone (Fairhead and Stuart, 1982).

Earthquakes also occur outside the rift zone. Examples include the area between Lake Kivu and the Congo River and a seismically active zone stretching northwest of Lake Kivu. The centres of activity are situated in the Semliki-Ruwenzori region (southwest of Lake Albert) and around the northern end of Lake Tanganyika. The focal depths of these earthquakes range down to 25 km except in the Ruwenzori area where the focal depths extend to 40 km (Rodrigues, 1970; Fairhead and Girdler, 1971).

The comparison of the epicentral map of the western rift with the map of the Precambrian basement (see Fig. 2.3, Chapter 2) shows that the seismicity of the western rift follows the western edge of the Tanganyika shield and is associated with the trends of the younger Precambrian basement rocks (i.e. Kibaride belt to the north and the Ubendian belt to the south) (Bath, 1975, and Fairhead and Stuart, 1982). The rift structures lie within and parallel to the trends of the Precambrian basement.

It is worth mentioning that in the Birunga volcanic field the active volcanoes of Niragongo and Nyamulagira appear to be almost aseismic, although a few earthquakes occur elsewhere. The earthquakes in this area are most likely to be volcanic instead of tectonic and the quiet zone in the vicinity of the active volcanoes is interpreted to be due to the existence of magma reservoirs beneath the volcanoes, which inhibits the accumulation of stresses within the area (Wohlenberg, 1969). The same is probably also true of the northern part of the eastern rift.

4.3.2 Eastern Rift

Seismically, the eastern rift can be divided into two zones, northern and southern, with the division being the east-west line of the northern Tanzania volcanic centres at 3°S . This line is coincident with a tectonic change of this branch and the termination of the Kenya rift valley. For the period from 1912 to 1983, the seismic activity of the eastern rift is concentrated south of 3°S . Although the northern section of the rift is reported to be almost aseismic, the large earthquake of 6 January 1928 with a body wave magnitude of 6.8 occurred in this area, with the epicentre at 5°N and 36.5°E (Richter, 1958; McCall, 1967).

From the epicentral map of the eastern rift several other features are evident: there are epicentres which are not connected to the rifting itself; in the Lake Victoria area there appears to exist a seismically active zone; and the Kavirondo rift (around 0° , 35°E) and the Speck Gulf region (around 2.25°S , 33.5°E) are seismically active (Bath, 1975). McConnell (1972) considers the faulting in Tanzania, and consequently the seismicity, to be controlled by the presence of the Tanzanian shield, which has to a great extent prevented the propagation of fractures through it.

4.4 MICROSEISMIC DATA

4.4.1 Western Rift

Microseismic studies of various districts of Uganda (Lake George, Ruwenzori, Lake Albert) by Rykounov et al. (1972) and Maasha

(1975) showed that a high rate of microseismic activity is associated with the Ruwenzori region and that this region is the most seismically active area in Uganda (Fig. 4.3). This is especially so in the northern part of the region near the contact with the Lake Albert rift. In this area an average of 23 events per day were recorded by Rýkounov et al. (1972). As mentioned in the previous section, this area is also very active macroseismically. The second region of high microseismic activity is around Lake George and Lake Albert.

It should be noted that the Lake Albert-Lake Edward (Lake Mobuto-Lake Amin) region is not associated with active volcanism and that in general the volcanic eruptions in the western rift occur at the structural nodes (i.e. intersections of the main rift with transverse features) as in the Ruwenzori region where the north-trending Kivu-Edward rift intersects the northeast-trending Lake Albert rift (Maasha, 1975). This is in contrast with the Kenya rift which has abundant volcanism associated with the rift floor. In the Lake Albert-Lake Edward region, the spatial trends of the microearthquakes are correlated with the major rift faults and volcanic zones.

4.4.2 Eastern Rift

Microseismic reconnaissance of the eastern rift has revealed that the floor of the Kenya rift is seismically very active (Tobin et al., 1969; Molnar and Aggarwal, 1971; Hamilton et al., 1973). These studies show that some parts of the rift are associated with an average of up to 47 small events per day and also that the grid faults along the axis of the rift are more active than the border faults (Girdler et al., 1969). The microseismic events in the eastern rift are found to

Fig. 4.3: Epicentral locations of micrpearthquakes in the northern Ruwenzori region for the period June 17 to August 29, 1973. The volcanic areas are shown by dotted lines. After Maasha (1975).

have shallow focal depths of 3 to 4 km.

Local seismic studies of this branch by Molnar and Aggarwal (1971) and Maguire and Long (1976) indicate an abundance of microseismic activity associated with the geothermal areas of the rift valley. These geothermal areas, consisting of steam jets and hot springs south and southeast of Lake Naivasha (0.75°S , 36.5°E) have an average of 26.2 events per day. Also the southern end of the eastern rift near Lake Manayara (3.5°S , 35.8°E) is very active. The number of events with magnitude between 0.5 to 3.5 are about 30 per day (Rykounov et al., 1972). These geothermal areas parallel the grid faults and the axis of the gravity high.

South of 5°S there is a decrease in geothermal activity and the microseismic events are few, but a large number of well recorded earthquakes occur. This change in seismicity is inferred to be due to the existence of a thicker lithosphere in the southern portion of the eastern rift. This conclusion is based on the interpretation of the regional gravity data (Fairhead, 1976; see also Chapter 3).

In the northern part of the Kenya rift, which is characterized by a high rate of microseismic (and a low rate of macroseismic) activity, the tensile stresses are adequately released by low magnitude shocks along the axial zone of the rift where the linear intrusion of hot igneous material at shallow depth reduces the strength of the crust. To the south of this area as well as in the western rift, in contrast, the tensional stresses are released by larger magnitude earthquakes occurring both within and outside the rift zone. These areas may not have well developed zones of lithospheric weaknesses like the northern part of the Kenya rift.

4.5 MAGNITUDE-FREQUENCY RELATIONS

The seismicity of an area can be expressed by the equation

$$\text{Log } N = a - bM \quad (\text{Gutenberg and Richter, 1954})$$

where a and b are parameters characteristic of the region and N is the expected number of earthquakes per year within the magnitude interval $M + dM$.

The seismicity of the two branches as well as of the whole east African rift is evaluated for the period from 1959 to 1982. For this period dM is taken to be 0.5 and the number N of earthquakes was applied to one year (i.e. $N = n/24$) for each magnitude interval. The magnitude interval starts from 4.5 in this study because for the catalogue used, the data for magnitudes less than 4.5 are not complete. This is shown in Fig. 4.4 where $\text{Log } N$ is plotted against the magnitude intervals for three arbitrary periods (1951-82, 1959-82, and 1965-82).

Table 4.2 shows the number of seismic events for the period 1959-1982 from 28 to 33°E, 5°N to 10°S (western rift); from 33 to 38°E and the same latitudes, i.e. an equal area (eastern rift); and for the whole area of the rift zone from 25° to 40°E. The fitted relationships, using the least squares method, are as follows:

$$\text{Whole rift} \quad \text{Log } N = 4.30 - 0.813 M$$

$$\text{western rift} \quad \text{Log } N = 3.99 - 0.764 M$$

$$\text{eastern rift} \quad \text{Log } N = 3.87 - 0.924 M$$

If m_b is taken to be more than or equal to 5 (to allow for the possibility that the catalogue in the interval $dM = 4.5-5.0$ is not complete) then the fitted relationships are expressed as:

Fig. 4.4: Plot of body-wave magnitude (m_b) versus $\text{Log } N$ for the earthquakes listed in Appendix 4.1.

Table 4.2: The number of seismic events in the rift zone. n is the number of earthquakes in 24 years and N is $n/24$.

Magnitude	4.5-4.9		5.0-5.4		5.5-5.9		6.0-6.4	
1959-82	n	N	n	N	n	N	n	N
eastern								
rift	15	0.63	9	0.38	1	0.04	1	0.04
(33-38°E)								
western								
rift	64	2.67	55	2.29	10	0.42	6	0.25
(28-33°E)								
The whole								
rift	86	3.58	66	2.75	11	0.46	7	0.29
(25-40°E)								

Table 4.3: The ranges in 'b' value for the two branches of EARS.

Location	'b'	Reference
Western Rift	0.65 to 0.94	Wohlenberg, 1969; Ndontoni, 1976
Eastern Rift	0.79 to 1.35	Wohlenberg, 1969; Ndontoni, 1976; Bath, 1975; Fairhead and Girdler, 1971

Whole rift	$\log N = 5.27 - 0.981 M$
western rift	$\log N = 5.11 - 0.963 M$
eastern rift	$\log N = 4.37 - 0.989 M$

The magnitude-frequency relationships drawn in Fig. 4.5 for $m_b > 4.5$ indicate that, for the period 1959-82, the seismicity of the western rift, for magnitudes $m_b > 4.5$ or $m_b > 5$, dominates over that of the eastern rift.

Table 4.3 indicates the ranges of the 'b' values recorded in the literature for the two branches. The 'b' values calculated above fall within these ranges, but the differences are not significant.

The differences in 'b', if they could be proven to be real, have implication for tectonic stress. Wyss (1973) has shown that low 'b' values indicate high stress in the source region. The 'b' value for the eastern rift (0.924) is higher than that of the western rift (0.764), but the statistical significance of the difference in 'b'-values cannot be assessed. Furthermore, the two values are about the same for $m_b > 5$, indicating that differences in b-values are catalogue-dependent and that therefore no geodynamic inferences can be drawn from them.

4.6 FOCAL DEPTHS

In general, focal depths of earthquakes are difficult to determine, especially if they occur at shallow depths. For well constrained focal depth determination it is necessary to have a local array of stations with at least one station at an epicentral distance less than the focal depth, such as in the Ruwenzori region where the uncertainty of 5 to 10 km is conventionally given to focal depths (Chen

Fig. 4.5: Magnitude- frequency relation for the two branches of EARS.

and Molnar, 1983).

Focal depths of earthquakes in east Africa are shown in the histograms of Fig. 4.6. It appears that there is a concentration of events at a depth of 33 km for both branches of the rift as well as the Precambrian shield. The high concentration of events at this depth is suspicious in that some events might have been arbitrarily assigned to this depth without proper depth determination. Therefore, although some events might have occurred at this depth, it is more likely that most of them occurred at shallower depths.

Shudofsky (1985) accurately determined (with the uncertainty of ± 4 km) the focal depths of 11 earthquakes in EARS (4 in the eastern rift region and 7 in the western rift region) (Table 4.4). He argued that the observed focal depths, as large as 25 to 30 km, are unusually deep for an extensional tectonic environment. However, this assertion is not entirely valid for the following reasons: First, in the eastern rift the deepest earthquake (event number 4) for which the well-constrained focal depth was determined by Shudofsky (1985) occurs outside the rift's major bounding faults. In this rift seismic studies indicate a shield-type crust and sub-Moho material within 25 km of the axis of the rift (Savage and Long, 1985) implying that outside this zone there is normal crustal structure and thickness such that relatively deep earthquakes are not unexpected. Second, the average heat flow value of the rift's shoulders is 54 ± 12 mW/m² (see Chapter 5), very similar to that of the Precambrian shield surrounding the rift indicating a much stronger lithosphere in these areas. The rheological profile based on the heat flow value of 54 mW/m² shows a crustal brittle layer down to the depth of around 30 km (see Chapter 6) immediately outside the rift, indicating that crustal earthquakes could

Fig. 4.6: Histograms of earthquake hypocenters in (a) the eastern rift; (b) the western rift; and (c) the Precambrian shield (see text for explanation).

Table 4.4: List of some of the earthquakes in the western and eastern rifts for which the fault plane solutions are determined.
Sources: Appendix 4.1 and Shudofsky (1985).

Solution number	Rift	Year	Month	Day	hr.	min.	Coordinates	depth (km)	Magnitude (mb)
1	ER	1928	01	06	19	31	0.1°N 36.5°E	--	6.8
4	ER	1964	05	07	05	45	4.0°S 34.9°E	28	6.2
5	ER	1977	12	15	23	21	4.8°S 34.9°E	12	5.2
6	ER	1972	02	13	10	03	4.5°S 34.1°E	6	5.0
7	ER	1967	10	14	23	29	3.3°S 38.2°E	10	5.1
10	WR	1966	03	20	08	55	0.8°N 29.9°E	05	5.3
11	WR	1966	05	17	07	03	0.7°N 29.9°E	07	5.5
12	WR	1966	03	20	01	30	0.8°N 29.9°E	29	6.0
13	WR	1966	03	21	01	30	0.8°N 29.6°E	07	5.8
14	WR	1972	04	25	00	03	1.1°N 30.1°E	10	4.9
15	WR	1966	10	05	08	34	0.0°N 29.9°E	23	5.3
16	WR	1977	07	06	08	48	6.3°S 29.6°E	14	5.2
17	WR	1975	03	26	03	40	5.3°S 30.1°E	16	5.0

be expected down to this depth.

From the above two reasons it is clear that in the eastern rift region lower crustal earthquakes are to be expected on the rift's shoulders north of 3°S even though this is the area where the lithosphere is most attenuated. South of 3°S faulting is diffuse, macroearthquakes are abundant and there is a decrease in geothermal activity, which are all attributed to the presence of an even cooler, thicker lithosphere in the region; yet this is where Shudofsky derives the data for his argument.

The focal depths of seven earthquakes in the western rift were determined by Shudofsky (1985). Since the rheological profile of this branch indicates a brittle layer down to the depth of 18±8 km (see Chapter 6), only two events (events number 12 and 15, Table 4.4) show unexpectedly deep focal depths (29 and 23 km respectively). Event number 12 is located about 30 km outside the rift's bounding fault in the northern Ruwenzori region and is suspected to have occurred in the Precambrian shield (Fig. 4.7). Event number 15 with a 23 km focal depth is located in the rift itself. Based on the evidence of along-strike hot and cold regions in the western rift (see Chapter 5), this earthquake might have occurred in a cold region where the brittle layer is thicker, although there is no heat flow data in this part of the rift to support this explanation. In any case the uncertainty in focal depths determined by different methods is +4 km (Shudofsky et al., 1987) which indicates a minimum depth of 19 km for this earthquake. This is very close to the depth of the brittle layer determined by the rheological profile of this branch. It is possible, moreover, that earthquakes may occur below the brittle-ductile transition along shear

Fig. 4.7

Fault plane solutions in the Ruwenzori region, western rift. Solutions 1 to 9 after Maasha (1975); 10, after Sykes (1967); 11 to 15 after Shudofsky (1985). Black areas denote compression.

zones where the strain rate is anomalously high.

It is concluded that the rift regions are characterized by shallow seismicity, although lower crustal earthquakes can occur outside of the rift valley itself where the lithosphere is thicker, and perhaps within the rift valley where there is a decrease in geothermal activity.

4.7 FOCAL MECHANISMS

Focal mechanism solutions collected from the literature are shown in Figs. 4.7 and 4.8. In general, for the whole east African rift zone the fault plane solutions indicate normal fault mechanisms with the extensional axis normal to the trend of the rift (Shudofsky, 1985). Therefore, these fault plane solutions point to the tensional nature of the rift system which results from horizontal tensional deviatoric stresses oriented normal to the overall strike of the rift. Table 4.4 lists some of the earthquakes in the western and the eastern rifts for which the fault plane solutions have been determined.

4.7.1 Western Rift

Solutions 8 and 9 show normal faulting. Solutions 10 and 11 (Fig. 4.7) show a northwest-southeast striking nodal plane, a feature consistent with the surface faulting observed in the area (Banghar and Sykes, 1969). These solutions indicate normal faulting and a least compressive stress oriented approximately ESE-WNW (Maasha and Molnar, 1972). Solution 12 indicates normal faulting on a NE-striking plane (Shudofsky, 1985). Solution 13 also indicates normal faulting. All other solutions from the Rwenzori area suggest the dominance of east-west least compressive (deviatoric tensional) stresses.

Fig. 4.8: Fault plane solutions in the eastern rift. Focal mechanism solution 1 and 2 after Hamilton et al. (1973); 3 after Maasha and Molnar (1972); 4 to 7 after Shudofsky (1985).

4.7.2 Eastern Rift

For this branch of EARS from north to south the fault plane solutions show the following: Solution 1 indicates a right lateral motion on a ENE-WSW fault plane; solution 2 shows an east-west tensional stress axis (Fairhead and Stuart, 1982). Solutions 3, 4, and 7 indicate normal faulting; solution 5 shows normal faulting with an ESE-trending tensional axis; solution 6 shows normal faulting mechanism along a NW-striking plane (Shudofsky, 1985).

Therefore in general there is an abundance of dip-slip, and a lack of strike-slip mechanisms in the rift zone (Shudofsky, 1985). The general picture emerging from the fault plane solutions is that a tensional horizontal stress axis trends E-W to ESE-WNW (perpendicular to the axis of maximum lithospheric thinning) and that the fault mechanisms mainly denote normal faulting.

Although strike-slip and compressional mechanisms predominate outside the rift system, the lack of strike-slip focal mechanisms associated with the rifting in EARS distinguishes it from the oceanic rift system. It may therefore be concluded that strike-slip displacements do not take place, or take place aseismically and/or are taken up by en echelon faults (Fairhead and Stuart, 1982).

4.8 TRAVEL-TIME ANOMALIES

Body and surface wave data from Africa away from the region of rifting have revealed high upper mantle seismic velocities typical of shield structure. For example, the mean P-wave travel-time anomaly is found to be -0.47 ± 0.21 s for the southern part of Africa, a small

negative delay typical of shield regions (Poupinet, 1979), and no S-wave velocities lower than 4.63 km s^{-1} have been found in the upper mantle (Gumper and Pomeroy, 1970). However, for the paths crossing the rift system the travel time delays are positive, i.e. the rays travelling close to or beneath the rift system are slowed down (Positive and negative delay times indicate the late or early arrival of the seismic signal.)

The delay times for several stations in EARS are listed in Table 4.5. It is evident that the rift system is associated with a significant anomaly compared to the stable African shield. In general, the delay times reveal the existence of a substantial low velocity zone in the upper mantle beneath EARS. This is particularly so for the eastern branch. It follows, therefore, that in the mantle below the western rift there exists less anomalously low velocity material compared to the eastern rift.

On the basis of teleseismic studies of 112 events, Savage and Long (1985) found that the pattern of delay times at the centre of the Kenya rift indicate significantly smaller delay times (more than 0.5 second difference) than at the edges. This pronounced reduction in delay time which seems to be coincident with the axial gravity high is interpreted to indicate that some anomalous mantle material has penetrated the crust to form an intrusion of relatively high velocity, high density material, as compared to normal crustal material, along the axis of the rift. From the plot of the delay time versus the distance from the rift axis, the width of this intrusion is inferred to be 30 km. This high velocity, high density crustal intrusion is also inferred from the gravity studies.

Table 4.5

Seismic delay times (in seconds) for three seismic stations in east Africa and their corresponding Bouguer anomaly values.

Source	AAE	LWI	NAI
Lilwall & Douglas (1970)	2.20	0.78	1.88
Long et al. (1972)	2.70	1.10	2.50
Fairhead & Reeves (1977)	1.61	-0.44	0.53
Poupinet (1979)	2.39	0.41	1.32
Average	2.23	0.46	1.56
Bouguer anomaly (mgals)	-230	-153	-186

Seismic refraction work in north Kenya and around Addis Ababa indicates the existence of low P_n velocities in the rift zone (P_n and S_n are body waves travelling in the uppermost mantle just below the Moho). Away from the rift, the velocities appear to be normal (Griffith et al., 1971; Berkhemer et al., 1974). This conforms to the data of Long et al. (1972) which show that in the region between the two branches of EARS, the P and S velocities are typical of the "standard" shield structure of the crust and topmost mantle. This shield-type crust overlying the normal sub-Moho material is reported to exist to within 50 km of the axis of the eastern rift (Long, 1975). This means that the anomalous material beneath the Gregory rift is of limited lateral extent in the uppermost mantle. This conclusion is in agreement with observation on the propagation of S_n waves between the two branches of EARS, as well as with the lithospheric profiles inferred from gravity data (see Chapter 3).

4.9 SURFACE WAVE DATA

In their investigation of phase and group velocities of seismic surface waves for the whole African continent Gumper and Pomeroy (1970) noticed that S_n waves are on the average strongly attenuated in EARS. More precisely, S_n -waves propagate across the southern part, i.e. south of 10°S , but do not cross the northern part of the rift zone above the Equator. In the area between the Equator and 10°S , S_n waves are strongly attenuated. This is consistent with the surface wave dispersion studies for the rift valley path AAE (station at Addis Ababa, Ethiopia) to NAI (Nairobi) (Fig. 4.1) by Long et al. (1972) and Knopoff and Schlue (1972), which indicate that an anomalously low

shear wave velocity upper mantle exists beneath the rift zone, and that this low-velocity zone is more pronounced in the northern sector.

In their studies of the S_n propagation paths over many regions of the world Molnar and Oliver (1969) observed that, although S_n propagates very inefficiently for paths crossing mid-ocean ridge systems, they propagate efficiently across the stable regions of the Earth characterized by low attenuation (high Q). Thus high attenuation of S_n waves indicates the existence of a low- Q zone beneath EARS. High and low Q zones correlate with the lithosphere and asthenosphere, respectively, and therefore the lack of effective S_n propagation indicates the presence of a high attenuation zone resulting from the replacement of the mantle part of the lithosphere with asthenospheric material.

If this interpretation is correct the data of Gumper and Pomeroy (1970) and Knapoff and Schlue (1972) indicate that under the northern part of the east-African rift zone, the mantle portion of the lithosphere is thinnest and/or lacking and it becomes thicker towards the south. Further evidence for this conclusion has already been independently drawn from the gravity data. As mentioned earlier, the broad negative anomaly dies out at about 5°S (Girdler and Sowerbutts, 1970) which is about where the travel time anomaly also dies out (Fairhead and Girdler, 1972).

Table 4.5 also shows the Bouguer anomaly for the stations AAE, NAI, and LWI. Plotting the Bouguer anomaly against the mean delay times (Fig. 4.9) indicates a clear correlation. This observation implies that positive delay times and the negative Bouguer anomalies are both manifestation of the same underlying structure, and that the most probable cause for it is the regional changes in lithospheric

Fig. 4.9: Plot of delay time vs Bouguer anomaly in EARS.

thickness.

The surface wave studies, therefore, indicate that the depth of the lithosphere-asthenosphere boundary varies considerably both along EARS and from EARS to the surrounding shield areas, i.e. the boundary deepens and the lithospheric thickness increases away from the rift zone so that beneath the shield regions the asthenosphere is very thin or even absent.

4.10 LITHOSPHERIC STRUCTURE IN THE RIFT ZONE

Some further studies using integrated seismological techniques, and general results on the lithospheric structure, are summed up in this section. Savage and Long (1985) used local, regional, and teleseismic events to measure the delay time and to map the anomalous upper mantle zone. A composite cross section of the Kenya dome based on the stations along and across the rift is shown in Fig. 4.10. This figure shows a rapidly varying lithospheric structure across the Gregory rift with a normal lithospheric structure immediately to the west of this rift.

The data from the only refraction line along the Gregory rift suggest an uppermost mantle layer associated with anomalously low sub-Moho velocities (about 7.5 km s^{-1}) (Fairhead and Reeves, 1977). Further support for these observations comes from the variation of P_n velocities. P_n velocity in both branches of the rift ranges from 7.1 to 7.6 km s^{-1} , indicating the presence of a low velocity upper mantle in both rifts, while away from the rift this velocity ranges from 7.9 to 8.3 km s^{-1} (Maguire and Long, 1976). The former range is typical of the asthenosphere (LVZ) and the latter of the mantle part of the

Fig. 4.10: Right: A composite true-scale cross-section of the Kenya dome based on delay times at the stations along and across the rift. The shaded area denotes the low-velocity zone (Savage and Long, 1985), the dashed-dotted line is from a model by Long and Baokhouse (1976). Left: the station numbers and their locations along and across the rift. The shaded areas denote the volcanic fields.

lithosphere.

Seismic refraction experiments by Griffiths et al. (1971) in the northern part of the Kenya rift (Lake Hannington- Lake Rudolf- Lake Turkana areas) provided evidence that in this part of the rift, the asthenosphere is cutting and penetrating into the lower crust. From these data a high (relative to the normal continental areas) P-wave velocity of $6.3-6.4 \text{ km s}^{-1}$ is found to correspond with the layers just beneath the volcanics. Griffiths et al. (1971) interpreted this high velocity as being due to the axial intrusion reaching to within 10 km of the surface.

The existence of a thinned crust under the eastern branch of EARS and a normal crust away from it is indicative of a sloping Moho with a compressional velocity varying from 7.9 km s^{-1} on the flanks to 7.5 km s^{-1} along the axis of the rift (Griffiths et al., 1971). For much of east Africa the seismological investigations indicate that the crustal thickness away from the rift zone is approximately constant at $40 \pm 5 \text{ km}$ (Maguire and Long, 1976).

It has been argued that if the isostatic compensation in east Africa were taken up at the Moho, a larger variation of the crustal thickness would occur. Therefore, a substantial part of the isostatic compensation is presumably being taken up in low density and low P-wave velocity structure at greater depth (Fairhead and Reeves, 1977), i.e. in the anomalous mantle.

It has so far been established that the low density, low velocity asthenosphere thickens and partly replaces the continental lithosphere in the area of the rift zone. This interpretation is compatible with the interpretations of the gravity anomalies (see

Chapter 3) and the travel time residuals. As mentioned earlier, the injection of mantle material reaches as close as 10 km of the surface in the eastern rift at the Equator.

It should be noted that in EARS the thinnest lithosphere occurs in the most uplifted areas, characterized by the highest volume of volcanics (e.g. Kenya dome). This zone of thin lithosphere, therefore, represents an area underlain by hot, low P-wave velocity, and low density material.

From the inversion of teleseismic data for the northern part of EARS, Nolet and Mueller (1982) derive models for the whole structure of the lithosphere underneath the rift zone. Their models, which are consistent with the previous interpretations (see Mueller and Bonjer, 1973), are presented in Table 4.6 and Fig. 4.11. The northern part of the western branch of EARS (extending northward from 1°S) is characterized by a 35 km crust and a thin high velocity lid overlying a channel possessing both low-S and low-P wave velocities (4.47 and 7.69 km s⁻¹ respectively). A reflector is detected at 140 km depth which marks the boundary to a high velocity material (4.67 and 8.44 km s⁻¹ for S and P wave respectively). This reflector, which was first detected by Bram (1975) is very sharp and is characterized by a significant jump in wave velocities.

The reflector is interpreted to mark the bottom of the low velocity channel. This discontinuity must be due to chemical or mineralogical changes, because thermal effects could not cause it to be so sharp, and a mechanical discontinuity is unlikely because temperature-induced changes in elastic parameters occur more gradually (Goetze, 1977). Furthermore, available phase diagrams show that sharp phase boundaries are rather unlikely considering the wide range of

Table 4.6: Lithospheric model of the eastern (a) and western (b) branch of EARS (after Nolet and Mueller (1982). ρ is the density and α and β are P and S-wave velocities.

Model EB (Eastern Branch)

Depth (km)	β km s ⁻¹	ρ g cm ⁻³
0- 20	3.521	2.760
20- 40	4.092	3.010
40- 78	4.433	3.248
78-101	4.095	3.377
101-161	4.215	3.377
161-220	4.412	3.377
220-300	4.591	3.487
300-400	4.822	3.592

(a)

Model WB (Western Branch)

Depth (km)	α km s ⁻¹	β km s ⁻¹	ρ g cm ⁻³
0 - 1.4	4.000	2.400	2.200
1.4- 14	5.815	3.607	2.789
14- 35	6.457	3.855	2.981
35- 55	7.973	4.700	3.288
55- 75	7.769	4.517	3.281
75-140	7.686	4.470	3.281
140-220	8.441	4.666	3.343
220-300	8.490	4.750	3.490
300-400	8.810	4.890	3.600

(b)

Fig. 4.11: Models resulting from simultaneous inversion of teleseismic data. After Nolet and Mueller (1982).

temperature at this depth.

In the northern part of the eastern rift the upper mantle, below 40 km depth, is characterized by low S velocities (4.43 km s^{-1}). This lid which extends to a depth of 78 km, is underlain by a channel possessing S velocities of 4.09 to 4.21 km s^{-1} extending to a depth of 161 km (Table 4.6).

The seismological data of Nolet and Mueller (1982) and their models for the eastern and the western rifts provide evidence that despite the assertions to the effect that the northern part of the western branch resembles the eastern one (Maasha and Molnar, 1972 and Green, 1976), there exist differences between the two branches which also extend to the northern parts.

4.11 DISCUSSION

The primary source for the tensional stress within the lithosphere in east Africa is most likely due to the regional uplift of the lithosphere which is the isostatic response to the rising of hot low density asthenosphere beneath the rift system.

Geological mapping, gravity, seismic-wave propagation, seismicity, and fault plane solutions clearly demonstrate the tensional nature of the rift system. The rift area is underlain by anomalously low velocity material which is assumed to represent a zone of partial melt at upper mantle depths (Banks and Swain, 1978). This zone has been shown to thicken towards the centre of the Kenya dome (Savage and Long, 1985).

The preferred interpretation of the P-wave delay times and their correlations with elevation and Bouguer anomaly is in terms of

variation of lithospheric thickness. The study of the propagation of S and P-waves, therefore, provides independent evidence for the existence of a thinned lithosphere underlain by anomalously hot, low density material beneath the rift system, as was inferred from gravity (Chapter 3). The gravity and seismic data imply the existence of an unmodified shield-type structure in the area between the two branches.

The seismic data indicate that EARS, like other continental rifts, is asymmetrical in cross section, normal to its strike. This asymmetry is further evidenced by the topography and the gravity data.

In this respect, it is interesting to note that Gibbs (1983) and Bosworth (1985) have suggested that low-angle normal faults play an important role in the overall geometry of rifts. The major boundary faults of the rift, which were formed when rifting initiated, change into low-angle faults at depth (Bally, 1982). If these low-angle "listric" faults cross cut at depth, then one of the detachments will lock, while the opposing detachment will remain active. As this fault propagates laterally, it has a tendency to curve inward toward the axis of the rift and die out (Bosworth, 1987). Beyond the point at which one main fault becomes inactive, the opposing locked main boundary fault may become active, in which case a reversal of asymmetry occurs (Fig. 4.12). This asymmetry alternates every 50-150 km along the axis of the rift (Mostafa, 1976); it can be documented by the change in direction of curvature of the rift in plan view, since the asymmetry is the surface manifestation of the polarity changes of the underlying low-angle fault systems (Bosworth, 1985).

Listric faults probably root at the brittle-ductile transition, since below this zone all displacements of the brittle layer can be

Fig. 4.12

Top: Schematic cross-section illustrating asymmetrical half graben structure above one dominant boundary fault. Bottom: diagrams showing the reversal of the main boundary fault along the rift axis at "accommodation zone" (AZ) due to curving inwards of the fault trace. T are transverse faults. After Bosworth (1985).

accommodated by ductile flow. From rheological profiles of the lithosphere (Chapter 6), it is apparent that the average depth of brittle- ductile transition is about 12 ± 5 km in the eastern rift and 18 ± 8 km in the western rift. Such a shallow brittle-ductile transition indicates that the lower crust in the rift zone is likely to be a zone of decoupling.

CHAPTER 5

HEAT FLOW AND REGIONAL GEOTHERMS

5.1 INTRODUCTION

The regional doming, rifting and magma genesis in the rift zones involve a variety of thermal processes. The study of heat flow and electrical conductivity can provide important clues concerning these thermal phenomena as well as constraints on temperature profiles under the rift zone.

It is well known that in contrast to the Precambrian shields which are associated with the lowest and least scattered values for heat flow averaging about 40 mW/m^2 * (Chapman and Pollack, 1974) the volcanic belts and the continental rift zones correlate with the highest and most scattered heat flow values on the continents (Bott, 1982a). In this chapter the heat flow data for the east African rift zone are studied and compared with the average heat flow of the surrounding Precambrian shield. The Precambrian shield average heat flow value will hereafter be referred to as the "normal" heat flow value.

The heat flow data used in this chapter were compiled from a data file provided by the University of Zurich. The data include the location of the sites (latitude and longitude), temperature gradient (mK/m), average conductivity (W/m K), and heat flow values (mW/m^2). However, the data file does not provide error margins for individual measurements.

* $1 \text{ mW/m}^2 = 41.84 \text{ HFU}$; $1 \text{ Heat Flow Unit (HFU)} = 1 \text{ } \mu\text{cal/cm}^2 \cdot \text{sec}$.

5.2 THE WESTERN RIFT

Heat flow data have been obtained from determinations in three deep lakes in the western branch of EARS, namely from north to south, Lakes Kivu, Tanganyika, and Malawi (Von Herzen and Vaquier, 1967). In these lakes the gradient was measured by a temperature probe which penetrated into the lake sediments and conductivities were either measured in situ, or on sediment cores brought up to the surface. The basic techniques for these measurements have been described by Langseth (1965).

5.2.1 Lake Kivu

Five heat flow values are available from Lake Kivu, which is surrounded by active volcanoes and geothermal springs (Table 5.1, Fig. 5.1). These values range from 17 to 185 mW/m^2 , indicating quite a wide scatter which has been confirmed by repeated measurements (Degens et al., 1973). These values, averaging about $73 \pm 70 \text{ mW/m}^2$ at one standard deviation level, indicate that heat flow is highly variable, a feature consistent with the recent volcanic activity in this region. It should be noted that the band of high heat flow has a width of approximately 50 km.

In this lake, characterized by a maximum water depth of 500 m, local effects such as high sedimentation rate have probably affected many of the measurements. Degens et al. (1973) have speculated that, after correcting for these effects the true heat flow would be two to three times higher than recorded. Also some heat is carried by hot springs, which indicates that 73 mW/m^2 is likely to be a lower bound (corrections to the observed heat flow values are discussed in section 5.4).

Table 5.1

Heat flow measurements in Lake Kivu

Latitude (°S)	Longitude (°E)	Heat Flow (mW/m ²)
1.77	29.02	99
1.82	29.18	25
1.95	28.99	17
1.99	29.02	41
2.13	29.15	185

Table 5.2

Heat flow measurements in Lake Tanganyika

Lat. (°S)	Long. (°E)	HF (mW/m ²)	Temp. Gradient mK/m	Conductivity W/m K
5.30	29.36	67.0	57	0.69
5.34	29.26	35.0	35	0.72
5.54	29.47	41.3	58	0.72
5.74	29.43	46.3	66	-----
5.77	29.54	39.3	55	0.71
5.92	29.54	57.0	72	0.80
5.97	29.45	151.3	190	-----
6.39	29.59	22.0	33	0.69
6.45	29.57	17.3	25	0.72
6.84	29.97	54.7	77	0.72
7.22	30.24	39.3	55	0.72
7.22	30.27	35.0	49	0.72

Fig. 5.1: Heat flow measurements in Lakes Kivu and Tanganyika. Values are in mW/m^2 .

5.2.2 Lake Tanganyika

In Lake Tanganyika twelve heat flow values are available (Table 5.2, Fig. 5.1). These values indicate that all except one of the measurements are less than 70 mW/m^2 . Disregarding this odd value the average of the eleven points indicates a heat flow value of about $41 \pm 15 \text{ mW/m}^2$. These eleven values indicate that a low to normal heat flow is associated with this lake. The odd value (151 mW/m^2 , at 5.97°S and 29.45°E) occurs near the locus where the linear trend of the lake is shifted by a transverse lineament. The average heat flow for the whole lake is $51 \pm 35 \text{ mW/m}^2$.

5.2.3 Lake Malawi

The measurements of twenty values of heat flow for this lake range from 6 to 120 mW/m^2 (Table 5.3, Fig. 5.2). Based on these data, the lake can be divided into three sections: a northern section down to 11.5°S ; a middle section from 11.5 to 12.5°S ; and a southern section from 12.5°S southward. The northern and southern sections of the lake are associated with low heat flow values, more so for the northern section, separated by a central region of high heat flow. Table 5.4 indicates the average heat flow values of these three sections. The average of all the data points gives a value of about $50 \pm 41 \text{ mW/m}^2$ which is similar to the average value for the South African Shield of 46 mW/m^2 (Lee and Uyeda, 1965).

Von Herzen and Vacquier (1967) noted that the high values in the middle section of the lake occur where the trend of the lake deviates from its linear form (Fig. 5.2), a feature already seen for Lake Tanganyika. This shifting or interruption is perhaps caused by

Table 5.3: Heat flow measurements in Lake Malawi

Lat. (°S)	Long. (°E)	HF (mW/m ²)

: Northern section :		
: 9.92	34.38	8.0 :
: 10.22	34.43	24.3 :
: 10.25	34.42	17.2 :
: 10.50	34.35	17.6 :
: 10.85	34.48	39.3 :
: 11.07	34.45	46.4 :
: 11.07	34.53	28.5 :
: 11.08	34.37	14.2 :
: 11.10	34.32	6.3 :
: 11.10	34.43	12.1 :
: 11.28	34.32	12.6 :
: 11.50	34.35	38.5 :
: Middle section :		
: 11.75	34.50	101.3 :
: 11.90	34.43	67.8 :
: 12.25	34.55	120.5 :
: 12.30	34.38	73.2 :
: 12.33	34.57	119.2 :
: Southern section :		
: 12.58	34.60	27.6 :
: 12.87	34.65	32.6 :
: 13.10	34.72	32.2 :

Table 5.4: Average heat flow values for the three sections of Lake Malawi

Area	# of measurements	Average HF

: Northern	12	22 ± 13 :
: Middle	5	96 ± 25 :
: Southern	3	31 ± 3 :

Fig. 5.2: Heat flow values in Lake Malawi. Circles with plus sign inside represent hot springs along the shore. After Von Herzen & Vaquier (1967).

or by magmatic intrusions to relatively shallow depths. Although the central section characterized by high heat flow is devoid of surface volcanism, the presence of magmatic intrusion is supported by the presence of hot springs along the lake shore (Fig. 5.2). The amount of the heat transported by these hot springs, however, is not known but indicates that the observed heat flow is a minimum.

The northern section of the lake shows very low heat flow values. Von Herzen and Vacquier (1967) suggest that these values are due to the high sedimentation rate. The anomalous heat flow recorded in almost all sedimentary basins is caused by factors such as sedimentation rate, water migration, etc. (Stegena, 1982) and does not reflect the true regional heat flow. In EARS, especially in Lakes Tanganyika and Malawi, bottom water warming since the last glacial period is another factor that reduces the measured heat flow. This phenomenon occurs in lakes with no bottom water circulation.

5.3 THE EASTERN RIFT

Thirty values of heat flow from different sites are available for the Kenya rift valley (Table 5.5). In general, on land the heat flow determinations are usually made by measuring the temperature gradient in a borehole deeper than 100 m and making laboratory thermal conductivity measurements on the rocks penetrated by the borehole. The basic techniques for the determination of heat flow on land are described by Beck (1965). Along the rift the heat flow values have a range of 6 to 194 mW/m² with most of them being above normal. The data are from the borehole based measurements and indicate significant convective ground water disturbances as in the western rift (Morgan,

Table 5.5: Heat flow measurements in the Kenya rift

Lat. (°N)	Long. (°E)	HF (mW/m ²)	Temperature Gradient mK/m	Conductivity W/m K
(A)				
: 3.35	38.32	78	17	4.70
: 3.18	38.15	87	17	5.25
: 2.77	38.12	59	34	1.77
: 1.97	37.93	53	30	1.76
: 1.92	37.90	45	--	--
: 1.87	37.88	37	20	1.90
: 1.03	36.15	111	66	1.69
: 1.00	35.97	43	27	1.56
: 0.80	36.27	101	66	1.31
: 0.75	36.27	87	66	1.31
: 0.70	35.22	64	48	1.34
: 0.68	35.48	47	37	1.25
: 0.65	35.22	68	48	1.34
: 0.57	35.32	76	51	1.49
: 0.53	35.30	42, 59	28	1.48
: 0.20	36.82	33	22	1.46
: 0.20	36.83	40	26	1.46
: 0.20	36.85	43	29	1.46
: 0.12	36.18	81	55	1.48
(B)				
Lat. (°S)	Long. (°E)	HF (mW/m ²)	Temperature Gradient mK/m	Conductivity: W/m K
: 0.07	36.07	54	47	1.17
: 0.09	36.00	55	47	1.17
: 0.10	36.90	22	15	1.48
: 0.13	36.12	103	82	1.26
: 0.17	36.13	194	150	1.29
: 0.23	35.82	56	65	1.04
: 0.23	35.90	67	65	1.04
: 0.28	36.00	176	199	0.96
: 0.35	36.00	190	199	0.96
: 0.35	36.13	137	150	1.29
: 1.30	36.78	6, 22, 39	20	1.10

1973). This is a factor that could significantly reduce the observed heat flow (note the presence of hot springs mentioned previously).

A typical example of the effect of thermal convection of ground water can be seen in the oceanic lithosphere near mid-ocean ridges. Anomalously low heat flow values (with respect to expected values on the basis of plate cooling models) and their large scatter are characteristic of mid-ocean ridges. This is interpreted to be evidence for convective water flow in the porous, fractured crustal rocks of the oceanic upper crust (Parsons and Sclater, 1977). Away from the ridge crest where the ground water convection is no longer effective, the heat flow values are higher. Therefore, it is speculated that the highly scattered heat flow values of the Kenya rift zone may also be partly explained by convective water flow.

Although the heat flow data do not show a trend along the rift, they reveal a definite pattern across it. Comparing the heat flow values of the rift floor with the ones corresponding to the rift shoulders, the values for the floor of the rift are seen to be higher (Fig. 5.3). The mean of the heat flow values increases from 61 ± 12 mW/m² (n=5) on the western flank to 111 ± 53 mW/m² (n=12) on the rift floor and falls subsequently to 49 ± 20 mW/m² (n=13) in the east of the rift. Table 5.6 shows the mean values for the floor and flanks of the rift.

The mean heat flow for the rift shoulders, which is about 54 ± 12 mW/m², is similar to the values determined for the coastal region of east Africa (Morgan, 1982). The values for this region which are determined from 55 measurements have a mean of 58 ± 16 mW/m² (Evans and Tammemagi, 1974; Evans, 1975). The mean heat flow value of the rift floor (111 ± 53 mW/m²) is more likely to be representative of the eastern

Fig. 5.3: Heat flow measurements in the Kenya rift. Values are in mW/m².

Table 5.6: Average heat flow values for the floor and flanks of the Kenya rift.

Area	# of measurements	Average HF
: Western flank	5	61 ± 12 :
: Rift floor	12	111 ± 53 :
: Eastern flank	13	49 ± 20 :

Table 5.7: The alternating high and low heat flow regions in the western rift from north to south.

Area	Average HF mW/m^2
: (A) :	
: Lake Kivu	99 ± 94 :
: :	
: Lake Tanganyika	73 ± 38 :
: :	
: northern	72 ± 24 :
: Lake Malawi central	178 ± 40 :
: southern	47 ± 3 :
: :	
: cold regions	64 ± 15 :
: hot regions	139 ± 51 :

rift. It should be noted that the band of high heat flow has a width of only approximately 70 km.

The high heat flow values associated with the eastern rift are substantiated by the presence of geothermal and Late Tertiary and Quaternary volcanic activity within and along the rift axis (Searle, 1970; Baker et al., 1972). In general, the study of the heat flow data from east Africa indicates that EARS, like other major continental rift systems of the world, is characterized by high values of heat flow (Morgan, 1982; see Chapter 1 and Table 1.2 for comparison).

5.4 HEAT FLOW CORRECTIONS

The measured heat flow values in lakes along the east African rift are subject to three main corrections: (i) bottom water warming, (ii) sedimentary blanketing, and (iii) hydrothermal circulation.

The first factor which may cause anomalous heat flow is the layer of deep cold water (if present) which does not mix with shallower water and has been gradually warming up since the last glacial period. In the absence of temperature sources or sinks, this would cause a decrease in the geothermal gradient through time (Von Herzen and Vacquier, 1967). This type of correction should be applied to Lake Malawi and Tanganyika because these two deep lakes lack bottom water circulation, as evidenced by their anaerobic conditions.

The effect of bottom water warming on the temperature distribution in the substratum can be approximated by the solutions given by Carslaw and Jaeger (1959, p. 63) for a semi-infinite solid whose surface temperature changes with time. If the surface temperature is initially T_0 , the temperature distribution within the solid is

expressed by equation (1). Equation (2) gives the temperature distribution at time $(t-t')$, assuming an instantaneous change in surface temperature (from T_0 to T_1) at time t' :

$$T = T_0 \left(1 - \operatorname{erf} \frac{z}{2\sqrt{Kt}} \right) = T_0 \operatorname{erfc} \frac{z}{2\sqrt{Kt}} \quad (1)$$

$$T = T_0 \operatorname{erfc} \frac{z}{2\sqrt{Kt}} + (T_1 - T_0) \operatorname{erfc} \frac{z}{2\sqrt{K(t-t')}} \quad (2)$$

where T is temperature, z is depth, and K is the thermal diffusivity.

Differentiating with respect to z gives the temperature gradient before (for equation 1) and after (for equation 2) the surface temperature change. Their difference gives the correction that should be added to the measured temperature gradient:

$$\left. \frac{dT}{dz} \right|_{z=0} = \frac{T_1 - T_0}{\sqrt{\pi K (t - t')}}$$

For a 10 degree temperature change occurring 10^4 years before present (this time being indicative of the climatic change following deglaciation), and a diffusivity of between 2×10^{-7} and $4 \times 10^{-7} \text{ m}^2 \text{ s}^{-1}$ (from measurements in the lake bottom sediments by Von Herzen and Vacquier, 1967), the correction to the temperature gradient is between 23 and 16 $^{\circ}\text{K/km}$. The correction decreases if the temperature change is smaller (for instance, it is approximately 11.5 $^{\circ}\text{K/km}$ for a temperature change of 5 $^{\circ}\text{K}$ and a diffusivity of $2 \times 10^{-7} \text{ m}^2 \text{ s}^{-1}$). The

values of 23 and 11.5 °K/km are taken here as the maximum and minimum corrections. Assuming that the same factor operates in both deep lakes, and taking the maximum corrections, the revised heat flow averages are 68 ± 35 mW/m² for Lake Tanganyika and 39 ± 13 , 113 ± 25 and 47 ± 3 mW/m² for the northern, middle and southern part of Lake Malawi, respectively.

The effect of sedimentary blanketing on heat flow is expressed by several but somewhat different relations (see for example Carslaw and Jaeger, 1959; Von Herzen and Uyeda, 1963; Birch et al., 1968). However, when applied, all these relations give similar results for the correction factor near the surface of the accumulating layer. The following formula by Von Herzen et al. (1974) expresses the effect of sedimentation on the measured heat flow values on the assumption of a constant sedimentation rate and zero rate of heat generation within the sediments:

$$\frac{1}{b} \left(\frac{dT}{dz} \right) \Big|_{z=0} = (1 + 2\tau^2) \operatorname{erfc}(\tau) - \frac{2}{\sqrt{\pi}} \tau e^{-\tau^2}$$

$$\text{where } \tau = \left(\frac{U}{2} \right) \left(\frac{t}{K} \right)^{1/2} \quad \text{is a dimensionless parameter,}$$

b is the undisturbed gradient before sedimentation, T is temperature, z is depth, U is the sedimentation rate, t is the time since the beginning of sedimentation, and K is the thermal diffusivity; dT/dz at $z=0$ is the present (measured) temperature gradient. A high sedimentation rate will reduce the surface heat flow. Therefore, depending on the sedimentation rate and its history in a lake, the

measured heat flow should be increased to correct for the sedimentary blanketing effect.

In Lake Malawi some C^{14} measurements at three different levels in one core in the central part of the lake indicated a high sedimentation rate in the range of 0.1 to 0.5 cm/yr (Von Herzen and Vacquier, 1967). The consistently low values of measured heat flow in the northern (deeper) part of this lake might be due to a higher sedimentation rate as compared to the rest of the lake. In the southern part, where the water depths are shallower than the rest of the lake, the sedimentation rate is, by comparison, effectively nil (Von Herzen and Vacquier, 1967). Therefore the heat flow measurements from the southern part of the lake are not corrected for sedimentation.

Using a thermal diffusivity $K = 2 \times 10^{-7} \text{ m}^2 \text{ s}^{-1}$ (Von Herzen and Vacquier, 1967) and a duration of sedimentation of 10^5 years, the average heat flow data are corrected for the northern and central parts of the lake. The duration of sedimentation at an average of about 0.2 cm/yr has been assumed to be 10^5 years, since a sedimentary thickness of about 200 m seems to be an estimate on the right order of magnitude. The heat flow (including the bottom water warming effect) is 39 ± 13 and $113 \pm 25 \text{ mW/m}^2$ for the northern and central regions, respectively. For the northern part of Lake Malawi, the corrected values of heat flow are calculated to be 72 ± 24 and $61 \pm 21 \text{ mW/m}^2$ for sedimentation rates of 0.4 and 0.3 cm/yr respectively. For the central region, the corrected values are 178 ± 40 and 152 ± 34 using sedimentation rates of 0.3 and 0.2 cm/yr.

In Lake Tanganyika the sedimentation rate is determined by C^{14} measurements to range from 0.03 to 0.05 cm/yr (Degens et al., 1971). If

these values are assumed to represent the minimum and maximum sedimentation rate for the whole lake, the corrected heat flow values are 71 ± 37 and 73 ± 38 mW/m² respectively.

The sedimentation rate in Lake Kivu is not known; however, the rate can be assumed to be similar to that of the shallower part of Lake Malawi. For this reason the minimum sedimentation rate has been assumed to be zero and the maximum rate has been estimated as 0.2 cm/yr. Using these sedimentation rates and the same thermal conductivity, the corrected heat flow values are calculated to be 73 ± 70 (not affected) and 99 ± 94 mW/m² respectively.

If the maximum corrections for sedimentation and bottom water warming are applied to the mean heat flow values of Lakes Tanganyika and Malawi, the revised heat flow values would be 73 ± 38 for Lake Tanganyika and 99 ± 16 mW/m² for Lake Malawi as a whole. The minimum corrections give values of 72 ± 37 and 84 ± 13 mW/m² respectively. For Lake Kivu only the correction for sedimentation has been applied since the bottom water is not anaerobic i.e. water circulates, and therefore the correction for bottom water warming is not applicable. The maximum and minimum corrections for sedimentations give heat flow values of 99 ± 94 and 73 ± 70 mW/m² respectively.

The average of all the measured values for the three lakes, taking into account the maximum and minimum corrections, gives a value of 90 ± 34 and 76 ± 27 mW/m² respectively. The value of 90 ± 34 mW/m² is here taken to represent the heat flow value of the western rift to compensate partly for the unknown hydrothermal factor. The above corrections have not been applied to the data from the eastern rift (111 ± 53 mW/m²) because the heat flow measurements were performed on land.

Although the measured heat flow is corrected for sedimentation and bottom water warming, the heat loss by hydrothermal circulation is not taken into account; so the actual heat flow could be higher. Also, the average heat flow values are not weighted with respect to area, because the sites from which data have been collected are scattered throughout the eastern rift and the lakes of the western rift.

The heat flow values of the eastern and western rifts of east Africa indicate that EARS, like other continental rift systems, is "hot". The values of 90 ± 34 and 111 ± 53 mW/m² compare closely to those of the Baikal (84 to 143 mW/m²), Rhine (109 ± 35 mW/m²) and Rio Grande (108 ± 27 mW/m²) rifts (see Table 1.2), and also with previous estimates for the eastern rift (105 ± 51) (Morgan, 1982).

5.5 REGIONAL GEOTHERMS

Geotherms (temperature as a function of depth) depend on the surface heat flow, the vertical distribution of heat sources and the variation of thermal conductivity with depth. The surface heat flow comprises the heat flowing from below into the upper layers of the crust plus the heat generated by radioactive decay within these layers. This can be written as:

$$q_0 = q^* + b A_0$$

where q_0 = surface heat flow; q^* = reduced heat flow (i.e. heat flowing from below the upper crust); b = characteristic depth of vertical source distribution (usually around 8-10 km); and A_0 = heat production of

the surface rocks (Birch et al., 1968; Roy et al., 1968; Blackwell, 1971).

On the global scale Pollack and Chapman (1977) derived an empirical relation between surface and reduced heat flow indicating that about 40% of the surface heat flow in any given region comes from the upper crustal zone enriched in radioactive material.

For the continental regions a family of geotherms have been constructed by Pollack and Chapman (1977). These geotherms, in which surface heat flow is the principal independent variable, are constructed using a steady-state conduction model with typical heat production of 2.5×10^{-3} mW/m³ for the granitic upper crust ($b = 8$ km) and 2.5×10^{-4} mW/m³ for the lower crust, assumed to consist of intermediate composition granulite facies rocks. From the base of the crust to the lithosphere/asthenosphere boundary a depleted ultrabasic zone with a characteristic heat production of 10^{-5} mW/m³ (down to 120 km depth) overlying pyrolite with heat production of 8.4×10^{-5} mW/m³ is used. The geotherms are shown in Fig. 5.4.

The petrology used in this model is not identical to the petrology of the rift as deduced from seismic data, and in the rift, as in oceanic regions, transient cooling plays an important role. However, geotherms are only weak functions of the petrology as long as the heat generation with depth follows approximately the standard model, and the geotherms for heat flow of 60 mW/m² or above are quite similar for the steady-state continental and transient oceanic models (see Fig. 3 of Pollack and Chapman, 1977). Therefore the geotherms of the continental model of Pollack and Chapman are applicable to EARS as a first approximation.

Fig. 5.4

Geotherm families of continental regions. Family parameter is heat flow in mW/m^2 . T_m is the mantle solidus. After Pollack and Chapman (1977).

Fig. 5.5

The geotherms for the eastern rift (III), the western rift (90) and the Precambrian shield (54 mW/m^2), and corresponding lithosphere thickness.

The heat flow values for the eastern rift and for the western rift are 111 ± 53 and 90 ± 34 mW/m² respectively. The heat flow outside the rift is 54 ± 28 mW/m². The corresponding geotherms, taken from Pollack and Chapman's model, are shown in Fig. 5.5. This figure also shows T_m and $0.85T_m$ where T_m is the estimated dry mantle solidus and $0.85T_m$ is the most likely estimate of the temperature at the top of the seismic low velocity zone (assuming the presence of volatiles such as CO₂ and H₂O which depress the solidus).

Since the heat flow measurements show large variations and the corrections are semi-quantitative, the inferred geotherms are subject to large uncertainties. Fig. 5.6 shows the geotherms for the eastern rift and western rift with their variability on the basis of the scatter in heat flow values. When taken in isolation, therefore, heat flow data provide only a weak constraint on the thickness of the lithosphere; the situation, however, improves when other geophysical data are included (see following sections).

A pattern of alternating "cold" and "hot" regions seems to exist along the western rift (see Table 5.7), with representative heat flows of 64 ± 15 and 139 ± 51 mW/m², respectively. Therefore, the thermal properties of the lithosphere may vary also along strike.

5.6 THERMAL THICKNESS OF THE LITHOSPHERE

The physical properties of the crust and upper mantle such as seismic velocities and attenuation, density, electrical conductivity and effective viscosity all depend on the homologous temperature T/T_m . Surface heat flow is an indication of the thickness of the lithosphere in that the thickness generally increases with decreasing surface heat

Fig. 5.6: Extreme geotherms and corresponding range of lithospheric thicknesses in (a) eastern, and (b) western rift.

flow.

Surface wave studies have demonstrated the existence of a shear wave low velocity zone in the mantle beneath oceanic and some continental regions. This low velocity zone, overlain by a high velocity lid, corresponds to the asthenosphere, and the high velocity lid to the subcrustal lithosphere. Heat flow and seismic studies in many regions of the continents have shown that in the regions of high heat flow there exists a well developed low velocity zone. This seems to disappear where the heat flow is less than about 40 to 50 mW/m², a heat flow value characteristic of the shields and platform regions. Thus the geotherm of a region is related to the depth to the low velocity zone, i.e. to the thickness of the lithosphere.

On the basis of an analysis of global heat flow data Pollack and Chapman (1977) showed that the low velocity zone is well correlated with heat flow data if the asthenosphere/lithosphere boundary is situated at the depth where the geotherm intersects the 0.85T_m isotherm. This conclusion is strengthened by heat flow and seismic studies in different regions, for example Australia (Goncz, 1974); the Canadian shield (Wickens, 1971); and the shield and platform of the United States (north-central and south-central regions respectively; Biswas and Knopoff, 1974).

A well-studied example is the area of Western North America (including the Basin and Range province) which is characterized by high mean heat flow of about 90 mW/m². Seismic studies in this region reveal a well developed low velocity zone beginning at a depth of 40 to 50 km. In Fig. 5.4 the 90 mW/m² geotherm intersects the 0.85T_m at 55 km depth indicating a remarkable agreement between the seismic and heat flow

data if the $T = 0.85T_m$ criterion is used.

Based on the above arguments Pollack and Chapman (1977) conclude that the lithosphere/asthenosphere boundary may begin at a temperature less than the solidus temperature and that $0.85T_m$ should be used to measure the thermal thickness of the lithosphere. This is natural, as shear wave velocity begins to decrease below the solidus. Also the mantle may be wet, in which case the water (and volatiles) would depress the solidus. Therefore it appears that the intersection of the geotherm with $T = 0.85T_m$ denotes a depth coincident with the top of the seismic low velocity zone, thus allowing the heat flow data to be used to measure the thickness of the lithosphere.

On the basis of the previous argument, the thermal thickness of the lithosphere in EARS is estimated to be approximately 40 to 55 km in the eastern and the western rifts, respectively (refer to Fig. 5.5). The lithospheric thickness in east Africa, in the areas outside the rift system (including the region between the two branches) is estimated to be 115 km. These estimates do not take into account the uncertainty in the heat flow data. Considering the variability on the heat flow values, the upper and lower bounds for the thickness of the lithosphere would be 17 to 92 km for the eastern rift, and 27 to 118 km for the western one. Furthermore, the existence of the "hot" and "cold" regions in the western rift indicates that the thickness of the thermal lithosphere varies along the strike of the rift (refer to Table 5.7). The thickness deduced from the hot and cold geotherms in this branch ranges from 30 km to 86 km. Although the variability is large, the good agreement with seismic data is an indication of the reliability of geotherms as estimators of lithospheric thickness.

5.7 ELECTRICAL CONDUCTIVITY

The knowledge of the distribution of electrical conductivity in the crust and upper mantle establishes limits on models for physical properties (mainly temperature) with depth. Geomagnetic deep sounding and magnetotelluric studies are two methods to measure the conductivity of the crust and upper mantle which in turn constrains the temperature structure.

Electrical conductivity is in general a function of temperature, composition, and water pressure (hydration). The relationship between the electrical conductivity and temperature can be expressed by:

$$\xi_c = \xi_0 e^{(E/kT)}$$

where ξ_c is electrical conductivity, ξ_0 and E are parameters characteristics of the material, k is the Boltzmann constant, and T is absolute temperature. Therefore, high values of conductivities within the crust and upper mantle may be due to a high geothermal gradient, or presence of conductive minerals, water and partial melting.

Magnetic deep sounding experiments were carried out in the Kenya rift near the Equator, 100 to 150 km north of Nairobi, by Banks and Ottey (1974) and Beamish (1976). These studies were based on an array of six recording magnetic variometers installed in the eastern rift. From these geomagnetic variation studies and the type of model which fits the experimental data, a strip of high conductivity material (about 0.1 S m^{-1}) at a depth of around 20 km was detected beneath the floor of the rift along the axis of the rift. In addition a thick zone (down to a depth of 100 km) of material of more moderate conductivity was also detected at a depth of no more than 50 km beneath the eastern

flank of the rift. These two regions of high conductivity were interpreted as zones of partial melting in the upper mantle. This interpretation seems to be valid since the position of the survey coincides with the regions of most recent (Late Quaternary) volcanic activity.

The results of geomagnetic deep soundings and magnetotelluric measurements at ten locations in and around the Kenya rift valley by Rooney and Hutton (1977) clearly confirm the existence of a good electrical conductor in the main rift valley of Kenya. Jiracek et al. (1979) also reached the same conclusion. Rooney and Hutton (1977) attribute the existence of high conductivity material to enhanced geothermal gradient and to the presence of electrolytic fluids in the rock pores and fissures beneath the rift zone.

The data gathered by the magnetotelluric method confirm and extend the conclusions drawn from the geomagnetic deep sounding experiments. From both sets of data it seems that the entire vertical section beneath the Kenya rift from 20 km down to the depth of 100 km (Banks and Beamish, 1979) is occupied by a zone of higher than normal conductivity. Electrical measurements do therefore confirm the presence of anomalously high temperature and/or electrolytic fluids beneath the rift system.

5.8 DISCUSSION.

Although quite variable, the heat flow measurements in EARS provide strong evidence that the rift system is associated with high heat flow and hot geotherms. By contrast, away from the rift the heat flow data seem to indicate values close to the typical Precambrian shield. The width of the high heat flow band seems to be no larger than

60±10 km for each of the western and eastern branches.

Within the rift itself the non-volcanic regions like Lake Tanganyika and parts of Lake Malawi reveal relatively low mean heat flow. On the contrary, high heat flow correlates with the volcanic sections, of which Lake Kivu and the eastern rift around the Equator are two examples. Also in the Kenya rift the heat flow varies locally. Higher heat flow values occur along the axial zone of the rift. The existence of high heat flow zones in the western branch and their correlation with the transverse features suggest that there is a definite heat flow pattern along strike as well. The average thickness of the lithosphere is calculated to be 40 and 55 km for the eastern and western rifts respectively, although this estimate is subject to a large uncertainty. The lithospheric thinning under the two branches of the East African rift system is clearly brought out by the heat flow in conjunction with other geophysical data.

Magnetotelluric and geomagnetic depth soundings in the eastern branch of EARS exhibit anomalously high values of electrical conductivity in two zones: one situated around 20 km of depth, and another extending from 50 km to 100 km of depth. The deep electrical studies support a model where ascending masses of hot material are displaced from the asthenosphere at the higher levels in the lithosphere. The initial heating for this rise can be attributed to upward transfer of heat by magma, hydrothermal activity, and/or thinning of the continental lithosphere.

If these zones of high electrical conductivity are associated with partial melting, they are likely to form the source from which the most recent volcanic rocks are derived. Therefore, the difference in

depth of these two sources and the degree of partial melting in the two branches of EARS might explain the difference in the chemistry of the rocks as well as the difference in the nature of the volcanic activity along the axis of the rift (see later chapters for discussion).

Fig. 5.7 shows a cross section of EARS using all available geophysical data. This profile, which is mostly representative of the area around the Equator, shows that despite the large error margins on the heat flow data, the resultant lithospheric thicknesses are fairly close to the ones modelled using gravity and seismic data.

The regional association of the rift zone with a pronounced upper-mantle seismic low velocity zone, negative gravity anomalies, low upper mantle density, high heat flow, and high electrical conductivity either in the crust itself or in the upper mantle, supports the conclusion that an anomalously hot, and probably partially molten asthenosphere is situated under the rift zone. Although individual sets of data (seismic, gravity, heat flow) are subject to large uncertainties, it is their convergence that highlights the differences in the lithospheric structure not only in the rift zone and the surrounding shield, but also between the two branches of the rift and along strike.

Fig. 5.7: A composite true-scale cross section of the east African rift system representative of the region around the Equator, incorporating gravity (this study), seismological (Savage and Long, 1985) and heat flow data model thickness ($0.85 T_m$; this study) of the lithosphere.

asthenosphere/lithosphere boundary inferred from:

- gravity anomalies
- - - seismic wave velocities
- heat flow

CHAPTER 6

THE RHEOLOGY OF THE LITHOSPHERE

6.1 INTRODUCTION

Lithospheric deformation depends on temperature, pressure, rock properties (e.g., composition, grain size, water content, etc.), and applied deviatoric stress. The lithosphere, at its simplest, consists of a top brittle layer overlying a ductile region, but the brittle-ductile transition can be repeated due to petrological layering. At low temperatures (shallow depths), the brittle strength (i.e. the stress at failure in the brittle regime) is less than the yield stress (the strength in the ductile regime) and brittle deformation predominates. The brittle strength is a linear function of pressure and therefore depth and as a first approximation does not depend on temperature. (See Paterson, 1978 for details on the brittle shear fracture criterion.) The yield stress, on the other hand, depends exponentially on temperature and decreases with depth. At some depth, the fracture stress exceeds the yield stress, which is rapidly decreasing with depth. Below this depth ductile deformation predominates.

Since geotherms are a function of surface heat flow, the thickness of the lithosphere beneath shields is larger than that beneath rift zones, taking the other properties to be equal in the two regions. Therefore, as stated above, there is a relationship between the thermal regime of the lithosphere and its rheological characteristics.

One way of determining the depth of brittle-ductile transitions and the thickness of the rheological lithosphere (defined as the outermost shell where the strength -brittle or ductile- is larger than some critical value) is to study the rheological properties of lithospheric material as a function of depth. The identification of this "rheological stratification" (Ranalli and Murphy, 1987) is possible using the deformation laws, for a given petrological model and a choice of geotherms, the latter of which is critical since the rheology is very sensitive to the temperature.

6.2 LAWS OF BRITTLE AND DUCTILE DEFORMATION

6.2.1 Brittle Deformation

Practically all rocks at relatively low temperatures and pressures are brittle solids and sufficiently large deviatoric stresses cause them to fracture. Consequently, fracture is important in upper crustal rocks (and possibly also in the uppermost mantle, as will be seen later). Given the compressive nature of all three principal stresses, the mode of failure is shear fracture and sliding along new or pre-existing fracture planes. Along these fracture surfaces (faults) a linear frictional failure criterion can be applied which is independent of temperature and rock type, and can be expressed as (Sibson, 1974)

$$\sigma_1 - \sigma_3 = \beta \rho g z (1 - \lambda)$$

where $\sigma_1 - \sigma_3$ is the difference between the maximum and minimum principal stresses, β is a parameter depending on the type of faulting (3, 1.2 and 0.75 for thrust, wrench, and normal faults, respectively),

ρ is the density of the material, g the acceleration of gravity, z the depth, and λ is the pore fluid factor, i.e. the ratio of pore fluid pressure to overburden pressure. If the pore fluid pressure is taken to be hydrostatic, $\rho_w g z / \rho g z = \rho_w / \rho$, and therefore $\lambda = 0.36$, an "intermediate" value which will be used in the calculations.

6.2.2 Ductile Deformation

The occurrence of ductile rather than brittle deformation depends on the relative magnitude of the ductile yield stress (or creep strength, the stress difference necessary to cause creep at geological strain rates) and brittle strength. If the critical stress difference for frictional sliding is less than the creep strength, brittle failure will occur; in the opposite case deformation will be by ductile mechanisms (Brace and Kohlstedt, 1980; Zuber and Parmentier, 1986).

Neglecting transient effects, power-law creep and Newtonian diffusion creep are the two most common steady-state (time-independent) flow mechanisms in silicate polycrystals at high temperature. Laboratory studies of creep of crystalline rocks give the following relationship between stress and strain rate (pressure effects are neglected since they are second-order with respect to temperature effects in the lithosphere-- see review by Kirby, 1983):

$$\dot{\epsilon} = A \sigma^n \exp(-Q/RT)$$

from which it follows that the stress (creep strength) necessary to achieve a given strain rate is

$$\sigma = (\dot{\epsilon}/A)^{1/n} \exp(Q/nRT)$$

where $\dot{\epsilon}$ is the strain rate, σ is the principal stress difference or twice the maximum shear stress, T the absolute temperature, and R the gas constant. A , n and Q are material parameters (the stress exponent n and the activation energy Q are temperature-independent, while A depends only weakly on temperature in comparison to the exponential term). In diffusion creep, A depends also on grain size; but the strain rate is independent of grain size in power-law creep. The stress exponent is $n=1$ for diffusion creep and $n=3$ for power-law creep. Experimental results indicate that power-law creep is the predominant mechanism of ductile deformation in the lithosphere for stresses of geodynamic significance.

6.3 RHEOLOGICAL PROFILES

The lithospheric models of Nolet and Mueller (1982) (see Chapter 4, Tables 4.6 and 4.7) based on the simultaneous inversion of teleseismic data are used to construct a rheological model for the lithosphere in east Africa. Model densities and seismic velocities are matched with rock types and are used to construct the petrological models (Fig. 6.1).

Using the criterion for frictional failure, the stress difference required to start frictional sliding as a function of depth is calculated for normal faults (the relevant type of fault in the rift region), using the actual rock densities of Nolet and Mueller's (1982) model and assuming hydrostatic pore water pressure at any depth (curve "a" in Figs. 6.2, 6.3 and 6.4).

The distribution of the ductile creep strength with depth depends on composition (rock type) and temperature. Different

Fig. 6.1: Petrological models based on the lithospheric model of Nolet and Mueller (1982).

Fig. 6.2: Creep strengths of several lithospheric materials in the eastern rift, assuming a steady-state strain rate of 10^{-15} s^{-1} and using the 111 mW/m^2 geotherm.

Fig. 6.3: Creep strength of several lithospheric materials in the western rift, with a strain rate of 10^{-15} s^{-1} and the 90 mW/m^2 geotherm. Symbols are the same as in Fig. 6.2.

Fig. 6.4: Creep strength of several lithospheric materials in the shield area, with a steady state strain rate of 10^{-15} s^{-1} and the 54 mW/m^2 geotherm. Symbols are the same as Fig. 6.2.

geological provinces have different temperature structures, as in the rift system of east Africa and the shield region surrounding it, i.e. transitions between the brittle- and creep-dominated behaviour in the lithosphere occur at different depths (shallower in hot regions and deeper in cold regions), and the strength-depth profiles are different. The brittle-ductile transition is, of course, gradual rather than sharp, so that the calculated depths are indicative of a range of a few kilometres rather than precise values.

Figs. 6.2, 6.3 and 6.4 give the creep strengths calculated for various rock types using the flow-law equation and the values for the creep parameters given in Table 6.1. Then, using the petrological models shown in Fig. 6.1, rheological profiles of the lithosphere are computed for a representative strain rate of $\dot{\epsilon} = 10^{-15} \text{ s}^{-1}$, and geotherms (from Pollack and Chapman, 1977) corresponding to heat flow values of 111, 90, and 54 mW/m², for eastern and western rifts and surrounding regions, respectively (Table 6.2). The strain rate of 10^{-15} s^{-1} is geologically appropriate for EARS and takes into account an overall extension of approximately 10 km since Late Miocene (10 Ma) in the Kenya rift (Baker and Wohlenberg, 1971).

For the eastern branch, a two layer crust is used with the upper 20 km being composed of plagioclase-rich rocks (modelled with the rheology of anorthosite), overlying 20 km of basic lower crust (modelled with the rheology of diabase) and a dunite (olivine) upper mantle. The petrological model of the western rift consists of a 15-km thick plagioclase-rich upper crust, a 20-km thick basic lower crust and an olivine upper mantle. The shield is assumed to have a 20-km thick plagioclase-rich upper layer, a 20-km thick basic lower crust,

Table 6.1: Creep parameters for lithospheric materials

Material	$A(\text{GPa}^{-n}\text{s}^{-1})$	n	$Q(\text{kJ mol}^{-1})$
: Granite	: 5.0	: 3.2	: 123
: Granite (wet)	: $1.0\text{E}+2$	: 1.9	: 137
: Quartzite	: $1.0\text{E}+2$	: 2.4	: 156
: Quartzite (wet)	: $2.0\text{E}+3$	: 2.3	: 154
: Quartz Diorite	: $2.0\text{E}+4$	: 2.4	: 219
: Albite rock	: $1.3\text{E}+6$	: 3.9	: 234
: Anorthosite	: $1.3\text{E}+6$	: 3.2	: 238
: Diabase	: $3.2\text{E}+6$	: 3.4	: 260
: Olivine	: $4.0\text{E}+15$	: 3.0	: 540

Sources: Kirby, 1983 Ranalli and Murphy, 1987

Table 6.2: Temperature structures in the eastern rift, western rift, and the shield region.

z	S	WR	ER
	54	90	111 mW/m^2
: km	Temperature ($^{\circ}\text{K}$)		
: 5	448	500	573
: 10	520	590	650
: 20	600	760	860
: 30	680	920	1080
: 40	770	1080	1300
: 50	860	1220	1510
: 60	960	1450	1650
: 70	1050	1550	1800
: 80	1130	1700	1950
: 90	1180	1820	2100
: 100	1390	1930	2200
: 110	1330	2000	----
: 120	1530	----	----

: z = depth, S = shield, WR = :
: western rift, ER = eastern rift, :

and an olivine upper mantle. The resulting profiles of strength versus depth are shown in Fig. 6.5 (a,b,c).

The rheological models indicate that the topmost brittle part of the crust in the eastern rift is confined approximately to the upper 12 km of the crust (Fig. 6.5a). However, for the western branch of EARS the brittle layer continues down to a depth of about 18 km, below which the deformation is ductile (Fig. 6.5b). In both cases, the Moho is a rheological discontinuity involving a sudden change in creep strength. The "hot" areas of the western rift, represented by the 139 mW/m^2 geotherm (see Chapter 5) indicate a 10 km-thick brittle layer. The cold regions in this area, represented by the geotherm of 64 mW/m^2 on the other hand indicate a 25 km-thick brittle layer, a ductile layer from 25 to 37 km and another brittle layer up to the depth of 50 km below which the deformation is ductile. For the shield region, the rheological model shows that the brittle zone has a thickness of 57 km, below which the deformation is ductile. A ductile but not weak layer is possibly present between 29 and 40 km of depth (Fig. 6.5c).

The base of the lithosphere in the rheological sense can be defined as the depth at which the yield stress is reduced to some low value (say, $\sigma = 1 \text{ MPa}$) and below which there are no further discontinuities in the creep strength. According to this definition, the lithospheric thicknesses for the eastern rift, western rift, and the Precambrian shield are 41, 55 and 115 km respectively. The lithosphere in the hot regions of the western rift is thinner than the lithosphere in the eastern rift (30 km for 139 mW/m^2 compared to 41 km). This is consistent with the presence of active volcanoes in the hot areas of the western rift, and reflects along-strike variations. In the cold regions, represented by the geotherm of 64 mW/m^2 , the

Fig. 6.5: Rheological models for: (a) eastern rift, (b) western rift, and (c) the Precambrian shield. σ_L is the inferred lithospheric strength.

thickness of the lithosphere is 100 km.

The lower boundary of the lithosphere can also be defined as the depth at which the viscosity reaches a value characteristic of the asthenosphere, say 10^{21} Pa s (see for example Walcott, 1973 and Cathles, 1975). The effective viscosity is the ratio between stress and strain rate, i.e. from the flow law and $\eta_{\dot{\epsilon}} = \sigma / \dot{\epsilon}$:

$$\eta_{\dot{\epsilon}} = \frac{\dot{\epsilon}^{(1-n)/n}}{A^{(1/n)}} \exp \frac{Q}{nRT}$$

Table 6.3 shows the effective viscosity as a function of depth for the east African region, and shows variations in lithospheric thickness in the rift zone and the Precambrian shield as well as along-strike variation in the thickness in the western branch. The effective viscosities gradually decrease with depth, and are in general lower in the rift compared to the shield region of east Africa. Using the above criteria, the thickness of the lithosphere is between 30 and 40 km for the eastern rift, between 50 and 60 km for the western rift and between 90 and 100 km for the shield.

The comparison of the lithospheric thicknesses deduced from yield stress criterion ($\sigma = 1$ MPa) and the effective viscosity criterion ($\eta_{\dot{\epsilon}} = 10^{21}$ Pa s) indicates that the two criteria give approximately the same results and are therefore interchangeable.

The rheological profiles are constructed using rheological parameters determined from laboratory experiments and various inferred geotherms. Taking into account the uncertainties in the parameters used in the flow laws (as they are extrapolated to geological conditions),

Table 6.3

Viscosity (in Pa s) of the crust and upper mantle in the rift zone and the Precambrian Shield of east Africa. Since the heat flow of the cold western rift is similar to that of the Precambrian shield, their corresponding viscosities are of comparable value and are combined in the Table.

z (km)	ER 111	WR 90 (mW/m ²)	HWR 139	S= CWR 54
: 10	2.4E+23	4.7E+23	1.4E+23	7.5E+24
: 20	8.3E+21	1.6E+23	2.5E+21	7.5E+23
: 30	2.4E+21	1.3E+22	1.2E+21	3.5E+23
: 40	5.6E+20	3.2E+21	3.0E+20	7.3E+22
: 50	1.1E+20	2.0E+21	-----	5.4E+24
: 60	-----	1.5E+20	-----	3.9E+23
: 70	-----	-----	-----	5.7E+22
: 80	-----	-----	-----	1.3E+22
: 90	-----	-----	-----	5.9E+21
: 100	-----	-----	-----	2.8E+21
: 110	-----	-----	-----	1.2E+21
: 120	-----	-----	-----	7.4E+20
: z=depth; WR=western; ER=eastern rift; :				
: H= "hot"; C= "cold" :				

Table 6.4

Uncertainties in flow law parameters

Rock type	A GPa ⁻ⁿ s ⁻¹	n	Q KJ mol ⁻¹
: Anorthosite	(1.4 _± 1.3)10 ⁶	3.2 _± 0.5	238 _± 27
: Diabase	(3.2 _± 2.7)10 ⁶	3.4 _± 0.6	266 _± 29
: Olivine	(4.0 _± 1.3)10 ¹⁵	3.0 _± 0.5	540 _± 61

3

NEC

the lack of precise knowledge of the mineralogy, the possible roles played by water and/or other volatiles, and the uncertainties in the geotherm, the profiles must be taken as indicative only. Nevertheless, their characteristics relative to one another are in all likelihood sufficiently reliable.

A quantitative assessment of uncertainties on the depth of brittle-ductile transition and lithospheric thickness can be obtained by considering the uncertainties in each parameter used in the flow law equation. Table 6.4 shows the uncertainty of each parameter (Shelton and Tullis, 1981; Kirby, 1983). The uncertainty in n and Q for anorthosite and diabase are calculated using the same percentage uncertainty in these values as was measured for olivine. The maximum possible value for the depth to the brittle-ductile transition and lithospheric thickness are to a first approximation determined using the minimum values for A and n and maximum values for Q and $\dot{\epsilon}$ (taken as 10^{-14} s^{-1}). The upper boundary of rheological profiles is calculated using maximum values for A and n and minimum values for Q and $\dot{\epsilon}$ (taken as 10^{-16} s^{-1}). The lower and upper boundaries of rheological profiles are plotted in Fig. 6.6. It is seen that the depth and depth-uncertainties of the brittle-ductile transition for the eastern rift, western rift and the Precambrian shield are 12 ± 5 km, 18 ± 8 km and 57 ± 15 km respectively. The lithospheric thicknesses and their uncertainties for the corresponding regions are 41 ± 7 km, 55 ± 9 and 115 ± 28 km respectively. The rheological profiles of eastern and western rifts show extra brittle layers at the depths of 20 to 28 and 35 to 40 km respectively, when the maximum depths of the brittle-ductile transition and the lithosphere are calculated.

The uncertainties shown in the depth to the brittle-ductile

Fig. 6.6: The upper and lower boundary of the rheological profiles using the uncertainties in the flow-law parameters given in Table 6.4.

transition and in the lithospheric thickness were calculated choosing those values for the flow law parameters which produced the maximum uncertainties possible. It is likely, however, that the error is considerably less since all variables would not simultaneously have values that would produce the maximum deviations. On the other hand, the above uncertainties reflect only the uncertainties in rheological parameters and not in the geotherms.

Since seismicity is mainly confined to the brittle layers of the lithosphere, the depth distribution of earthquakes in different areas is a possible check for the reliability of the rheological models. The seismicity in EARS is mostly concentrated at a depth of 5 to 15 km which is within the brittle regime defined by the rheological models and at a depth of 30 to 40 km. The high concentration of events at the depth of 33 km (Fig. 4.6 and Appendix 4.1) is suspicious, in that some events might have arbitrarily been assigned to this depth without proper depth determination to indicate "lower crustal" or "just subcrustal" shocks (see Chapter 4 for discussion). At first sight, the earthquake focal depths do not fully coincide with brittle zones as expected from the rheological profiles. This is because EARS, especially the western branch, is characterized by along strike variation in the lithospheric structure (as evidenced from gravity, seismicity and heat flow data) but the rheological profiles take into account the average heat flow value of the rift and are not sensitive to the alternating hot and cold regions. Besides, the average heat flow value of the floor of the eastern rift represent the values from 1°N to 1.5°S where macroseismic activity is practically nonexistent. The lack of heat flow data from south of 1.5°S restricts the possibility of comparing the depth of the brittle ductile transition from the

rheological profiles to the earthquakes focal depths in the region. Nevertheless, disregarding the 33 km deep events, the seismicity in EARS along the axis of the eastern and western rifts is mostly concentrated at a depth of 5 to 10 km in the eastern rift and 0 to 15 km in the western rift (Fig. 4.6) (see section on focal depth, Chapter 3). These depths coincide with the depth of the brittle layers in the eastern rift (12+5 km) and the western rift (18+8 km) deduced from the rheological profiles.

6.4 THE TOTAL STRENGTH OF THE LITHOSPHERE

One of the basic uses of rheological models of the lithosphere is the calculation of "total lithospheric strength" (Kusznir and Park, 1984).

The area delimited by each curve in Figs. 6.5a, 6.5b and 6.5c gives the force per unit width necessary to deform the lithosphere at a constant strain rate of 10^{-15} s^{-1} . These areas have been calculated using Simpson's rule of integration and their values are 1.09×10^{12} , 3.53×10^{12} and $2.16 \times 10^{13} \text{ N m}^{-1}$ for the eastern rift, western rift and the Precambrian shield respectively. Normalizing the above values by the total thickness of the lithosphere defined by the rheological criterion, i.e. depth at which $\sigma = 1 \text{ MPa}$, gives the "total strength" σ_L :

$$\sigma_L = \frac{1}{L} \int_0^L \sigma dz$$

where: σ_L = the total strength of the lithosphere; $\sigma = (\sigma_1 - \sigma_3)$ = differential stress; L = the rheological thickness of the lithosphere

and z = depth. Lithospheric strengths of 27, 62 and 188 MPa are obtained for the eastern and western rift and the Precambrian shield, respectively.

Since in east Africa rifting developed within the Precambrian shield, the force per unit width necessary to initiate rifting is that necessary to overcome the strength of the Precambrian shield. There is a close agreement between the force per unit width necessary to initiate rifting in EARS ($2.16 \times 10^{13} \text{ N m}^{-1}$) and the one necessary for mechanical failure of the lithosphere throughout its thickness in a plate boundary such as northern Red Sea region ($2.10 \times 10^{13} \text{ N m}^{-1}$, Steckler and ten Brink, 1986).

As expected, the lithospheric strength is lower for the eastern rift than the western rift, and again lower for the western rift compared to the shield. However, the failure of the lithosphere does not depend only on its strength, since the ductile deformation in the lower lithosphere causes stress relaxation in the region undergoing ductile flow, which is accompanied by redistribution and amplification of stress in the upper lithosphere leading to its fracturing. In other words, the ductile flow in the lower lithosphere concentrates stress in the upper lithosphere ("stress concentration factor" of Kusznir and Park, 1984) which results in complete brittle failure in the upper lithosphere when the concentrated stress level becomes larger than the failure strength. The higher the heat flow, and consequently the thinner the brittle layer, the higher will be the stress concentration factor.

"Whole Lithospheric Failure" (WLF) (Kusznir and Park, 1984) will occur when the competent elastic layer yields throughout and the whole

Fig. 6.7: Field of significant tensional lithosphere deformation (dotted) compared with estimated stress levels for possible sources of tensional stress (such as plateau uplifts and continent margin stresses) (after Kusznir and Park, 1984).

lithosphere is undergoing deformation, either brittle or ductile. Fig. 6.7 (from Kusznir and Park, 1984) shows the critical stress necessary for regions with different heat flow to undergo WLF. From this diagram it seems that the extensional stress required for failure in the Precambrian shield lithosphere with the heat flow value of 54 mW/m^2 is 35 MPa. Therefore an average tensional stress of 35 MPa throughout the lithosphere is required to initiate rifting; this is likely to reflect the initial condition of the rifting. Once rifting has begun, the lithosphere weakens and lower stresses are required to maintain extension.

In applying Fig. 6.7 to EARS several factors should be considered. In this model, the Griffith failure phenomenon is used for the brittle upper lithosphere. The brittle strength is controlled by the values of tensile strength (T_0) and coefficient of friction (μ). Laboratory experiments indicate that values for T_0 and μ range from 10 to 40 MPa and 0.5 to 1.0 for crystalline rocks (Brace, 1964; Jaeger and Cook, 1971; Byerlee, 1978). Values of 20 MPa and 0.5 are used in Fig. 6.7 for T_0 and μ respectively. Also although the creep parameters (A , n , Q) used by Kusznir and Park (1984) are fairly close to the one used in this study, their petrological model, consisting of a wet quartzite upper crust and a dry quartzite lower crust, differs from the ones used here. As a consequence, therefore, the rheological structure is different.

The strength of the lithosphere in EARS and the possibility of WLF in the region allow evaluation of different proposed sources of tectonic stresses in the rift zone (see Chapter 1, section 1.7). Membrane stresses due to the northward movement of the African plate on a non-spherical Earth were calculated by Oxburgh and Turcotte (1974).

Given a movement rate of 0.25 degree/Ma beginning at 100 Ma before present, the cumulative stress is calculated to be 13.5 MPa in the centre of a plate of the size of the African plate. This stress is insufficient to initiate the rifting process, although it could be significant when combined with other sources of stress. However, after rift initiation membrane and thermal stresses are not likely to be the major source for continued tectonic deformation because they are "non-renewable" (Bott and Kusznir, 1984), i.e. they are not continuously generated at plate boundaries.

Other possible kinds of stresses are associated with plateau uplifts (up to 30 MPa) and with isostatically compensated loads in the continental lithosphere (of the order of 50 MPa) (Bott and Kusznir, 1984). These kinds of stresses are "renewable", and are most likely the source for much of the tectonic deformation in the rift zone. Fig. 5.7 shows that plateau uplifts can generate tensional stresses of the order of 30 MPa which are very close to the magnitude of the stress required for failure in the lithosphere of the Precambrian shield, i.e. the condition for initiation of the rift. It is therefore possible that a combination of intraplate stresses could have initiated the rifting. The stresses generated at the plate boundaries (interplate stresses), such as "ridge push" could also have played a role, although this is difficult to quantify.

CHAPTER 7

GENERAL INTRODUCTION TO THE BIRUNGA FIELD

7.1 INTRODUCTION

The Tertiary to Recent alkaline volcanic rocks of the western branch of EARS occupy three areas. From north to south these constitute the: (i) Toro-Ankole, (ii) Birunga (Virunga), and (iii) south Kivu fields (Fig. 2.8). Although the present study deals mainly with the volcanic rocks of the Birunga field, also included are some samples from the Toro-Ankole field.

Igneous provinces that include rock types similar to those found in the Birunga field are the Roman region, Italy (Cundari and LeMaitre, 1970; Appleton, 1972), New South Wales and Kimberley, Australia (Prider, 1960; Cundari et al., 1978), the East Eifel region, Germany (Hoefs and Wedepohl, 1968; Wörner et al., 1986), Leucite Hills (Carmichael, 1967; Vollmer et al., 1984), and Highwood and Bearpaw Mountains, U.S.A. (Larsen et al., 1941; Chadwick, 1970).

The Birunga field occupies the extreme southwest corner of Uganda (at border of Zaire, Rwanda and Uganda), and stretches about 80 km north and northeast of Lake Kivu, occupying an area of about 3500 km² (Fig. 7.1). The Bufumbira area forms the extreme northeastern corner of the Birunga field, covering regions in Rwanda and Uganda (Combe, 1933; Holmes and Harwood, 1937) (Fig. 7.1). A detailed discussion of the Birunga field is given in section 7.2.

The Toro-Ankole field lies about 160 km north-northeast of

Fig. 7. The location of the eight major volcanoes of the Birunga field; After Holmes and Harwood (1932). The dashed lines show the international boundaries.

Birunga field to the east and southeast of the Ruwenzori massif (northeast of Lake Amin, formerly Lake Edward) and contains numerous areas of tuffs and associated explosion craters. A suite of Middle-Late Pleistocene volcanics penetrate the Precambrian basement and Pleistocene lacustrine-fluvial deposits of the Kaiso Series. The Toro-Ankole field is composed of several small, isolated, volcanic areas. The isolated areas are from north to south: Rusekere, Fort Portal, Ndale (formerly Kyatwa), Katwe-Kikorongo, Bunyaruguru (formerly Kichwamba) and Katunga (Holmes and Harwood, 1932). The activity of the Toro-Ankole field is dated at Middle to Late Pleistocene, terminating at the beginning of Holocene (Bagdasaryan et al., 1973). Although the structural relations are not well known, it appears that the volcanic activity is associated with SSW-NNE faults.

The South Kivu field, Eocene to Pleistocene in age, lies southwest of Lake Kivu and covers a total area of about 5500 km² (Holmes, 1940). Basalts and alkaline trachytes are the main rock types. Most of the volcanic activity took place during the Miocene (Pouclet, 1977; Bellon and Pouclet, 1980), along N-S and SW-NE faults associated with the rift. The basaltic lavas of the South Kivu field are unusually poor in K.

7.2 THE BIRUNGA FIELD

The Birunga field, first studied by Von Götzen (1896), consists of 8 major volcanoes (Fig. 7.1) and over 100 minor vents. From east to west the major volcanoes are: Muhavura, Mgahinga, Sabinyo, Visoke, Karisimbi, Mikenno, Niragongo (Nyiragongo, Ninagongo), and Nyamлагira (Nyamuruguru). The last two volcanoes are still active. The lava

flows of these major volcanoes and numerous other minor volcanoes form a natural dam at the northern side of Lake Kivu. Niragongo contains a permanent lava lake.

7.2.1 SPATIAL EVOLUTION AND THE AGE OF THE VOLCANOES OF THE BIRUNGA FIELD

Geographically the major volcanoes of the Birunga field can be divided into three groups (Kampunzu et al., 1986 and in press):

- 1) an eastern group including Mgahinga, Muhavura, and Sabinyo;
- 2) a central group including Visoke, Karisimbi and Mikeno;
- 3) a western group including the active volcanoes of Niragongo and Nyamulagira (see Fig. 7.1).

Sabinyo (at the border of Uganda, Rwanda and Zaire) and Mikeno are the oldest volcanoes of the Birunga field, and both are greatly dissected by erosion. Some of the Sabinyo flows are older than the Mikeno volcanics. The K/Ar dates of between 0.3 and 0.14 Ma (Bagdasaryan et al., 1973; Katabarwa, 1983) for the youngest flows from the Sabinyo and Mgahinga volcanoes mark the final activity of the eastern group volcanoes. The K-Ar dating of the Mikeno volcanics reveals major activity between 2.6 and 0.8 Ma and later activity (deduced from the activity of the lava lake) between 0.5 Ma and the present (Guibert et al., 1975).

Potassium-argon dates from the central group of volcanoes indicate that this group is older than the western group. Dating of the samples from Mikeno give K-Ar ages of 2.6 ± 0.4 Ma (Guibert, 1978).

Nyamulagira and Niragongo, the two westernmost volcanoes, are still active. On December 26, 1981 a 400 m-long fissure opened between the two active volcanoes which reactivated Nyamulagira after its last

eruption on February 23, 1980. Nyamulagira has erupted 15 times during the interval 1894 to 1981 and has extruded a far greater volume of lava than Niragongo (Tazieff, 1977). In 1977 Niragongo drained about 20×10^6 m³ of melilitite nephelinite lava from its 3270 m suspended lava lake. In 1976-1977 lavas 1 m thick erupted from both volcanoes; the measured temperature was 1000 °C (Tazieff, 1977). Both Nyamulagira and Niragongo are characterized by violent explosive activity.

The ¹⁴C dating of organic materials in volcanic ashes from Lake Kivu indicate an age of 15,000 to 25,000 years for the formation of the active volcanoes of the western group (Degens et al., 1973; Pouclet, 1977). This estimate is corroborated by estimates of age based on the total volume of volcanics and the mean annual flow of the various volcanoes (Pouclet, 1975).

From the age of the volcanoes it appears that volcanic activity started first in the east and continued later in the west (Guibert, 1978). Kampunzu et al. (1986) relate the volcanic activity to movement of fracture zones that lie normal to the major rift faults.

7.2.2 VOLCANIC CYCLES IN THE BIRUNGA FIELD

The mineralogy and ages of the Birunga volcanics indicate that this field contains two distinct volcanic cycles. The initial cycle (Oligocene to Miocene; 28 ± 1.4 Ma to 8.9 ± 0.5 Ma) is characterized by a series of sodic lavas (Bellon and Pouclet, 1980), restricted to north and west of the Birunga field. Also associated with this cycle are tholeiitic basalts (Kampunzu et al., 1983a). This was then followed by the eruption of a series of Na-alkaline rich lavas such as basanites and alkali basalts similar to those seen in the South Kivu field

(Kampunzu et al., 1982). The volcanism in the latter is, for the most part, older than that of Birunga (Sahama, 1972). This initial cycle is associated with rift faults oriented SW-NE and N-S.

The second cycle formed the eight major volcanoes and took place during Plio-Pleistocene to Recent time. Rocks of the second cycle are strongly undersaturated and alkalic and are concentrated along faults oriented SW-NE and SSE-NNW, transverse to the floor of the rift (Pouclet, 1977).

The chemical analyses of the rocks of the two volcanic cycles in the western branch of EARS indicate that the alkali content ($K_2O + Na_2O$) of the magmas has increased both in time and in space (from south to north) and that they have also changed in composition from sodic to strongly potassic (Pouclet, 1980).

7.3 ASSOCIATION OF VOLCANOES WITH GEOLOGICAL FEATURES

Holmes and Harwood (1937) state that there is an underlying relationship between rift movement, the direction of rift faults and magma generation. The direction of the rift faults and their intersections, therefore, might have a significant effect on the magma type. Evrard and Jones (1963) note that a number of faults in the Lake Kivu region lie mainly ~~northeast-southwest~~, in the general direction of the rift. There are also fractures that strike north-south and even northwest-southeast, at right angles to the rift faults. This observation is also supported by seismological and photogeological data (Villeneuve, 1980).

Seismological and photogeological studies also show that the Birunga volcanoes are located on faults of different orientations.

Nyamlagira, for example, is located on a major fault oriented NNW-SSE, while Niragongo and Mikenno lie on a N-S fault (Pouclt, 1977). Sahama (1972) pointed out that the active volcano, Niragongo, is located at the fracture intersections, while Nyamlagira is located on a SSE-NNW fracture zone, different both from the general SSW-NNE trend of the main rift, and the approximate E-W trend of the Birunga field.

7.4 ROCK SAMPLES

The rock samples for this study are part of a collection made by Dr. Keith Bell in 1965. A total of 38 samples from the Birunga field and 5 samples from the Toro-Ankole field were analysed. The 38 samples of the Birunga field are from 20 volcanoes and the 5 other samples are from the Katwe-Kikorongo, Katunga and Bunyaruguru regions of the Toro-Ankole field. Many of these samples are from the collection used by Holmes and Harwood (1937) in their classic memoir "The Volcanic Area of Bufumbira". Table 7.1 shows the sample number, rock type, volcano and the corresponding volcanic field for each sample. Figs. 7.2 and 7.3 indicate localities for the Birunga samples.

7.5 CLASSIFICATION AND NOMENCLATURE

The volcanic rocks of the western branch of the east African rift system are unusual and their names are not familiar to most petrographers. The classification adopted in this study is that of Holmes and Harwood (1937). Although there have been several recent attempts to reclassify the K-rich rocks (Borton, 1979; Foley et al., 1987), the scheme of Holmes and Harwood (1937) is chosen because it

Table 7.1: Sample numbers (C), rock type (R.T.), volcano (VOLC.) and the volcanic field for all the samples analyzed.

C	R.T.	VOLC.	FIELD	C	R.T.	VOLC.	FIELD
999	13	20	K.	2888	1	1	Bir.
1904	4	3	Bir.	2889	8	1	"
1906	3	15	"	2894	1	1	"
2766	7	3	"	2895	8	1	"
2768	7	3	"	2896	6	9	"
2772	3	18	"	2897	10	9	"
2774	6	12	"	2905	11	11	"
2777	3	12	"	2909	2	2	"
2781	6	4	"	2911	2	2	"
2782	10	13	"	2912	2	2	"
2792	10	13	"	2915	4	10	"
2794	3	3	"	3021	5	22	"
2799	9	3	"	3022	5	5	"
2805	6	17	"	3023	11	5	"
2816	4	6	"	3032	4	8	"
2820	10	13	"	3033	12	8	"
2824	3	16	"	4414	14	19	Kat.
2829	3	3	"	5628	10	24	K.
2849	6	3	"	6066	15	21	Bun.
2856	4	14	"	6095	12	21	"
2871	9	7	"	9872	5	23	Bir.
				9958	11	5	"

Rock-Types

1. Latite	6. Murambite	11. Nepheline Leucitite
2. Mikenite	7. Leucite Basanite	12. Kentallenite
3. Absarokite	8. Shoshonite	13. Ankaratrite
4. Kivite	9. Banakite	14. Katungite
5. Nephelinite	10. Olivine Leucitite	15. Mafurite

Volcanoes

1. Sabinyo	9. Kako	17. Kigoma
2. Miken	10. Rulenga	18. Nyakimango
3. Muhavura	11. Kitale	19. Katunga
4. Mihanga	12. Lwankone	20. Nyamununka
5. Niragongo	13. Lutale	21. Bunyaruguru
6. Lugandable	14. Mgahinga	22. Bushwaga
7. Bitare	15. Nyarobebsa	23. Goma
8. Nyamlagira	16. Nyagisenyi	24. Mbuga

Bushwaga and Goma are two subsidiary cones of Niragongo volcano.

Bir. = Birunga Bun. = Bunyaruguru Kat. = Katunga K. = Katwe-Kikorongo

Fig. 7.2: The locations of the samples from the Birunya-Bufumbira field.

2. 1904	3. 1906	4. 2766	5. 2768	6. 2772	7. 2774	8. 2777	9. 2781
10. 2782	11. 2792	12. 2794	13. 2799	14. 2805	15. 2816	16. 2820	17. 2824
18. 2829	19. 2849	20. 2856	21. 2871	22. 2888	23. 2889	24. 2894	25. 2895

Fig. 7.3: Samples locations in the Birunga field. Sources: Holmes and Harwood (1937) and Bell's (1965) field notes. The big symbols show the volcanic vents, the equivalent smaller ones indicate sample locations.

26. 2896	27. 2897	28. 2905	29. 2909	30. 2911	31. 2912	32. 2915
33. 3021	34. 3022	35. 3023	36. 3032	37. 3033	42. 9872	43. 9958

covers the rock types for Birunga and because of the lack of consensus in the more recent attempts at classification. The rock types of the Birunga field were divided by Holmes and Harwood (1937) into 6 series, namely:

(1) The olivine-leucitite series. This series contains olivine leucitite, olivine nephelinite and olivine melilitite. The Mikenite type of leucitite is named mikenite, characterized by considerably less MgO and notably more Na_2O than the average leucitites;

(2) The melilite-feldspathoidal series contains melilite nephelinite, melilite leucitite and katungite (K-rich olivine leucitite);

(3) The leucite-basanite series which ranges from murambite (feldspathic basanite or melanocratic kivitite) through kivitite to leucite basanite;

(4) The absarokite-shoshonite-banakitite series;

(5) The potash-trachyandesite-latitude series contains latites and shoshonites. This group is represented by lavas from only one volcano, Sabinyo;

(6) The limburgite-trachybasalt series.

Table 7.2 summarizes the petrology of the main volcanic rocks of the Birunga and Toro-Ankole fields analyzed in the present study. Detailed petrographic descriptions of some of the samples analysed in this thesis are given in Holmes and Harwood (1937).

Table 7.2: Petrographic Summary of the Rocks of the Birunga
and Toro-Ankole Fields (Holmes and Harwood, 1937).

ROCK TYPE	PHENOCRYSTS	GROUNDMASS
Absarokite	corroded olivine, augite, rare biotite, plagioclase more than murambite, black ores as inclusions	augite, ilmenite, magnetite, few leucite, plagioclase (labradorite), xenoliths of rounded quartz with green augite reaction rims, orthoclase
Ankaratrite	augite, olivine, occasional perovskite, black ore	augite, biotite, leucite, black ore
Banakitite	augite, biotite, hornblende rare xenocrysts of olivine, plagioclase, apatite, ores	biotite , augite, alkali feldspar (sanidine) apatite, ore, groundmass similar to absarokite but higher proportion of feldspar
Leucite Basanite	augite, olivine, leucite, plagioclase & ores	leucite, plagioclase, augite, rimmed quartz inclusions
Katungite	mellilite, olivine, no augite perovskite and black ore	nepheline, kalsilite, leucite, biotite
Kentallenite	augite, olivine, black ore	mainly plagioclase laths, black ore, few biotite flakes, common apatite needles, leucite, olivine
Kivite	titan-augite, olivine, plagioclase, leucite, biotite and magnetite	augite, abundant plagioclase (mostly labradorite), leucite, apatite (conspicuous), magnetite

Table 7.2 ctd.

ROCK TYPE	PHENOCRYSTS	GROUNDMASS
Latite	hypersthene, plagioclase, biotite, ore, olivine (rare accessory), apatite, and ore	augite, alkali feldspar, laths of hypersthene & no olivine
Nepheline leucite	leucite and minor nepheline	leucite, abundant corroded augite, few nepheline, rare corroded olivine and a few apatite needles (some varieties are free from olivine)
Olivine leucite	olivine, leucite, and minor augite	mostly leucite, augite, glass, a few olivine, black ores, no feldspar
Mafurite	augite, olivine	augite, olivine, perovskite, black ore, kalsilite, biotite, glass
Mikenite	leucite, titan-augite black ore, no olivine	augite, rare olivine, little perovskite, nepheline, leucite, melilite, apatite & glass
Murambite	olivine, feldspar, rare augite, quartz xenocrysts	abundant augite, black ores, laths of plagioclase, orthoclase, leucite, glass
Nephelinite	nepheline, melilite, accessory olivine	nepheline, leucite, melilite, rare olivine, perovskite, titan-augite, magnetite
Shoshonite	corroded olivine, augite, few biotite, plagioclase, apatite inclusions & ore	mostly alkali feldspar (oligoclase & orthoclase) augite, rare biotite, & ore

CHAPTER 8

PETROCHEMISTRY

8.1 INTRODUCTION

The general petrography of the lavas of the Birunga as well as Toro-Ankole fields has been described by Combe and Simmons (1933). Subsequently Holmes and Harwood (1937), in their classic work, carried out a detailed petrographic and geochemical study of the Birunga rocks. Since then several researchers have analyzed these rocks; e.g. Higazy (1954), Bell and Powell (1969), Bell and Doyle (1971), Gerasimovsky and Polakov (1974), Mitchell and Bell (1976), Gerasimovsky (1977), Hawkesworth and Vollmer, (1979), Vollmer and Norry (1983a;b), and Vollmer et al. (1985).

8.2 MAJOR, TRACE AND RARE EARTH ELEMENT CHEMISTRY

Some new chemical analysis of 38 samples from both the Birunga and Toro-Ankole fields are given in Table 8.1. This table includes the whole-rock analyses for major elements as well as Ba, Cr, Nb, Ni, Rb, Sr, Y, Zn and Zr determined by X-ray fluorescence. The analytical details are given in Appendix 8.1. The CIPW normative mineral compositions for these rocks are given in Appendix 8.2.

The new data highlight the following features:

- 1) silica contents that range between 34 to 61 weight % (herein after indicated as %).

Table 8.1

Whole-Rock Major and Trace Element Analyses

C	SiO ₂	Al ₂ O ₃	Total Fe	MgO	CaO	Na ₂ O	K ₂ O	TiO ₂	P ₂ O ₅	MnO
999	41.9	8.9	13.6	9.0	13.6	1.9	3.9	4.8	0.5	0.19
1904	47.4	13.1	12.3	7.8	10.9	2.2	3.3	3.3	0.5	0.18
1906	50.3	16.5	11.1	4.4	7.1	2.8	4.6	2.8	0.8	0.18
2766	46.5	12.7	12.6	7.3	11.8	1.8	3.1	3.6	0.4	0.19
2768	48.0	16.2	12.3	4.0	7.9	2.9	4.6	3.5	0.8	0.18
2772	47.5	14.4	12.4	7.2	9.7	2.3	3.3	3.1	0.6	0.18
2774	47.3	14.7	12.1	6.1	9.4	2.3	3.6	3.1	0.6	0.18
2777	47.5	15.4	12.7	5.5	9.2	2.4	3.8	3.3	0.7	0.18
2781	45.5	11.0	12.4	11.4	11.6	1.9	3.0	2.9	0.4	0.18
2782	44.6	14.1	12.0	6.4	10.6	3.9	4.3	3.1	0.9	0.20
2792	44.0	12.0	12.4	11.2	10.1	3.1	3.6	3.1	0.7	0.20
2794	50.8	15.6	10.7	4.8	7.6	3.0	4.6	2.7	0.8	0.16
2799	53.0	17.5	9.2	2.5	6.0	3.7	5.4	1.9	0.7	0.19
2805	48.5	12.8	11.0	10.0	10.0	2.1	2.8	2.4	0.4	0.17
2816	44.4	13.7	13.1	7.1	9.5	2.4	3.5	3.5	0.6	0.19
2820	44.2	12.3	12.1	10.8	9.9	3.1	3.7	3.0	0.7	0.20
2824	47.1	14.5	12.6	7.0	9.5	2.9	3.6	3.0	0.7	0.20
2829	50.3	16.6	11.2	4.3	7.2	2.7	4.5	2.8	0.8	0.18
2849	48.0	13.1	11.7	7.9	10.0	2.5	3.6	2.9	0.6	0.17
2856	44.3	14.2	12.1	5.5	7.3	2.5	3.8	3.1	0.5	0.17
2871	53.6	17.9	8.7	2.5	5.7	3.6	5.2	2.1	0.9	0.16
2888	60.0	15.4	7.2	2.6	4.6	3.0	4.9	1.7	0.5	0.10
2889	55.9	15.5	9.0	3.4	5.6	3.0	4.7	2.4	0.6	0.13
2894	60.8	15.6	6.6	2.1	3.8	3.1	5.1	1.6	0.5	0.09
2895	53.4	16.9	9.8	2.8	5.2	3.5	5.1	2.5	0.7	0.14
2896	45.9	11.8	12.5	10.8	11.2	2.0	2.6	2.8	0.3	0.18
2897	46.3	12.3	12.3	9.3	11.8	2.2	2.9	2.9	0.5	0.18
2905	42.8	14.2	11.5	5.3	9.8	3.6	5.7	3.6	0.7	0.19
2909	43.3	14.9	11.3	4.9	9.4	3.7	6.1	3.5	0.8	0.18
2911	47.7	16.8	9.4	2.8	7.0	5.4	6.6	2.8	0.8	0.17
2912	49.2	17.1	8.9	2.5	6.4	5.3	6.9	2.6	0.9	0.17
2915	48.0	15.5	12.7	5.1	8.4	2.8	3.6	3.2	0.6	0.18
3022	37.0	13.9	12.1	4.7	15.2	6.3	5.6	2.3	1.7	0.29
3023	38.7	15.2	14.7	4.2	12.3	4.8	5.0	3.2	2.1	0.28
3033	42.9	12.0	14.5	10.1	13.7	1.7	1.7	3.9	0.0	0.20
4414	34.2	5.8	12.1	12.9	16.3	1.1	4.5	4.2	0.9	0.21
6066	35.5	6.5	10.0	15.9	9.5	0.7	6.8	3.3	0.4	0.15
9958	39.5	18.5	11.0	3.3	8.9	6.3	5.7	2.2	1.3	0.24

Table 8.1 continued

C	Sr	Rb	Ba	Cr	Zr	Y	Nb	Zn	Ni	$\frac{K_2O}{Na_2O}$	K/Rb	Mg'
999	1379	101	934	314	314	17	154	80	68	2.1	323	0.59
1904	885	100	933	281	277	28	91	86	27	1.5	275	0.57
1906	1165	125	916	93	371	31	120	96	8	1.7	304	0.45
2766	857	86	773	161	264	30	88	89	1	1.7	302	0.55
2768	1476	136	1186	68	385	31	154	108	4	1.6	280	0.41
2772	925	132	827	192	294	29	90	97	8	1.4	204	0.55
2774	940	130	867	138	296	29	93	89	0	1.6	230	0.52
2777	1019	110	912	47	320	31	102	106	0	1.6	287	0.48
2781	891	89	767	652	238	28	82	77	158	1.6	284	0.66
2782	1548	119	1032	156	400	35	169	95	35	1.1	302	0.53
2792	1258	101	889	684	331	30	138	99	216	1.2	316	0.66
2794	1049	152	998	128	360	28	117	86	25	1.6	250	0.50
2799	1372	164	819	36	480	35	161	102	0	1.5	275	0.36
2805	761	89	709	711	239	37	75	89	81	1.4	263	0.66
2816	992	91	790	198	297	32	102	96	36	1.5	315	0.53
2820	1255	96	927	660	336	34	143	92	209	1.2	322	0.66
2824	1030	106	733	268	316	31	109	59	47	1.2	279	0.54
2829	1171	125	891	97	364	36	116	106	8	1.7	300	0.45
2849	997	118	808	369	298	28	102	89	69	1.4	255	0.59
2856	922	120	893	179	297	30	97	52	40	1.5	264	0.47
2871	1316	158	979	13	424	32	143	72	0	1.4	275	0.47
2888	642	189	872	44	330	25	89	17	1	1.6	215	0.44
2889	790	172	925	42	342	27	102	95	0	1.6	228	0.44
2894	602	200	893	30	331	24	86	56	0	1.6	212	0.44
2895	1008	150	1146	46	378	24	121	96	0	1.5	283	0.38
2896	709	79	622	615	222	31	69	79	147	1.3	275	0.65
2897	883	90	762	375	246	29	81	86	99	1.4	270	0.62
2905	1406	176	1398	67	333	24	183	89	31	1.6	269	0.49
2909	1470	144	1405	45	344	22	194	87	17	1.7	352	0.48
2911	1750	153	1340	20	412	29	232	82	11	1.2	354	0.39
2912	1696	167	1315	16	402	29	229	101	4	1.3	344	0.38
2915	994	114	942	63	309	31	103	100	0	1.3	261	0.46
3022	3401	132	1626	21	338	43	314	102	9	0.9	352	0.46
3023	2318	112	1423	26	371	38	188	89	17	1.0	366	0.38
3033	625	43	332	204	200	29	46	77	96	1.0	332	0.60
4414	2831	116	1602	425	377	20	254	88	196	4.3	322	0.69
6066	1109	183	552	822	186	9	145	67	470	9.4	310	0.77
995	2730	160	2182	22	422	46	270	124	10	0.9	294	0.39

* $Mg' = Mg / (Mg + Fe^{2+})$ (moles) where Fe^{2+} has been calculated from total Fe using the ratio of Fe_2O_3 (%) / $FeO = 0.15$ (Brooks et al., 1976).

- 2) extreme enrichment in alkalis, with K_2O usually in excess of Na_2O and reaching 7% of the total.
- 3) high TiO_2 contents that range from 1.6 to 4.8%.
- 4) high and variable contents of Ba (332 to 2182 ppm), CaO (3.8 to 16.3%), MgO (up to 15.9%), Nb (46 to 314 ppm), Rb (43 to 200 ppm), Sr (625 to 3401 ppm), Y (9 to 46 ppm), and Zr (186 to 480 ppm).

Compositional extremes are represented by lavas from Sabinyo, (the oldest and most eastern) and Niragongo (the youngest and most western) of the Birunga volcanoes. The Sabinyo latites and shoshonites are the only Si-saturated to oversaturated rocks of the Birunga field, and contain up to 61% SiO_2 . On the other hand the nephelinites and nepheline leucitites of Niragongo contain about 38% SiO_2 . The Sabinyo volcanics also have much less CaO, total Fe, K_2O , MgO, Na_2O , P_2O_5 , and TiO_2 than those of Niragongo (Table 8.1).

Table 8.1 also indicates that the nephelinites, nepheline leucitites and mikenites are characterized by a high Na_2O content, with values almost equal to or even exceeding their K_2O abundances. Of all the Birunga rock types, the Niragongo nephelinite and nepheline leucitites have the lowest K_2O/Na_2O ratios (0.90).

Table 8.2 shows averages of some element abundances from the eastern (Mgahinga, Muhavura and Sabinyo), central (Mikeno), and western (Niragongo) groups of volcanoes based on data given in Table 8.1. The analysis of the ejected block, sample C3033 from Nyamulagira, is not included. Table 8.2 indicates that the SiO_2 and Cr abundances decrease from east to west, while the concentrations of Ba, CaO, total Fe, MnO, Na_2O , Nb, P_2O_5 , Sr, and Zn increase from E to W.

Another example of the chemical diversity of the Birunga

Table 8.2: The Average of Some Elements for the Three Groups of Volcanoes of the Birunga Field.

	Eastern	Central	Western
SiO ₂ (wt%)	50.31	46.70	38.40
Fe ₂ O ₃ T	9.82	9.85	12.59
CaO	6.96	7.59	12.13
Na ₂ O	2.78	4.76	5.81
MnO	0.16	0.17	0.27
P ₂ O ₅	0.56	0.84	1.69
Sr (ppm)	924	1639	2816
Ba	968	1353	1744
Cr	120	27	23
Nb	129	218	257
Zn	86	90	105

Table 8.3: The Rare Earth Element and Th Analyses.

Sample	Ce	Eu	La	Lu	Nd	Sc	Sm	Tb	Th	Yb	Sum REE
C 999	265	3.2	132	0.3	98	31.4	13.5	1.2	15	1.4	546
*C1904	141	2.4	65	0.3	--	29.0	9.4	1.0	9	2.1	250
C2768	273	3.5	145	0.4	89	16.1	13.6	1.2	21	3.9	546
*C2794	183	2.6	91	0.3	--	17.0	11.1	1.1	17	2.4	309
*C2799	262	3.0	139	0.4	--	8.5	12.8	1.4	24	2.8	430
C2894	241	2.1	128	0.3	83	11.5	12.5	1.2	34	2.9	483
*C2895	206	2.5	103	0.3	--	11.0	11.2	1.0	24	2.2	337
*C2896	125	2.4	57	0.3	--	33.0	8.8	0.9	11	1.9	229
C2897	161	2.8	88	0.4	53	34.4	10.6	0.9	12	2.7	354
*C2905	217	3.1	110	0.3	--	18.0	11.7	1.0	18	1.9	363
*C2909	201	2.8	101	0.2	--	17.0	9.7	1.1	17	1.8	335
C3022	442	5.1	246	0.4	139	4.1	19.2	1.6	23	4.8	862
C3033	86	2.1	40	0.5	40	43.4	7.6	1.0	2	2.3	223
*C4414	486	5.7	236	0.1	--	26.0	23.1	2.1	27	0.7	780
*C6066	285	3.3	134	0.1	--	24.0	13.1	1.2	18	0.9	462

RhY-1

X	40	0.9	19	0.8	19	8.9	5.0	1.1	2	5.4	-----
R.V.	41	1.0	18	0.9	21	---	5.5	1.1	--	5.6	-----

Chond.	.957	.087	.367	.038	.711	7.8	.231	.058	.051	.248	-----
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RhY-1 = Internal Standard (Rhyolite).

X = Values obtained during this study.

R.V. = Recommended Value (University of Ottawa).

Chond. = Chondritic values.

All the values are in ppm. The chondritic values are from Taylor (1980). The samples with (*) are from Mitchell and Bell (1976). The rest are from this study.

volcanics is shown by the lavas of the two active volcanoes, Niragongo and Nyamulagira, only 15 km apart. Despite being geographically close and having high Na_2O contents, these volcanoes exhibit pronounced differences in chemistry. The Niragongo rocks show lower Si, higher Al, Fe, K, Na, and P than those of Nyamulagira (Sahama, 1972).

The volcanic rocks of the Toro-Ankole field are characterized by undersaturated, kalsilite-bearing, potassic rocks (Holmes, 1940; King and Sutherland, 1960). Sample C4414, which has the lowest SiO_2 content (34.2%), is a katungite from the Katunga volcano, and like the flows of the Birunga field, contains high concentrations of Ba, CaO, MgO, Nb, Rb, Sr, TiO_2 , Y and Zr relative to chondrites. The volcanics of the Bunyaraguru have higher concentrations of Cr and Ni than the rocks of the Birunga field.

The average K/Rb ratio of the samples from the Birunga and Toro-Ankole fields is very close to the estimates of about 290 for average continental crust (Taylor et al., 1956; Horstman, 1957; Gast, 1968; Taylor, 1960; Heier and Adams, 1964; Bell and Doyle, 1971; Nyambok, 1977). The new analyses reveal an average of 318 for the Toro-Ankole and 285 for the Birunga fields (Table 8.1). The weighted average for all samples is 288.

8.2.1 RARE EARTH ELEMENT CHEMISTRY

REE data were obtained for five samples (samples C2768, C2894, C2897, C3022 and C3033) by neutron activation analysis by Bondar-Clegg Company. The precision of the data, expressed as one standard deviation, is better than 4% for all the elements determined except Eu, Lu, Nd, and Sm. The accuracy has been estimated by comparing the data

obtained on the standard sample RHY-1 with values reported by Abbe (1983). In addition Nd and Sm concentrations were determined by isotope dilution analysis for four samples (C2889, C2894, C3022 and C3023) (see Appendix 9.1 for method). The results from both methods are in excellent agreement.

Mitchell and Bell (1976) have also presented REE data for the rocks of the Birunga and Toro-Ankole fields. The REE data from all rock types in the Birunga and Toro-Ankole fields are represented in Table 8.3 which includes analyses both from Mitchell and Bell (1976) and this study. The chondrite-normalized plot of the five samples analysed in this study (chondrite values taken from Taylor, 1980) is shown in Fig. 8.1.

In spite of the analytical uncertainties associated with the REE data it can be seen that:

- 1) All the analysed samples are enriched in REE and highly enriched in the light REE (LREE) relative to chondrites.
- 2) All the samples display sub-parallel REE patterns.
- 3) There is no negative Eu anomaly, except possibly for the latite (C2894).
- 4) The nephelinite from Niragongo, C3022, is more enriched in total REE (862 ppm) than the latite of Sabinyo, C2894 (483 ppm).

8.3 INCOMPATIBLE ELEMENT PATTERNS

The elements that partition strongly into the residual liquids in mafic and ultramafic magmas are useful in investigating the relative roles of crust and mantle in magma genesis (Bailey, 1982; Thompson et al., 1984). One way of comparing one rock to another is to normalize

Fig. 8.1: The Chondrite normalized plot of the Rare Earth Elements.

the elements to a reference composition that can be represented by chondritic abundances or hypothetical primordial mantle composition (Peccerillo et al., 1984). These patterns or "spidergrams" can be used to compare trace element abundances of different rock types and different tectonic settings (Thompson et al., 1983). Because of the large amount of analytical information that was obtained for the Birunga and Toro-Ankole samples, it was felt that the spidergrams were the simplest way of handling the data. The data from the Birunga samples are normalized to the primitive mantle abundances of Taylor (1980) and are plotted in the order of decreasing incompatibility for mantle mineralogy of a garnet lherzolite (Wood, 1981).

Very similar overall patterns are observed for all samples. The patterns indicate enrichment in Ba, K, Nb, P, Rb, Sr, Ti and Zr relative to primitive mantle. The degree of enrichment of the incompatible elements in the rocks of the Birunga field relative to the primitive mantle is variable. In general the enrichment decreases with decreasing incompatibility of the elements. Fig. 8.2 shows "spidergrams" for individual rock types of the Birunga field. An average metasomatic mantle composition, based on data from nodules of metasomatized mantle lherzolites, (Frey and Green, 1974; Carter et al., 1978a; Frey and Prinz, 1978; Frey et al., 1978; Hawkesworth and Vollmer, 1979; Sun et al., 1979; Wood, 1979; Boettcher and O'Neill, 1980; Menzies and Murthy, 1980a, b, and c; Wass and Rogers, 1980) has a very similar shape to the patterns based on data from the Birunga and Toro-Ankole volcanics (Fig. 8.3).

Fig. 8.4 is a plot that shows the average values for the mikenites and latites normalized to an average nephelinite. From this it can be seen that the nephelinites are enriched in Ba, Nb, P, Sr, and

ABSAROKITES.

□ 1906

+ 2772

○ 2777

Δ 2794

x 2824

▽ 2829

BANAKITES

□ 2799

+ 2871

KIVITES

LEUCITITES

MIKENITES

LEUCITE BASANITES

MURAMBITES

□ 2774 + 2781 ○ 2805 △ 2849 x 2896

NEPHELINITES

□ 3022 + 3023 ○ 9958

Fig. 8.2: Spidergrams for different rock types of the Birunga field. Normalizing values are taken from Taylor (1980).

Fig. 8.3: Spidergram for different rock types of the Toro-Ankole-field and an average metasomatized mantle nodule. The values for the metasomatized lherzolite nodule are taken from Frey and Green (1974), Menzies and Murthy (1980c).

Fig. 8.4: The nephelinite normalized plot of incompatible elements for the banakite, latite, mikenite and shoshonite using the average values for each rock type.

Y relative to the other rock types. Average values for the remaining samples investigated in this study fall between the limits set by the latites and nephelinites.

Niobium is an element commonly enriched in alkalic granites, metasomatically altered granites, pegmatites, carbonatites and hydrothermal veins in alkalic complexes (Clague and Frey, 1982) and is easily carried in H_2O - and CO_2 -rich fluids (Heinrich, 1974). Nephelinites have the highest Nb concentration of any Birunga samples. A plot of Nb vs Na_2O (Fig. 8.5) indicates an interesting positive correlation (correlation coefficient = 0.95) for all the rocks of the Birunga field. For comparison the average Nb content of the lower crust is 28 ppm, for average crust 14 ppm (Thompson et al., 1984), for MORB 4 ppm (Hofmann et al., 1986), and for Niragongo 257 ppm. There is also a strong positive correlation between Ba and Na_2O for all the Birunga samples (correlation coefficient = 0.88).

8.4 PRIMARY AND DERIVATIVE MAGMAS IN THE BIRUNGA AND TORO-ANKOLE FIELDS

The choice of primary (or primitive) mantle magmas for the Birunga and Toro-Ankole rocks is difficult to establish. From the known phase relationships in a diopside - forsterite - nepheline - silica and diopside - nepheline - silica system, Sood and Edgar (1972) suggest two parental magmas for the alkaline rocks of the east African rift system: 1) a nephelinitic magma for the sodic volcanics ($Na > K$); and 2) a magma similar in composition to the olivine leucitites.

The $Mg / (Mg + Fe^{2+})$ atomic ratio known as the Mg number (Mg') is frequently used to identify the more primitive magma type in

Fig. 8.5: The plot of Na_2O (wt%) vs Nb (ppm) for all the rocks of the Birunga field.

continental alkaline suites (Sahama, 1972; LeBas, 1977; Edgar et al., 1980; Civetta et al., 1981; Peccerillo et al., 1984; Nelson et al., 1986). Based on experimental studies, Edgar et al. (1975) state that the Mg' of around 0.75 is appropriate for a primary liquid derived from a peridotite or a pyrolite source. Studies of lherzolite xenoliths in kimberlites indicate that magmas derived from a peridotite upper mantle composition (Mg' of 0.88-0.89) must have Mg' of around 0.68-0.75 for up to 30 percent melting (Roeder and Emslie, 1970).

Magnesium numbers have been calculated for all samples listed in Table 8.1. Because Fe^{++} was not measured directly, the ratio $Fe_2O_3/FeO = 0.15$ (Brooks et al., 1976) has been used to calculate the Mg' . Although this is only an approximation, it is interesting to note that most samples have Mg' that fall between 0.36 and 0.66. Only the mafurite sample from the Toro-Ankole field (C6066) has a Mg' (0.77) similar to those from primary melts. The Niragongo nephelinites and nepheline leucitites have very low Mg' of 0.41. Sahama (1972) has shown that in order to produce the nephelinite with such a low Mg' and Mg content, at least 20% olivine fractionation is needed.

It should, however, be pointed out that Vollmer (in press) argues that primary magmas can be characterized by low concentration of Cr and Ni and a low Mg' if the partition coefficient of these elements in the magma is low and/or if the olivine is consumed by metasomatic reaction in the source region of this magma type. This argument is based on the presence of an unusual (metasomatised) mantle composition in the area of rifting.

CHAPTER 9

ISOTOPIC SYSTEMATICS

SR ISOTOPE SYSTEMATICS

9.1 INTRODUCTION

Four naturally occurring isotopes of Sr, namely ^{84}Sr , ^{86}Sr , ^{87}Sr , and ^{88}Sr exist. Of these, only ^{87}Sr is radiogenic, formed by beta-decay of the parent, ^{87}Rb , which has a decay constant of $\lambda = 1.42 \times 10^{-11} \text{ yr}^{-1}$. The $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic ratio in any system is a function of the time-integrated $^{87}\text{Rb}/^{86}\text{Sr}$ ratio of the system (Faure, 1986).

Processes other than radioactive decay, for example isotope mass fractionation in the mass spectrometer, can affect the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio. In practice these mass dependant effects are removed by normalizing the measured $^{87}\text{Sr}/^{86}\text{Sr}$ ratio to a value of $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$. Igneous rocks from the same magma should all have the same $^{87}\text{Sr}/^{86}\text{Sr}$ ratio, if the system remains chemically closed during crystallization. If different $^{87}\text{Sr}/^{86}\text{Sr}$ initial values are recorded from the same lava flows, this suggests contamination prior to or during magma ascent to the surface.

9.2 Sr ISOTOPIC ANALYSIS

Sr was extracted from sample aliquots by conventional cation exchange techniques. Isotope ratios of 43 whole rock samples from the Birunga and Toro-Ankole fields were measured using a Finnigan-MAT 261 solid source, mass spectrometer. Appendix 9.1 explains in detail the

chemical procedure and instrumental techniques.

Rubidium and Sr concentrations were determined by X-ray fluorescence technique. Rubidium and Sr abundances for four samples (C2889, C2894, C3022 and C3023) were determined by isotope dilution (see Appendix 9.1).

The uncertainty of the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios is based on replicate measurements. Sample (C2782) was prepared three times. The data (0.70663, 0.70663 and 0.70664) demonstrate the excellent overall reproducibility. The reproducibility is considered to be 3 parts in 70000 at the two sigma level. Accuracy of the measurements was monitored by analyzing the standard sample NBS-987 which yielded an average $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.71023 ± 0.00003 .

9.3 RESULTS

Table 9.1 gives the values of the initial $^{87}\text{Sr}/^{86}\text{Sr}$ for the samples of the Birunga and Toro-Ankole fields. The $^{87}\text{Sr}/^{86}\text{Sr}$ ratios range from 0.70447 to 0.70971 in the Birunga field and cover a range of values similar to those obtained by other workers (Bell and Powell, 1969; Vollmer and Norry, 1983a;b). The two extreme values of 0.70447 and 0.70971 belong to a nepheline leucitite from Niragongo and a latite from Sabinyo, respectively. Fig. 9.1a indicates the frequency distribution of Sr isotopic compositions of Birunga volcanics; the dashed lines indicate the samples from the Toro-Ankole field.

The Niragongo nephelinites and nepheline leucitites are characterized by low initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.70447 to 0.70477), high Sr concentrations (2199 to 3333 ppm) and low Rb/Sr ratios

Table 9.1

Sr Isotopic Composition of the Birunga and Toro-Ankole Volcanics

C	RT	V	$^{87}\text{Sr}/^{86}\text{Sr}$	Sr	Rb	Rb/Sr	$^{87}\text{Rb}/^{86}\text{Sr}$
999	13	20	0.70520	1379	101	0.073	0.211
1904	4	3	0.70686	885	100	0.113	0.327
1906	3	15	0.70777	1165	125	0.107	0.310
2766	7	3	0.70710	857	86	0.100	0.289
2768	7	3	0.70716	1476	136	0.092	0.266
2772	3	18	0.70741	925	132	0.143	0.414
2774	6	12	0.70747	940	130	0.138	0.399
2777	3	12	0.70732	1019	110	0.108	0.312
2781	6	4	0.70625	891	89	0.099	0.289
2782	10	13	0.70663	1548	119	0.077	0.223
2792	10	13	0.70651	1258	95	0.076	0.220
2794	3	3	0.70732	1049	152	0.145	0.419
2799	9	3	0.70714	1372	164	0.120	0.347
2805	6	17	0.70787	761	89	0.117	0.338
2816	4	6	0.70676	992	91	0.092	0.266
2820	10	13	0.70656	1255	96	0.077	0.223
2824	3	16	0.70628	1030	106	0.103	0.298
2829	3	3	0.70777	1171	125	0.107	0.310
2849	6	3	0.70680	997	118	0.118	0.341
2856	4	14	0.70726	922	120	0.130	0.376
2871	9	7	0.70742	1316	158	0.120	0.347
2888	1	1	0.70946	642	189	0.294	0.851
2889	8	1	0.70863	790	172	0.218	0.631
*2894	1	1	0.70971	623	241	0.387	1.164
2895	8	1	0.70776	1008	150	0.149	0.431
2896	6	9	0.70654	709	79	0.111	0.321
2897	10	9	0.70595	883	90	0.102	0.295
2905	11	11	0.70553	1406	176	0.125	0.362
2909	2	2	0.70543	1470	144	0.098	0.283
2911	2	2	0.70573	1750	153	0.087	0.252
2912	2	2	0.70584	1696	167	0.099	0.286
2915	4	10	0.70699	994	114	0.115	0.333
3021	5	22	0.70463	2866	136	0.048	0.139
*3022	5	5	0.70469	3333	108	0.032	0.094
*3023	11	5	0.70447	2199	110	0.050	0.145
3032	4	8	0.70566	----	----	-----	-----
3033	12	8	0.70535	625	43	0.069	0.200
4414	14	19	0.70482	2831	116	0.041	0.119
5628	10	24	0.70481	3074	68	0.022	0.064
6066	15	21	0.70506	1109	183	0.165	0.477
6095	12	21	0.70521	1932	172	0.089	0.257
9872	5	23	0.70450	2421	110	0.045	0.130
9958	11	5	0.70477	2730	160	0.059	0.171

* The Rb and Sr abundances for these samples were determined by isotope dilution analysis, the remainder by XRF.

Fig. 9.1: Histograms of Sr (a) and Nd (b) isotopic ratios for Birunga field (solid lines) and Toro-Ankole field (dashed lines).

(average = 0.047, Table 9.1). Sabinyo shoshonites and latites, on the other hand, are characterized by high and variable $^{87}\text{Sr}/^{86}\text{Sr}$ initial ratios (0.70776 to 0.70971), much lower Sr concentrations (623 to 1008 ppm), and higher Rb/Sr values (average = 0.262) than Niragongo (Table 9.1).

The variation in Sr isotopic composition, even within rocks of the same volcano such as Sabinyo, indicates simple magmatic differentiation from a single parent magma cannot be the only factor involved in the petrogenesis of these rocks. The high initial $^{87}\text{Sr}/^{86}\text{Sr}$ values from the Sabinyo volcano may be due to: (i) assimilation of the continental crust (Bell and Powell, 1969); (ii) inheritance from an unusual source region (Thompson, 1985); or (iii) radiogenic enrichment following metasomatism (Vollmer and Norry, 1983a).

The plots of $^{87}\text{Sr}/^{86}\text{Sr}$ vs $^{87}\text{Rb}/^{86}\text{Sr}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ vs Sr content (Figs. 9.2 and 9.3) for all the Birunga volcanics indicate excellent correlations (correlation coefficients of 0.83 and -0.67 respectively). The positive correlation between $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{87}\text{Rb}/^{86}\text{Sr}$ ratios can be interpreted either as (i) the mixing of at least two end members with distinct isotopic signatures and Rb/Sr ratios or as (ii) a time dependent relationship between Rb and Sr within the source regions. If two end-members with different $^{87}\text{Sr}/^{86}\text{Sr}$ and Rb/Sr ratios are mixed in varying proportions, the mixtures have isotope compositions that lie along a straight line on a plot of $^{87}\text{Sr}/^{86}\text{Sr}$ vs Rb/Sr (Riley and Compston, 1962; Lamphere et al., 1964). Furthermore, Bell and Powell (1969) and Barberi et al. (1980) point out that plots of initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios vs other elements or oxides for such mixtures would produce hyperbolic curves. Thus the positive correlation between $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{87}\text{Rb}/^{86}\text{Sr}$ shown in Fig. 9.2 could reflect a simple,

Fig. 9.2: $^{87}\text{Sr}/^{86}\text{Sr}$ vs $^{87}\text{Rb}/^{86}\text{Sr}$ for the Birunga volcanics.
 Symbols the same as in Fig. 8.5.

Fig. 9.3: $^{87}\text{Sr}/^{86}\text{Sr}$ vs Sr for Birunga samples. Symbols as in Fig. 8.5.

binary mixing process. On the other hand, Vollmer and Norry (1983b) state that this straight line of positive slope may reflect the preservation of an isochron, corresponding to an age of around 500 Ma that corresponds to the age of an event that affected the source region of the Birunga volcanics. The slope of the best-fit line for all of the data from the Birunga samples (except Sabinyo shoshonites and latites) give an age of 484 Ma. This is in excellent agreement with the secondary Pb isochron obtained by Vollmer and Norry (1983a) from the Niragongo lavas. Fig. 9.2 indicates that the data from Sabinyo shoshonites and latites fall on a straight line with a shallower slope than the rest of the Birunga samples. This shallow slope corresponds to an age of 187 Ma.

If the linear relationship between $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{87}\text{Rb}/^{87}\text{Sr}$ ratios indicates that the observed volcanics are produced by mixing, then the slope of the straight lines would have no real age significance (Bell and Powell, 1969). If this were so, then one end-member has a low $^{87}\text{Sr}/^{86}\text{Sr}$ initial ratio, high Sr content and low Rb/Sr ratio, and the other end member shows a high $^{87}\text{Sr}/^{86}\text{Sr}$ ratio, low Sr content and high Rb/Sr ratio.

The graph of $^{87}\text{Sr}/^{86}\text{Sr}$ vs SiO_2 for all the Birunga samples (Fig. 9.4) indicates a very strong positive correlation (correlation coefficient = 0.89). In Fig. 9.3 the samples from Sabinyo show a marked correlation between $^{87}\text{Sr}/^{86}\text{Sr}$ initial ratio and Sr content. This contrasts with data from Mikenno and Niragongo.

The $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic ratios of Sabinyo and Mikenno, the two oldest volcanoes of the Birunga field, are quite different from each other. The Sabinyo volcanics have higher $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (average

Fig. 9.4: $^{87}\text{Sr}/^{86}\text{Sr}$ vs SiO_2 for the Birunga rocks.

0.70889) than the mikenites of Mikenö (average 0.70567). Differences also exist in the major and trace elements. The Sabinyo samples have higher SiO_2 , lower Ba, CaO, K_2O , Na_2O , Nb, Ni, P_2O_5 , Sr, and Zr, than the Mikenö samples (see Table 8.1).

Nd ISOTOPIC SYSTEMATICS

9.4 Nd ISOTOPIC COMPOSITION

Both samarium (Sm) and neodymium (Nd) are rare earth elements. One of the Sm isotopes (^{147}Sm) is radioactive and decays by alpha emission to a stable isotope of neodymium (^{143}Nd) with a half life of 1.06×10^{11} yr and a decay constant of $6.54 \times 10^{-12} \text{ yr}^{-1}$ (Carter et al., 1978b; Faure, 1986). The $^{143}\text{Nd}/^{144}\text{Nd}$ ratio is a function of the time integrated $^{147}\text{Sm}/^{144}\text{Nd}$ ratio (Faure, 1986).

Samples from both the Birunga and Toro-Ankole fields were prepared for Nd isotopic analysis using a reverse-phase chromatography technique (Richard et al., 1976). The Nd isotopic abundances were measured on a Finnigan-MAT 261 multicollector mass spectrometer. Appendix 9.1 gives the details of sample preparation and the mass spectrometer techniques. The Nd and Sm concentrations of samples C2889, C2894, C3022, and C3023 were determined using isotope dilution (Table 9.2, also see Appendix 9.1).

The reproducibility of $^{143}\text{Nd}/^{144}\text{Nd}$ measurements is shown by the duplicate values given in Table 9.3 for samples C2889, C2894, and C9958. The reproducibility is considered to be ± 0.00003 at the 2 sigma level.

Table 9.2

The Nd and Sm concentrations of four samples
using the isotope dilution method

Sample C	Nd (ppm)	Sm (ppm)	Sm/Nd weight	$^{147}\text{Sm}/^{144}\text{Nd}$ atomic
2889	145	22	0.1517	0.09342
2894	80	12	0.1500	0.09342
3022	138	20	0.1449	0.08517
3023	117	17	0.1453	0.08868

Table 9.3

Reproducibility of Nd isotopic ratios

Sample C	$^{143}\text{Nd}/^{144}\text{Nd}$		average
	first	second	
2889	0.51211	0.51215	0.51213
2894	0.51203	0.51205	0.51204
9958	0.51265	0.51267	0.51266

9.5 RESULTS

The initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratios for 33 samples from both the Birunga and Toro-Ankole volcanic fields are given in Table 9.4. The measured initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratios are normalized to $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$ to correct for mass-dependant isotope fractionation in the mass spectrometer. The average $^{143}\text{Nd}/^{144}\text{Nd}$ value for the standard sample, BCR-1, is 0.51263 ± 0.00002 . The range in $^{143}\text{Nd}/^{144}\text{Nd}$ ratios for the Birunga samples is 0.51202 to 0.51269. Fig. 9.1b indicates the frequency distribution of Nd isotopic compositions for all the analysed samples. The dashed lines indicate samples from the Toro-Ankole field. Sabinyo shoshonites and latites show the lowest initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratios (0.51202 to 0.51220) for the Birunga field, while the ratios for the Niragongo nephelinites are the highest. The initial $^{143}\text{Nd}/^{144}\text{Nd}$ of the three Niragongo samples are the same within experimental uncertainty (0.51266 - 0.51269).

Total Nd and Sm were obtained by isotope dilution for four samples. These results are given in Table 9.2. The $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{147}\text{Sm}/^{144}\text{Nd}$ ratios for these samples are plotted in Fig. 9.5. The negative slope indicated in this figure shows that the Sm/Nd system behaved quite differently from the Rb/Sr and U/Pb systems.

Sr AND Nd ISOTOPIC SYSTEMATICS

9.6 ANTI-CORRELATION PLOT (ACP)

In a study of mid-ocean ridge basalts (MORBs) and some oceanic island basalts (OIBs), DePaolo and Wasserburg (1976a;b) and O'Nions et al. (1977) noticed a strong negative correlation between present day

Table 9.4

The Isotopic Analyses of the Birunga and Toro-Ankole Fields

C	R.T.	V.	$^{87}\text{Sr}/^{86}\text{Sr}$	$^{143}\text{Nd}/^{144}\text{Nd}$	ϵ_{Sr}	ϵ_{Nd}
999	13	20	0.70520	0.51250	2.84	-2.73
1904	4	3	0.70686	0.51238	26.38	-5.07
1906	3	15	0.70777	---	39.29	---
2766	7	3	0.70710	---	29.79	---
2768	7	3	0.70716	0.51238	30.64	-5.07
2772	3	18	0.70741	---	34.18	---
2774	6	12	0.70747	0.51233	35.04	-6.05
2777	3	12	0.70732	---	32.91	---
2781	6	4	0.70625	---	17.73	---
2782	10	13	0.70663	---	23.12	---
2792	10	13	0.70651	0.51245	21.42	-3.71
2794	3	3	0.70732	0.51228	32.91	-7.02
2799	9	3	0.70714	---	30.36	---
2805	6	17	0.70787	0.51228	40.71	-7.02
2816	4	6	0.70676	0.51236	24.97	-5.46
2820	10	13	0.70656	---	22.13	---
2824	3	16	0.70628	0.51240	18.16	-4.68
2829	3	3	0.70777	0.51233	39.29	-6.05
2849	6	3	0.70680	0.51237	25.53	-5.27
2856	4	14	0.70726	0.51234	32.06	-5.85
2871	9	7	0.70742	0.51236	34.33	-5.46
2888	1	1	0.70946	0.51202	63.26	-12.09
2889	8	1	0.70863	0.51213	51.49	-10.34
2894	1	1	0.70971	0.51204	66.81	-11.70
2895	8	1	0.70776	0.51220	39.15	-8.58
2896	6	9	0.70654	0.51242	21.84	-4.29
2897	10	9	0.70595	0.51247	13.48	-3.32
2905	11	11	0.70553	0.51258	7.52	-1.17
2909	2	2	0.70543	0.51254	6.10	-1.95
2911	2	2	0.70573	0.51255	10.36	-1.76
2912	2	2	0.70584	0.51246	11.92	-3.51
2915	4	10	0.70699	---	28.23	---
3021	5	22	0.70463	0.51267	-5.25	0.59
3022	5	5	0.70469	0.51268	-4.40	0.78
3023	11	5	0.70447	0.51268	-7.52	0.78
3032	4	8	0.70566	0.51255	9.36	-1.76
3033	12	8	0.70535	0.51260	4.97	0.78
4414	14	19	0.70482	0.51250	-2.55	-2.73
5628	10	24	0.70481	---	-2.70	---
6066	15	21	0.70506	0.51249	0.85	-2.93
6095	12	21	0.70521	0.51247	2.98	-3.32
9872	5	23	0.70450	0.51269	-7.09	0.98
9958	11	5	0.70477	0.51266	-3.26	0.39

$^{143}\text{Nd}/^{144}\text{Nd}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic ratios. Because oceanic basalts originated within the mantle and are most probably unaffected by crustal contamination, they referred to the distribution of the data as the "mantle array".

Assuming that the Sm/Nd ratio of the Earth is equal to that of chondritic meteorites, an average value of 0.1967 (from the analysis of chondrites and the achondrite Juvinas; Jacobsen and Wasserburg, 1980), can be used to calculate the $^{143}\text{Nd}/^{144}\text{Nd}$ ratio of a chondritic reservoir (CHUR) at the present time using an age of 4.6 Ga for the Earth. In this model the assumption is made that Nd has evolved in a uniform reservoir with the Sm/Nd ratio equal to that of chondritic meteorites. The present-day $^{143}\text{Nd}/^{144}\text{Nd}$ ratio of this reservoir is 0.51264. The Rb/Sr ratio of the uniform reservoir (UR), also known as bulk Earth (BE), cannot be determined from chondritic meteorites because Rb is a volatile element. Instead the mantle array is used to obtain the present-day $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of BE. Estimates of the present-day Sr ratio of UR, therefore, range from 0.7045 to 0.7052 depending on which set of data is used for construction of the mantle arrays (DePaolo and Wasserburg, 1976a and b; O'Nions et al., 1977; Allègre et al., 1979; Zindler et al., 1982). Using present-day $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.705, the primordial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.69899 (basaltic achondrite best initial; BABI) and an age of 4.6 Ga for the Earth, a corresponding ratio of 0.089 can be calculated for $^{87}\text{Rb}/^{86}\text{Sr}$. The latter corresponds to a value of 0.03 for the Rb/Sr ratio of BE (DePaolo and Wasserburg, 1976a; O'Nions et al., 1977).

The values of 0.705 and 0.51264 are used for the Sr and Nd isotopic ratios of BE respectively to define the four quadrants on the ACP. The upper left-hand quadrant which contains the mantle array

encompasses the data from samples which have lower $^{87}\text{Sr}/^{86}\text{Sr}$ and higher $^{143}\text{Nd}/^{144}\text{Nd}$ than the BE, indicating an origin from sources that had lower Rb/Sr and higher Sm/Nd ratios than UR. The bottom right-hand quadrant includes the data from rocks with higher $^{87}\text{Sr}/^{86}\text{Sr}$ and lower $^{143}\text{Nd}/^{144}\text{Nd}$ values than BE, i.e. those rocks derived from sources that have higher Rb/Sr and lower Sm/Nd ratios than BE. Mantle-derived magmas, contaminated by continental crust (Faure, 1986), would generate data that would also lie within this quadrant. Rocks whose data plot in this quadrant originate from enriched sources. Data falling in the upper right-hand quadrant would indicate a coupled enrichment in Sm and Rb which is inconsistent with the geochemical properties of these two elements. Few fresh igneous rocks yield data that fall in either the upper right or lower left quadrants. The lower left-hand quadrant indicates lower Rb/Sr and Sm/Nd ratios than the uniform reservoir. Although data from basalts of the Tertiary volcanic province of Scotland fall in this quadrant, these basalts are considered to be the products of a magma contaminated with basement granulites characterized by low $^{143}\text{Nd}/^{144}\text{Nd}$ and low $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (Carter et al., 1978a).

The variation in Nd isotope ratios can also be conveniently measured in terms of the "epsilon parameter" which measures deviations of an isotopic value from a reference value (DePaolo and Wasserburg, 1976a and 1977). An $\epsilon(0)$ parameter can be defined as follows:

$$\epsilon_{\text{Nd}}(0) = \left(\frac{(^{143}\text{Nd}/^{144}\text{Nd})_{\text{initial}}}{(^{143}\text{Nd}/^{144}\text{Nd})_{\text{CHUR}}(0)} - 1 \right) \times 10^4$$

A similar parameter can be written for Sr:

$$\epsilon_{\text{Sr}}(0) = \left(\frac{(^{87}\text{Sr}/^{86}\text{Sr})_{\text{initial}}}{(^{87}\text{Sr}/^{86}\text{Sr})_{\text{CHUR}}(0)} - 1 \right) \times 10^4$$

$\epsilon_{\text{Nd}}(0)$ and $\epsilon_{\text{Sr}}(0)$ are the present-day Nd and Sr epsilon values. This convention is used for young rocks of essentially zero age.

The epsilon (ϵ) values for both Sr and Nd isotopic values are calculated and recorded in Table 9.4, using the values of 0.705 and 0.51264 for $(^{87}\text{Sr}/^{86}\text{Sr})_{\text{UR}}(0)$ and $(^{143}\text{Nd}/^{144}\text{Nd})_{\text{CHUR}}(0)$ respectively.

Fig. 9.6 is a plot of $^{87}\text{Sr}/^{86}\text{Sr}$ vs $^{143}\text{Nd}/^{144}\text{Nd}$ isotopic ratios for all Birunga samples. This data shows an extremely good negative correlation (correlation coefficient = 0.98) extending from the depleted quadrant close to BE well into the enriched quadrant. The extremes are represented by the data from the Niragongo nephelinites and nepheline leucitites that fall in the depleted quadrant ($\epsilon_{\text{Nd}} = 1.0$, $\epsilon_{\text{Sr}} = -7.1$) and the latites from Sabinyo with ϵ_{Nd} of -11.7 and ϵ_{Sr} of 66.8 in the enriched quadrant. The distribution of data is similar to that found by Vollmer and Norry (1983a;b) although their array was based on only 12 samples. Their array extends from $\epsilon_{\text{Nd}} = 2.5$ for Niragongo to $\epsilon_{\text{Nd}} = -10.0$ for Sabinyo samples.

The data from the Birunga rocks are compared with the data from similar volcanics e.g. the Pliocene-Recent alkalic rocks of central Italy in Fig. 9.7. This figure shows that most of the $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios of the Italian volcanics lie in the enriched quadrant although they show considerably greater variation than the Birunga rocks (Carter et al., 1978b; Hawkesworth and Vollmer, 1979).

Fig. 9.6; Anti-correlation plot (ACP). Symbols as in Fig. 8.5.

Fig. 9.7: Comparison of the isotopic ratios from Birunga and Italian volcanic provinces. The data from the Italian province is from Hawkesworth and Vollmer (1979) and Carter et al. (1978b).

9.7 Pb ISOTOPIC SYSTEMATICS

The analyses of Pb isotopic compositions of some of the Niragongo nephelinites and nepheline leucitites by Vollmer and Norry (1983a;b), given in Table 9.5, indicate that despite their fairly constant Nd and Sr isotopic compositions similar to BE, they are characterized by an unusually large variation in the $^{206}\text{Pb}/^{204}\text{Pb}$ ratios (19.101 to 60.230). On the other hand the Sabinyo Pb isotope compositions show only small variations ($^{206}\text{Pb}/^{204}\text{Pb} = 19.338 \pm 18$ to 19.435 ± 22) and are close to values for typical continental crust. The $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ ratios of the Sabinyo rocks also indicate minor variations (15.798 to 15.828 and 41.088 to 41.260 respectively).

Crustal contamination is rejected as an explanation for the extreme Pb isotopic variation of Niragongo samples by Vollmer and Norry (1983a) because one of the samples is more radiogenic than any Pb found in common crustal rocks. The identical Pb isotopic composition of all Sabinyo samples, also, argues strongly against crustal contamination (Vollmer and Norry, 1983b), since very large Pb isotopic variations are displayed by crustal rocks (Doe, 1970; Doe and Zartman, 1979).

The Pb isotopic data suggests that the mantle beneath Niragongo is not primitive; rather it has experienced LIL enrichment and depletion at different times (Vollmer and Norry, 1983b). A $^{207}\text{Pb}/^{204}\text{Pb} \pm ^{206}\text{Pb}/^{204}\text{Pb}$ isochron gives an age of 484 ± 28 Ma for the Niragongo samples and is interpreted by Vollmer and Norry (1983a) to be the age of a metasomatic event which inhomogeneously enriched the Niragongo upper mantle source in U and Th relative to Pb. This metasomatic event, if real, does not seem to have affected the Sm-Nd system.

Table 9.5

Pb isotopic composition for some of the samples of Birunga field. Sources: Vollmer and Norry (1983a;b).

C	$\frac{208\text{Pb}}{206\text{Pb}}$		$\frac{207\text{Pb}}{206\text{Pb}}$		$\frac{206\text{Pb}}{204\text{Pb}}$		$\frac{207\text{Pb}}{204\text{Pb}}$		$\frac{208\text{Pb}}{204\text{Pb}}$		Pb		U		$\frac{238\text{U}}{204\text{Pb}}$	
1904	2.0888	+ 4	0.81089	+ 12	19.420	+ 9	15.748		40.565		6.99		2.34		22.2	
2794	2.1060	+ 2	0.81357	+ 4	19.402	+ 16	15.785		40.861		12.02		3.83		21.2	
2799	2.1128	+ 9	0.81601	+ 20	19.334	+ 11	15.777		40.849		18.50		5.15		18.4	
2824	2.1062	+ 9	0.81406	+ 20	19.408	+ 24	15.799		40.877		---		---		---	
2888	2.1257	+ 4	0.81792	+ 12	19.360	+ 8	15.835		41.154		36.50		6.07		11.1	
2889	2.1263	+ 13	0.81653	+ 26	19.405	+ 18	15.845		41.260		---		---		---	
2894	2.1247	+ 5	0.81849	+ 11	19.338	+ 18	15.828		41.088		---		---		---	
2895	2.1189	+ 2	0.81291	+ 7	19.435	+ 22	15.798		41.180		---		---		---	
2897	2.0828	+ 8	0.81101	+ 18	19.409	+ 30	15.741		40.426		---		---		---	
2909	2.0763	+ 5	0.81062	+ 9	19.396	+ 12	15.723		40.272		4.53		4.18		60.4	
2911	2.0810	+ 2	0.81282	+ 6	19.409	+ 11	15.776		40.390		---		---		---	
3021	2.0501	+ 2	0.81416	+ 5	19.192	+ 12	15.625		39.346		---		---		---	
3022	2.0382	+ 5	0.81326	+ 16	19.218	+ 22	15.629		39.171		11.30		13.99		80.0	
3032	2.0940	+ 4	0.82228	+ 9	19.101	+ 27	15.706		39.997		---		---		---	
9872	1.6983	+ 7	0.29783	+ 9	60.230	+ 42	17.940		102.300		14.00		50.30		573.0	
9958	1.9573	+ 3	0.68310	+ 6	23.267	+ 34	15.894		45.543		7.74		15.39		148.0	

CHAPTER 10

HYPOTHESES FOR THE GENESIS OF ALKALINE ROCKS

10.1 INTRODUCTION

Mechanisms for the genesis of alkaline rocks include: (i) crystal fractionation (Holmes and Harwood, 1937; Pouclet, 1980; Katabarwa, 1983); (ii) magma mixing (Bailey, 1982; Carlson et al., 1983), (iii) assimilation and fractional crystallization (Holmes and Harwood, 1937; Green and Ringwood, 1967), and (iv) metasomatism (Sahama, 1972; Menzies and Murthy, 1980a and b; Kampunzu et al., 1983a and b; Vollmer et al., 1985).

Any model for the origin and evolution of the Birunga volcanics must explain the following features:

- (i) the wide range in SiO_2 contents of 36% to 61%;
- (ii) the extreme enrichment in alkalis;
- (iii) high concentrations of REE, especially light REE;
- (iv) very high Nb content of the nephelinites and the excellent positive correlation of Nb and Na_2O for all the Birunga volcanics;
- (v) the wide variation of 0.36 to 0.77 for Mg';
- (vi) higher Cr and Ni contents in the olivine leucitites and murambites than in the nephelinites;
- (vii) similar overall patterns of the incompatible elements for all samples;
- (viii) the variation in $^{87}\text{Sr}/^{86}\text{Sr}$ ratios from 0.70447 to 0.70971;
- (ix) positive correlation between $^{87}\text{Sr}/^{86}\text{Sr}$ and Rb/Sr ratio;

- 242
- (x) the variation in $^{143}\text{Nd}/^{144}\text{Nd}$ ratios from 0.51202 to 0.51269;
 - (xi) the negative slope in the plot of $^{143}\text{Nd}/^{144}\text{Nd}$ vs Sm/Nd ratio;
 - (xii) the excellent negative correlation between $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios for all of the Birunga samples.

10.2 THE ORIGIN OF ALKALIC ROCKS OF THE BIRUNGA FIELD AS INFERRED FROM ELEMENTAL AND ISOTOPIC STUDIES

10.2.1 CRYSTAL FRACTIONATION

The variation in $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ isotopic ratios of the Birunga samples rules out the generation of the Birunga rocks by simple magmatic fractionation. The fact that the Birunga rocks are not related by crystal fractionation, i.e. latites cannot be produced by crystal fractionation of nephelinites, is supported by phase equilibrium studies (Yoder, 1979). The phase equilibrium diagrams show a thermal barrier that prevents most undersaturated liquids from developing into oversaturated liquids (or vice versa) by crystal fractionation.

The variation in both Nd and Sr isotopic compositions places constraints on models proposed for the origin of the Birunga volcanics. The variations observed in the isotopic compositions of the Birunga volcanics can be explained by a variety of processes that include (O'Nions and Pankhurst, 1973):

- 1) disequilibrium melting of a homogeneous source.
- 2) melting of an inhomogeneous source.
- 3) mixing of two or more end members in varying proportions.

Hofmann and Hart (1978) have discussed the geochemical arguments for disequilibrium melting at length and concluded that because the

mantle is heterogeneous there is no need to resort to such a mechanism. Isotopic differences between mid-ocean ridge basalts and the oceanic island basalts as well as differences in basalts from different oceanic islands lead to the conclusion that the mantle is inhomogeneous both laterally and vertically (Hofmann and Hart, 1978). In addition, the linear correlation between $^{206}\text{Pb}/^{204}\text{Pb}$ vs $^{207}\text{Pb}/^{204}\text{Pb}$ for oceanic basalts has been interpreted, by some, as indicating an event that occurred ca 2 Ga ago. The latter, confirmed by Rb-Sr data (Sun and Hanson, 1975), indicates that isotopic heterogeneities have persisted for long periods of time within the upper mantle. Therefore the variation in isotopic ratios need not be explained by a disequilibrium partial melting phenomenon as long as a heterogeneous mantle exists to become locally homogenized during partial melting.

In the second model the assumption is made that Sr isotope equilibration occurs between the liquid and residual phases, and that the initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of a magma reflects the isotopic composition of the source region. Partial melting of an inhomogeneous source is one possible way to generate the Birunga volcanics, since the major and trace elements and isotopic data of the more primary magmas of the Birunga field (e.g. nephelinites and nepheline leucitites) are grossly different. The distinct variation in incompatible element concentrations, even in rocks with very similar major element abundances, suggests that these rocks are derived from a heterogeneous source by varying degree of partial melting. (Auchapt et al., 1987).

The third model involves mixing. Mixing can occur in the solid state such as convection of an inhomogeneous mantle, in the liquid state, such as between magmas, or between liquid and solid, such as

between a magma and its wall rock. Mixing can be modelled quantitatively.

10.2.2 QUANTITATIVE SIMPLE BINARY MIXING

The following equations can be used to quantitatively model the simple mixing of two components with different isotopic compositions (Faure, 1986):

$$\left(\frac{{}^{87}\text{Sr}}{{}^{86}\text{Sr}} \right)_M = \frac{[({}^{87}\text{Sr}/{}^{86}\text{Sr})_1 \times \text{Sr}_1 \times F] + [({}^{87}\text{Sr}/{}^{86}\text{Sr})_2 \times \text{Sr}_2 \times (1-F)]}{\text{Sr}_1 F + \text{Sr}_2 (1-F)} \quad (1)$$

A similar equation can be written for the Nd isotopes:

$$\left(\frac{{}^{143}\text{Nd}}{{}^{144}\text{Nd}} \right)_M = \frac{[({}^{143}\text{Nd}/{}^{144}\text{Nd})_1 \times \text{Nd}_1 \times F] + [({}^{143}\text{Nd}/{}^{144}\text{Nd})_2 \times \text{Nd}_2 \times (1-F)]}{\text{Nd}_1 F + \text{Nd}_2 (1-F)} \quad (2)$$

where: $({}^{87}\text{Sr}/{}^{86}\text{Sr})_M$ and $({}^{143}\text{Nd}/{}^{144}\text{Nd})_M$ indicate the isotopic character of the magma produced by mixing of different proportions of the two end-members. These differing proportions are defined as F where F varies from 0 to 1. $F = A/A+B$, where A and B are the weights of the two components in a given mixture. Subscript 1 refers to the first, and 2 the second end member.

The ${}^{87}\text{Sr}/{}^{86}\text{Sr}$ and ${}^{143}\text{Nd}/{}^{144}\text{Nd}$ isotopic ratios of the Birunga samples form a linear array in Fig. 9.6 (ACP). This linear array can be

interpreted in two ways: (i) mixing of two magmas, nephelinite and latite, and (ii) contamination of a nephelinitic magma by continental crust. The binary mixing equations (above) are used to determine the effect of mixing a nephelinite ($\text{Sr} = 3333 \text{ ppm}$; $\text{Nd} = 138 \text{ ppm}$; $^{87}\text{Sr}/^{86}\text{Sr} = 0.70469$; $^{143}\text{Nd}/^{144}\text{Nd} = 0.51268$) with a latite and also with various continental crust end-members (Fig. 10.1). Curve (A) indicates mixing with latite ($\text{Sr} = 623 \text{ ppm}$, $\text{Nd} = 80 \text{ ppm}$, $^{87}\text{Sr}/^{86}\text{Sr} = 0.70971$, $^{143}\text{Nd}/^{144}\text{Nd} = 0.51204$); curve (B) with an average granulite (lower crust) ($\text{Sr} = 150 \text{ ppm}$, $\text{Nd} = 20 \text{ ppm}$, $^{87}\text{Sr}/^{86}\text{Sr} = 0.713$, $^{143}\text{Nd}/^{144}\text{Nd} = 0.51106$; Hamilton et al., 1979; DePaolo et al., 1982); curve (C) with rocks of the Tanzanian shield (upper crust) ($\text{Sr} = 416 \text{ ppm}$, $\text{Nd} = 32 \text{ ppm}$, $^{87}\text{Sr}/^{86}\text{Sr} = 0.747$, $^{143}\text{Nd}/^{144}\text{Nd} = 0.511$; average values for granitoid rocks of the shield taken from Bell and Blenkinsop, 1987); curve (D) with 40% upper crust and 60% lower crust ($\text{Sr} = 256 \text{ ppm}$, $\text{Nd} = 25 \text{ ppm}$, $^{87}\text{Sr}/^{86}\text{Sr} = 0.7266$ and $^{143}\text{Nd}/^{144}\text{Nd} = 0.51104$).

Curve (A), Fig. 10.1, shows that it is difficult to generate the rocks of the Birunga field by mixing a nephelinite and a latite. The results also show that mixing of a nephelinitic magma (C3022) and a magma with a composition of a latite (C2894) cannot produce the intermediate members. Curves (B) and (C) clearly indicate that it is extremely difficult to generate the ACP array using either average granulite or upper crust as assumed crustal end-members. An end-member, however, that consists of 60% granulite and 40% upper crust, curve D, can produce a mixing curve similar to that found for the Birunga volcanics. A magma mixing program, IGPET, was also used to evaluate the mixing model using trace and major element data and conclusions similar to those above were obtained.

Fig. 10.1. The binary mixing model for the two end-members, nephelinite and latite (A); lower crust (B); upper crust (C); and 40% upper + 60% lower crust (D).

10.2.3 ASSIMILATION AND FRACTIONAL CRYSTALLIZATION (AFC) MODEL

The simple binary mixing model, discussed in section 10.2.2, is perhaps an oversimplification for the evolution of the Birunga igneous rocks. A more realistic model for the evolution of igneous rocks involves simultaneous assimilation and fractional crystallization. A model, formulated by DePaolo (1980), calculates the effects of simultaneous assimilation and fractional crystallization on the composition of a magma in which the mass assimilation rate ($\dot{M}_a$) is a fraction "r" of the fractional crystallization rate ($\dot{M}_c$), i.e. $r = \dot{M}_a / \dot{M}_c$. This is known as the assimilation and crystal fractionation (AFC) model.

The following equations can be used to monitor the effect of simultaneous assimilation and fractional crystallization on elemental abundances (a) as well as on the isotopic ratios (b): Note: $r \neq 1$.

$$\frac{C_m}{C_m^0} = F^{-z} + \left(\frac{r}{r-1} \right) \frac{C_a}{z C_m^0} (1 - F^{-z}) \quad (a)$$

$$I_m = \frac{\frac{r}{r-1} \frac{C_a}{z} (1 - F^{-z}) I_a + C_m^0 F^{-z} I_m^0}{\frac{r}{r-1} \frac{C_a}{z} (1 - F^{-z}) + C_m^0 F^{-z}} \quad (b)$$

Where:

$\dot{M}_a$ = the rate of mass assimilation

$\dot{M}_c$ = the rate of fractional crystallization

$r = \dot{M}_a / \dot{M}_c$ i.e. (if $r > 1$, the mass of the magma increases while if $r < 1$)

the mass of the magma decreases).

C_m = concentration of the element in the magma;

C_m^0 = the original concentration in the magma;

$F = M_m/M_m^0$, M_m = mass of the magma body, M_m^0 = initial mass of the magma;

$z = (r+D-1)/(r-1)$, D = solid/liquid partition coefficient = C_a/C_m ;

C_a = the concentration of the element in the assimilated wallrock;

I_m = isotope ratio in the magma;

I_a = isotope ratio in the wallrock;

I_m^0 = the original isotope ratio in the magma. "I" can be replaced by any isotope ratio (e.g. $^{87}\text{Sr}/^{86}\text{Sr}$, $^{143}\text{Nd}/^{144}\text{Nd}$) or any normalized parameter describing such ratios (ϵ_{Sr} , ϵ_{Nd}). The derivation of the above formulas are given in DePaolo (1980).

In the above equations of the AFC model it is assumed that "r" is not equal to 1, i.e. the mass of the magma does not remain constant, and that the isotope ratio of the crystallizing phases is always the same as the magma (the crystallizing phases are in isotopic equilibrium with the magma).

The AFC model was used to evaluate the effect of combined assimilation and contamination on an assumed nephelinite parent magma. In order to model the AFC process quantitatively, a primary magma, a contaminant and the two variables, "r" and "D" are chosen. The initial magma is assumed to have the same isotopic composition of one of the Niragongo nephelinites, C3022, with $\text{Sr} = 3333 \text{ ppm}$; $^{87}\text{Sr}/^{86}\text{Sr} = 0.70469$; $\text{Nd} = 138 \text{ ppm}$; $^{143}\text{Nd}/^{144}\text{Nd} = 0.51268$. The other end-members chosen are: 1) the upper crust, 2) the lower crust, 3) 1:1 ratio of the upper and lower crust, and 4) a mixture of upper and lower crust (40% Tanzanian shield + 60% granulite, upper and lower crust respectively). This end-

member is known to give a mixing curve which fits the data using the simple binary mixing model (see section 10.2.2).

For the purposes of calculation "r" was arbitrarily assigned values of 0.005, 0.01, 0.05, 0.1, 0.2, 0.3, 0.4 and 0.5. In order to cover most of the fractionating minerals, a range of possible "D" values, 0.003 to 4.4, were chosen for Sr and Nd for each of these "r" values. The straight line relationship of $^{143}\text{Nd}/^{144}\text{Nd}$ vs $^{87}\text{Sr}/^{86}\text{Sr}$ ratios on the ACP can be produced using the AFC model only if the D^{Sr} and D^{Nd} have the same magnitude for specific value of "r". If D^{Sr} differs from D^{Nd} , the mixing line is curved, i.e. the curvature of the mixing line on the ACP is related to the Sr/Nd ratio of the two end-members and a straight line indicates a simple mixing of two end-members with the same Sr/Nd ratio.

The minerals that are found as phenocrysts in the Birunga rocks do not have the same D values for Sr and Nd, and it therefore seems unlikely that the genesis of these rocks can be explained by the AFC model. In addition, the clustering of the data points close to the nephelinitic end-member on the AFC plot for all possible r and D values indicates that this model is not applicable to the genesis of Birunga rocks and cannot explain the intermediate members of the array shown on the ACP.

10.2.4 METASOMATISM

Metasomatism is the chemical transformation of a rock in the solid state either by diffusion or infiltration of a fluid and/or vapour from deeper sources (Hofmann and Hart, 1978). Since solid-state diffusion is a slow process, a fluid or melt phase is required to

enhance the effect. These fluid phases include, $\text{CO}_2\text{-H}_2\text{O}$ fluid (Eggler, 1978), hydrous silicate melt (Frey and Green, 1974), and carbonated alkaline melt (Meen, 1987).

There is a well-developed linear relationship between $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{87}\text{Rb}/^{86}\text{Sr}$ for all the Birunga samples. If a line is drawn through the data points and it is assumed to be an isochron then it yields a date of around 500 Ma for all samples except Sabinyo shoshonites and latites. This date agrees with the date based on the Pb isotopic ratios obtained by Vollmer and Norry (1983a) for nephelinites from Niragongo. On the basis of this similarity the 500 Ma figure may reflect an event that produced uniform Pb and Sr isotopic compositions within the source region of the lavas. The uniform Sr isotopic composition and variable Rb/Sr ratios can be considered the result of metasomatism. The constant Rb and K contents of the Birunga volcanics suggests that the variation in Rb/Sr ratios is mainly due to variation in Sr content.

Three other aspects of the geochemistry of these lavas can also be explained by metasomatism: (i) the high Nb content of some of these rocks, (ii) the variation in the REE abundances and incompatible trace element patterns, and (iii) similarity of the trace element pattern to the average for metasomatized mantle nodules. The Niragongo nephelinites and nepheline leucitites, which are the most deep-seated of the rock types from Birunga, also contain the highest REE, Ba, Na_2O , Nb, P and Sr abundances of any of the volcanics. If these geochemical characteristics can be attributed to metasomatizing fluids from a deeper source, then the fluid would have to have these geochemical features.

The latites and shoshonites of Sabinyo are the most evolved of the Birunga rock types in terms of chemical and isotopic compositions.

They have the highest $^{87}\text{Sr}/^{86}\text{Sr}$ and lowest $^{143}\text{Nd}/^{144}\text{Nd}$ ratios. On the plot of $^{87}\text{Sr}/^{86}\text{Sr}$ vs Rb/Sr the slope based on the data for shoshonites and latites of Sabinyo is different from that obtained from the data for the remainder of the samples. If the slope is interpreted as an isochron then it corresponds to a date of 187 Ma and an initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.70624. It is possible that a second metasomatic event occurred that affected only the source below Sabinyo.

The excellent linear correlation between $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ in the ACP for most of the Birunga samples means that these rocks are genetically related. The quantitative mixing modelling indicates that mixing of two end-members represented by the nephelinites and murambites can produce the correlation observed in the ACP (Fig. 9.6). However, the Rb-Sr data still fall along a line on an isochron diagram with a slope corresponding to 500 Ma which means that the mixing event that produced the correlation on the ACP did not destroy the evidence for the Lower Paleozoic event.

The high SiO_2 content (average 58%), high $^{87}\text{Sr}/^{86}\text{Sr}$ and low $^{143}\text{Nd}/^{144}\text{Nd}$ ratios of the Sabinyo samples are difficult to explain if only a mantle source is involved. The similarity of the Sm/Nd ratio of 0.15 of the samples to a lower crustal value of 0.18 (the average of 17 granulite samples from Zaire, Delhal et al., 1986), and the similarity of the average Nb/U ratio (15) to an average continental crustal value of 10 (Hofmann et al., 1986) (Table 10.1), suggests involvement with continental crust. Since the Sabinyo rocks have high REE abundances and their trace element pattern parallels the one for known metasomatized mantle nodules, a metasomatized crustal source could be one explanation of the high $^{87}\text{Sr}/^{86}\text{Sr}$ and low $^{143}\text{Nd}/^{144}\text{Nd}$ ratios. The involvement of

Table 10.1

Nb and U concentrations for several Birunga samples.
 U concentrations are taken from Vollmer and Norry
 (1983a), Nb concentrations from this study.

Sample #	Rock type	Nb (ppm)	U (ppm)	Nb/U
C1904	kivite	91	2.34	38.9
C2794	absarokite	117	3.83	30.6
C2799	banakite	161	5.15	31.3
C2888	latite	89	6.07	14.5
C2909	mikenite	194	4.18	46.4

both metasomatized mantle and crustal sources could also explain the wide variation of the Pb isotopic data for the Niragongo volcanics and the restricted range for the Sabinyo samples.

10.2.5 DISCUSSION

From the previous discussions it can be seen that many of the features of the Birunga volcanics, but not all, can be explained by either metasomatism or mixing between nephelinite and continental crust. Because of the high REE concentrations and high Nb and Ba abundances for the nephelinites, it seems unlikely that a "normal" upper mantle was involved in the genesis of the Birunga volcanics. It could well be that both metasomatism and mixing with continental crust were involved in the genesis of the Birunga rocks. Many problems in the interpretation of the isotopic data still remain unresolved. These problems include:

- (i) the nature and source of the metasomatizing fluids,
- (ii) the precise time of mixing that generated the distribution of the data on the ACP,
- (iii) the nature of the end-members involved in mixing,
- (iv) the isotopic characteristics of the pre-metasomatic mantle, and
- (v) the retention of the 500 Ma event in the Birunga volcanics.

CHAPTER 11

CONCLUSION

The major findings of this thesis are summarized below:

11.1 GRAVITY

(1) A very long wavelength negative gravity anomaly exists over the rift zone of east Africa. The largest negative anomaly is located at a latitude of 0 in the eastern rift (ER) and at 1.7°S in the western rift (WR). In ER there is a small axial linear positive anomaly superimposed on the broad negative anomaly.

(2) The negative anomaly has a deep source and is due to variation in lithospheric thickness. The positive anomaly in ER is due to the injection of mantle material into the crust.

(3) Gravity modelling shows that the lithosphere is thinnest at the Equator in ER and at 1.7°S in WR. Quantitative modelling also indicates that lithospheric thickness in the rift zone varies both across and along the rift axis.

11.2 SEISMICITY AND SEISMIC WAVE PROPAGATION

(1) WR is more seismically active than ER. The latter is aseismic north of 3°S . Microseismic data indicate activity at the intersection of faults of different directions and at shallow focal depths.

(2) The rift system is associated with a delay of body waves travel time which reveal a substantial low velocity zone in the upper mantle. The delay time is less in the western rift compared to the eastern rift.

(3) The attenuation of surface waves indicates that lithospheric mantle is missing and that the asthenosphere has come up to the level

of the Moho in ER at around the Equator. The thickness of the lithospheric mantle increases towards the south. The S-wave velocities are higher under the crust in WR as compared to ER but still lower than normal. ●

(4) There is an inverse correlation between the frequency of the microseismic activity and the thickness of the lithosphere. These properties vary asymmetrically across EARS.

11.3 HEAT FLOW

(1) The rift system is associated with a zone of high heat flow approximately 60 ± 10 km wide. In ER, the heat flow data shows no trend along the rift but does show a systematic pattern across it. In the WR heat flow variation is systematic both along and across the rift.

(2) Using the geotherms corresponding to heat flow values of 111, 90 and 54 mW/m^2 for the ER, WR and the Precambrian shield respectively, the thermal lithospheric thicknesses are estimated to be 40 km in ER, 55 km in WR and 115 km in the Precambrian shield. The thermal lithospheric thickness varies not only across the rift but also along it, at least in the western branch.

(3) In ER a zone of high conductivity is detected at a depth of 20 km. This zone is due to enhanced geothermal gradients and possibly partial melting in the upper mantle.

11.4 RHEOLOGY

(1) the rheological structure of the lithosphere derived from geophysical data is as follows: In ER only the upper 12 km deforms by brittle fracture below which the deformation is by ductile flow; in the

WR the brittle layer continues down to a depth of about 18 km below which the deformation is ductile; in the Precambrian shield the brittle zone has a thickness of 57 km, and below this depth the deformation is ductile; a ductile layer is also possibly present between 29 and 40 km.

(2) Lithospheric thicknesses deduced from the rheological profiles (defined as the depth at which the creep strength is 1 MPa) are: 41 ± 7 km in ER, 55 ± 9 km in WR, and 115 ± 28 km in the adjacent Precambrian shield. Along-strike variation in the lithospheric thickness is evident.

(3) Calculation of the effective viscosities confirms the lithospheric thicknesses of 30-40 km for ER, 50-60 km for WR and 90-100 for the Precambrian shield, taking 10^{21} Pa s as the lower boundary of the lithosphere. Therefore, either the creep strength or the effective viscosity can be used to infer the lithospheric thickness.

(4) The average total strength of the lithosphere is 27 MPa in ER, 62 MPa in WR and 188 MPa for the Precambrian shield. The total strength of the lithosphere in the Precambrian shield (taken to represent the condition required for initiation of the rift) indicates that a combination of several sources of lithospheric stress is necessary to initiate rifting (e.g. plateau uplift plus other intraplate stresses).

11.5 MAJOR AND TRACE ELEMENT DATA

(1) Major element data indicate two extreme chemical end-members in the Birunga field, the nephelinites of Niragongo which are the most silica-undersaturated and the latites of Sabinyo which are the most silica-saturated rocks of the Birunga samples.

(2) Major and trace element data indicate chemical changes from

east to west. Cr and SiO_2 decrease from east to west, whereas Ba, CaO, total Fe, MnO, Na_2O , Nb, Ni, P_2O_5 , Sr, and Zn increase from east to west.

(3) Incompatible element patterns (spidergrams) are very similar for all samples. Relative to the primitive mantle the Birunga volcanics are enriched in Ba, Rb, K, Nb, Sr, P, Zr, Ti, and Y. These elements are listed in order of decreasing enrichment relative to the primitive mantle.

(4) All samples analysed show enrichment in the REE, especially in the LREE, relative to chondrites and show subparallel REE patterns. The nephelinites are more enriched in REE than latites.

(5) There is an excellent positive correlation between Nb and Na_2O (correlation coefficient = 0.95) as well as between Ba and Na_2O (correlation coefficient = 0.88) for all of the Birunga volcanic rocks. Nephelinites have the highest Nb and Na_2O contents of any rock types from the Birunga field.

11.6 ISOTOPIC SYSTEMATICS

(1) The $^{87}\text{Sr}/^{86}\text{Sr}$ ratios for the Birunga volcanics range from 0.70447 to 0.70971.

(2) Nephelinites show low $^{87}\text{Sr}/^{86}\text{Sr}$ (average 0.70461), low Rb/Sr ratio (0.042) and high Sr (2873 ppm) content. Latites, on the other hand, have high $^{87}\text{Sr}/^{86}\text{Sr}$ (average 0.70959), high Rb/Sr ratio (0.341) and low Sr (633 ppm) contents.

(3) A negative correlation exists between $^{87}\text{Sr}/^{86}\text{Sr}$ and Sr, plus a positive correlation between $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{87}\text{Rb}/^{86}\text{Sr}$.

(4) The range in $^{143}\text{Nd}/^{144}\text{Nd}$ for 33 Birunga samples is 0.51202 to

0.51269. Nephelinites show the highest and latites the lowest values for the $^{143}\text{Nd}/^{144}\text{Nd}$ ratios.

(5) There is a negative slope when $^{143}\text{Nd}/^{144}\text{Nd}$ values are plotted against $^{147}\text{Sm}/^{144}\text{Nd}$.

(6) A very strong negative correlation exists between $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ (ACP) for all Birunga samples (correlation coefficient = 0.98). The Niragongo nephelinites and nepheline leucitites plot in the depleted quadrant very close to the bulk Earth. The rest of the samples fall within the enriched field.

11.7 MODELLING

(1) The variation in Sr and Nd isotopic ratios of the Birunga samples indicates that Birunga volcanics cannot be produced by simple magmatic fractionation.

(2) The intermediate compositions between the nephelinites and latites cannot be explained by mixing of nephelinite and latite magmas.

(3) The quantitative binary mixing modelling for the isotopic data indicates that it is very difficult to generate the array of data on the ACP by mixing nephelinites with (a) latites, (b) with upper crust, and (c) with lower crust. But mixing of a nephelinitic magma with a 60% lower crust + 40% upper crust can indeed explain the generation of different rock types in the Birunga field.

(4) A metasomatic model can explain some of the characteristics of the Birunga volcanics. For example, the high abundances of Nb and REE and the similarity of incompatible trace element patterns to that for the average metasomatized mantle nodules can be explained by such a model. The agreement between the Rb-Sr and U-Pb dates of about 500 Ma suggests that metasomatism may have taken place at this time.

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Appendix 3.1

The average $1^\circ \times 1^\circ$ altitude, Free Air (FA) and the calculated Bouguer anomaly (BA) in mgals.

Lat	Long	FAA	Alt	BA					
15	20	10	474	43.0	13	23	-22	425	69.6
15	21	5	503	51.3	13	24	11	401	33.9
15	22	3	530	56.3	13	25	10	494	45.3
15	23	13	587	52.7	13	26	14	642	57.8
15	24	12	603	55.5	13	27	15	697	63.0
15	25	14	651	58.8	13	28	19	868	78.1
15	26	16	734	66.1	13	29	4	966	104.1
15	27	15	688	62.0	13	30	42	1437	118.8
15	28	15	715	65.0	13	31	-34	883	132.8
15	29	17	808	73.4	13	32	2	1043	114.7
15	30	18	822	74.0	13	33	18	1104	105.5
15	31	-4	732	85.9	13	34	10	1276	132.8
15	32	28	991	82.9	13	35	9	873	88.7
15	33	33	1100	90.1	13	36	15	762	70.3
15	34	21	661	53.0	13	37	21	834	72.3
15	35	8	672	67.2	13	38	1	650	71.7
15	36	-21	690	98.2	13	39	10	500	45.9
15	37	41	1323	107.0	13	40	-9	377	51.2
15	38	36	1674	151.3	12	20	-28	356	67.8
15	39	25	1082	96.1	12	21	-40	368	81.2
14	20	10	488	44.6	12	22	-27	372	68.6
14	21	11	505	45.5	12	23	-31	380	73.5
14	22	11	525	47.7	12	24	-43	399	87.6
14	23	9	450	41.3	12	25	-28	414	74.3
14	24	11	523	47.5	12	27	-16	625	85.9
14	25	2	628	68.3	12	28	13	808	77.4
14	26	20	666	54.5	12	29	7	894	93.0
14	27	15	658	58.6	12	30	1	1132	125.6
14	28	18	834	75.3	12	31	-15	1035	130.8
14	29	6	834	87.3	12	32	1	1084	121.3
14	30	14	1119	111.2	12	33	-17	1067	136.4
14	31	-7	877	105.1	12	34	7	1524	163.5
14	32	11	1078	109.6	12	35	36	1453	126.6
14	33	26	1222	110.7	12	36	31	1237	107.4
14	34	44	1054	73.9	12	37	32	971	76.6
14	35	3	608	65.0	12	38	-10	434	58.6
14	36	15	645	57.2	12	39	-10	256	38.6
14	37	-2	720	82.6	12	40	-6	193	27.6
14	38	28	885	71.0	11	20	-36	366	77.0
14	39	36	1048	81.2	11	21	-26	381	68.6
14	40	12	798	77.3	11	22	-42	383	84.9
13	20	-33	385	76.1	11	23	-35	393	79.0
13	21	-62	375	104.0	11	24	-36	368	77.2
13	22	-35	393	79.0	11	25	-26	424	73.4

11	26	7	625	62.9	09	35	24	1846	182.5
11	27	9	715	71.0	09	36	1	1528	170.9
11	29	4	1215	131.9	09	37	42	1286	101.9
11	30	15	1280	128.2	09	38	-21	631	91.6
11	31	7	1239	131.6	09	39	-11	173	30.4
11	32	30	1143	98.9	09	40	-13	66	20.4
11	33	18	1132	108.6	08	20	-55	390	98.6
11	34	1	1469	164.3	08	21	-57	486	111.4
11	35	50	1910	163.7	08	22	-46	489	100.7
11	36	20	1698	170.0	08	23	-31	566	94.3
11	37	18	1366	134.8	08	24	-39	598	105.9
11	38	14	583	51.2	08	25	-19	580	83.9
11	39	-7	255	35.5	08	26	-17	675	92.5
11	40	-9	144	25.1	08	27	-25	885	124.0
10	20	-59	353	98.5	08	28	2	1384	152.8
10	21	-49	336	86.6	08	29	63	1770	135.0
10	22	-37	344	75.5	08	30	6	1517	163.7
10	23	-42	439	91.1	08	31	2	1338	147.7
10	24	-37	530	96.3	08	32	-7	1186	139.7
10	25	-17	536	77.0	08	33	-11	1237	149.4
10	26	-22	600	89.1	09	34	-1	1455	163.8
10	27	-19	797	108.2	08	35	-3	1623	184.6
10	28	8	1242	131.0	09	36	16	1392	139.7
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10	31	8	1223	128.8	08	39	-16	191	37.4
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10	38	-8	563	71.0	07	26	-11	723	91.9
10	39	-13	225	38.2	07	27	-13	925	116.5
10	40	-12	139	27.6	07	28	70	1934	146.4
09	20	-65	383	107.9	07	29	10	1455	152.8
09	21	-51	430	99.1	07	30	18	1489	148.6
09	22	-42	442	91.5	07	31	6	1352	145.3
09	23	-44	446	93.9	07	32	4	1333	145.1
09	24	-53	566	116.3	07	33	-26	1215	161.9
09	25	-19	553	80.9	07	34	-19	1483	184.9
09	26	-9	754	93.4	07	35	-1	1559	175.4
09	27	-11	954	117.7	07	36	2	1342	148.1
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09	29	87	1829	117.6	07	38	37	710	42.4
09	30	24	1555	150.0	07	39	-5	163	23.2
09	31	-1	1281	144.3	06	20	-30	570	93.8
09	32	-7	1134	133.9	06	21	-32	572	96.0
09	33	-9	1170	139.9	06	22	-20	590	86.0
09	34	1	1477	165.2	06	23	-34	597	100.8

06	24	-32	600	99	1	04	36	31	1272	111	3
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06	26	-28	601	95	2	04	38	-10	206	33	1
06	27	-20	813	111	0	03	20	-2	944	96	4
06	28	7	1146	121	2	03	21	-11	807	101	3
06	29	-23	961	135	5	03	22	-20	825	112	3
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06	38	-1	695	78	8	03	31	-19	1315	166	1
05	20	-11	601	78	2	03	32	6	1397	150	3
05	21	-19	581	84	0	03	33	15	1457	148	0
05	22	-12	610	80	2	03	34	5	1390	150	5
05	23	-33	605	100	7	03	35	-6	1216	142	0
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05	29	18	916	84	5	02	22	-7	956	114	0
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05	31	-13	1088	134	7	02	24	15	1061	103	7
05	32	4	1291	140	4	02	25	-4	962	111	6
05	33	1	1304	145	9	02	26	-36	679	112	0
05	34	4	1355	147	6	02	27	30	1378	124	2
05	35	-15	1232	152	8	02	28	-1	1185	133	6
05	36	26	1471	138	6	02	29	-16	1122	141	5
05	37	20	1190	113	1	02	30	-32	1009	144	9
05	38	7	342	31	3	02	31	21	1575	155	2
04	20	-7	628	77	3	02	32	-29	1124	154	8
04	21	-7	622	76	6	02	33	5	1435	155	5
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04	25	-15	713	94	8	02	37	-27	314	62	1
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04	27	-28	627	98	2	01	20	-7	1000	118	9
04	28	-9	941	114	3	01	21	-7	1000	118	9
04	29	33	1095	89	5	01	22	-2	1000	113	9
04	30	1	1193	133	5	01	23	13	1021	101	2
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04	32	-7	1220	143	5	01	25	-29	817	120	4
04	33	17	1490	149	7	01	26	1	1156	129	3
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04	35	-32	1048	149	2	01	28	-9	1105	132	6

01	29	1	1200	134	3
01	30	12	1430	148	0
01	31	21	1474	143	9
01	32	36	1470	128	5
01	33	15	1433	145	3
01	34	35	1641	148	9
01	35	-15	921	118	0
01	36	-16	649	88	6
01	37	-1	549	62	4
01	38	-17	343	55	4
00	20	-7	1200	141	3
00	21	-72	1200	206	3
00	22	13	1200	121	3
00	23	-47	1073	167	0
00	24	-22	1135	149	0
00	25	9	1287	135	0
00	26	3	1248	136	6
00	27	1	1203	134	6
00	28	-9	1129	135	3
00	29	-42	1192	175	4
00	30	10	1405	147	2
00	31	16	1430	144	0
00	32	5	1450	157	2
00	33	10	1770	188	0
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00	35	17	987	93	4
00	36	25	845	69	5
00	37	15	647	57	4
00	38	-15	419	61	9
00	39	13	345	25	6

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CANADIAN THESES

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NOTICE

AVIS

APPENDIX 4.1

Time, location, depth and magnitude of all the earthquakes in EARS from 1912 to 1983

Year	Month	Day	Hour	Minute	Second	Latitude	Longitude	Depth	Magnitude
1983	01	15	16	34	078	0513N	30199E	1052MB	
1982	07	24	03	51	016	1220N	30107E	1045MB	
1979	03	09	00	14	087	1252N	30567E	3550MB	
1979	12	04	23	44	48	1754N	31294E	3350MB	
1979	02	26	44	03	59	1858N	30936E	1049MB	
1981	07	23	06	21	520	2183N	20076E	1347MB	
1979	11	01	21	50	069	2627N	22132E	1046MB	
1982	01	05	08	37	537	4667N	31733E	1051MB	
1945	03	18	08	01	260	0000N	03200E		
1977	12	29	11	50	380	0013N	029683E	03343MB	
1971	04	18	05	49	000	0169N	030196E	03247MB	
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1928	01	06	19	31	580	0050N	036500E		
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APPENDIX 8.1

MAJOR ELEMENT CHEMISTRY (XRF METHOD)

Analyses of major elements, Rb, Sr, Ba, Zr, Y, Nb, Zn, Ni, and Cr were carried out on a Phillips 1410 automated x-ray fluorescence (XRF) unit at University of Ottawa, Ottawa. The determinations were made on a fused disk composed of 3.9g of $\text{Li}_2\text{B}_4\text{O}_7$, 0.433g of Li_2CO_3 , and 1.3g of whole rock powder. The standard used was SY-2 (see Abbey, 1983). Using this standard, the precision of the results (given in percentage of amount present) was determined to be:

- + 2% for SiO_2 , Al_2O_3 , Fe_2O_3 (total), and K_2O
- + 3% for MgO , TiO_2 , Na_2O , and CaO
- + 10% for P_2O_5 , MnO , Rb, Sr, Ba, Zr, Cr, and Ni

Accuracy - felsic standard SY-2

Major-elements

	SiO_2	Al_2O_3	Fe_2O_3	MgO	CaO	Na_2O	K_2O	TiO_2	P_2O_5	MnO
R.V.	60.09	12.15	6.29	2.61	8.00	4.35	4.51	0.15	0.42	0.32
X	60.18	12.23	6.25	2.69	7.96	4.37	4.52	0.14	0.43	0.33
S.D. (+)	.17	.10	.04	.05	.06	.07	.04	.01	.02	0.00

Trace-elements

	<u>Ba</u>	<u>Cr</u>	<u>Zr</u>	<u>Sr</u>	<u>Rb</u>	<u>Y</u>	<u>Zn</u>	<u>Ni</u>
R.V.	460	10.0	270	270	220	130	250	10.0
X	467	9.8	269	256	222	123	239	2.1
S.D. (+)	5.2	8.1	3.7	7.2	3.9	2.5	31.7	3.2

R.V. = recommended value- Abbey (1983).

X = mean measured value, based on 10 fusions of SY-2, all run once.

S.D. = 1 standard deviation of the measured value.

11

[illegible]

Appendix 8.2 (ctd)

	2849	2856	2871	2888	2889	2894	2895	2896	2897
Ac	-----	-----	-----	-----	-----	-----	-----	-----	-----
Ab	13.17	17.39	30.88	25.55	25.64	26.59	29.44	8.19	7.49
An	13.58	17.59	16.97	14.03	14.79	13.68	15.52	15.41	15.19
Ap	1.33	1.25	2.14	1.07	1.35	1.10	1.64	0.72	1.21
CO	-----	-----	-----	-----	-----	-----	-----	-----	-----
Di	25.73	14.38	4.30	4.74	7.22	1.96	4.89	30.54	31.81
He	-----	-----	-----	-----	-----	-----	-----	-----	-----
Hy	-----	-----	3.24	5.59	6.74	5.20	7.01	-----	-----
Il	5.58	6.37	3.92	3.28	4.50	3.05	4.77	5.42	5.53
Ka	-----	-----	-----	-----	-----	-----	-----	-----	-----
Le	-----	-----	-----	-----	-----	-----	-----	-----	-----
Mg	6.44	7.21	5.17	4.68	5.62	4.53	5.83	6.33	6.40
Ne	4.54	2.99	-----	-----	-----	-----	-----	4.94	5.86
Ol	8.16	8.50	2.37	-----	-----	-----	13.00	-----	9.25
Or	21.51	24.34	31.05	29.12	28.04	30.52	30.43	15.50	17.28
Q	-----	-----	-----	11.95	6.13	13.40	0.51	-----	-----
ST	-----	-----	-----	-----	-----	-----	-----	-----	-----
Wo	-----	-----	-----	-----	-----	-----	-----	-----	-----

	2905	2909	2911	2912	2915	3022	3023	3033	4414
Ac	-----	-----	-----	-----	-----	11.25	-----	-----	4.10
Ab	-----	-----	2.16	5.00	21.13	-----	-----	0.88	-----
An	6.21	6.30	2.34	2.78	19.00	-----	5.31	20.22	-----
Ap	1.81	2.04	2.01	2.02	1.43	4.00	4.96	0.05	2.28
CO	-----	-----	-----	-----	-----	20.33	5.84	-----	22.53
Di	30.79	28.02	16.43	14.39	15.27	-----	19.28	37.92	7.28
He	-----	-----	-----	-----	-----	-----	-----	-----	-----
Hy	-----	-----	-----	-----	-----	-----	-----	-----	-----
Il	7.01	6.88	5.36	5.04	6.13	4.48	6.04	7.37	8.78
Ka	-----	-----	-----	-----	-----	14.03	-----	-----	-----
Le	18.82	18.67	-----	-----	-----	6.97	23.14	-----	22.83
Mg	7.60	7.49	6.29	6.03	6.87	-----	6.80	7.81	7.02
Ne	16.82	17.27	23.73	21.48	1.65	22.35	22.10	7.24	2.73
Ol	0.08	-----	-----	-----	7.12	16.55	6.63	8.44	22.49
Or	10.89	13.27	39.16	41.12	21.41	-----	-----	10.07	-----
Q	-----	-----	-----	-----	-----	-----	-----	-----	-----
ST	-----	-----	-----	-----	-----	0.14	-----	-----	-----
Wo	-----	0.11	-----	2.17	-----	-----	-----	-----	-----

Appendix 8.2 (ctd)

	6066	9958
Ac	6.16	-----
Ab	-----	-----
An	-----	5.59
Ap	0.99	3.30
CO	15.65	4.51
Di	-----	13.72
He	-----	-----
Hy	-----	-----
Il	7.07	4.30
Ka	0.39	-----
Le	31.07	27.27
Mg	4.77	5.55
Ne	-----	30.03
Ol	32.38	5.79
Or	-----	-----
Q	-----	-----
ST	-----	-----
Wo	-----	-----

The following abbreviations have been used in the table above:
 Acmite, Ac; Albite, Ab; Anorthite, An; Apatite, Ap; Calcium
 orthosilicate, CO; Diopside, Di; Hematite, He; Hypersthene, Hy;
 Ilmenite, Il; Kaliophilite, Ka; Leucite, Le; Magnetite, Mg;
 Nepheline, Ne; Olivine, Ol; Orthoclase, Or; Quartz, Q; Sodium
 metasilicate, ST; Wollastonite, Wo.

APPENDIX 9.1: PROCEDURES FOR ISOTOPIC STUDIES

DISSOLUTION AND PREPARATION OF THE WHOLE-ROCK POWDER FOR ISOTOPIC STUDIES

- 1) 200 mg of rock powder is taken into solution in a Teflon evaporating dish using 20 ml of HF and 3-4 drops of HClO_4 .
- 2) The sample is taken to dryness on a hot plate at low setting (100 °C).
- 3) The sample is heated strongly (about 250 °C) to drive off any remaining acid.
- 4) After the dish has cooled, the sample is taken into solution with about 5 ml of 2.5N HCl.

Note: If the sample is to be spiked, the solution is split and transferred into two weighed disposable beakers. The precise weight of beakers plus solutions are recorded. One aliquot is spiked, and both aliquots are then taken to dryness.

- 5) The sample is taken into solution with 5 ml of 2.5N HCl, transferred into a centrifuge tube and is centrifuged for approximately 10 minutes.
- 6) The solution is then transferred to a disposable beaker and taken to dryness at 100 °C.
- 7) Then 2 ml of 2.5N HCl is added to the dried sample and the solution is centrifuged again for 10 minutes.

The sample is now ready to be loaded on to the columns.

CALIBRATION FOR SEPARATING Rb, Sr AND REE FROM WHOLE ROCK POWDERS FOR MASS SPECTROMETRY: FIRST COLUMNS

Materials:

Teflon tube 22 cm in length with an internal diameter of 1.1 cm, Bio-Rad AG50W-X8 cation exchange resin (200-400 mesh, hydrogen form). The resin is equilibrated with 2.5N HCl. The columns are packed with 13 ml of the resin, cleaned with 100 ml of 6N followed by 2.5N HCl.

8) The prepared sample (after step 7 above) is loaded directly on top of the resin.

9) After the sample is completely absorbed into the resin, it is washed with 5 ml of 2.5N HCl in 1 ml portions to ensure that the sample begins to pass through the resin.

10) When all of the acid is absorbed, the resin is washed with 30 ml of 2.5N HCl.

11) Rb is then collected with 10 ml of 2.5N HCl in clean, labelled disposable beakers.

12) The columns are washed with 25 ml of 2.5N HCl.

13) The Sr is eluted with 15 ml of 2.5N HCl.

14) The resin is then washed with 20 ml of 2.5N HCl.

15) The resin is also washed with 15 ml of 6.2N HCl.

16) 25 ml of 6.2N HCl is added and the rare earth elements are collected.

17) The collected solutions are put on the hot plate.

18) When the Rb and Sr aliquots are evaporated to 1-2 ml, they are transferred into clean, small Teflon dishes and taken to dryness. These samples are now ready for analyses in the mass spectrometry laboratory.

19) The REE aliquot is taken to dryness and prepared for Nd-Sm separation in the second set of columns.

20) The columns are washed with 200 ml of 6N HCl followed by 100 ml of 2.5N HCl.

SUMMARY

Load Sample	2 ml	2.5N HCl	
wash	5 ml	2.5N HCl	
wash	30 ml	2.5N HCl	
collect	10 ml	2.5N HCl	Rb
wash	25 ml	2.5N HCl	
collect	15 ml	2.5N HCl	Sr
wash	20 ml	2.5N HCl	
wash	15 ml	6.2N HCl	
collect	25 ml	6.2N HCl	REE

IF ONLY COLLECTING REE:

Load Sample	2 ml	2.5N HCl	
wash	5 ml	2.5N HCl	(in 1 ml portion)
wash	100 ml	2.5N HCl	
wash	15 ml	6.2N HCl	
collect	25 ml	6.2N HCl	REE

SEPARATION OF Nd AND Sm FROM REE: SECOND SET OF COLUMNS

Materials and column preparation:

Supersil quartz (vycor) column with 100 ml reservoir and 50 cm of tubing with 0.5 cm internal diameter and a 2 mm diameter hole 12.5 cm from the bottom of the tube, porous polyethelene disc 35 microns in thickness, Teflon powder (Marque Voltalef Deposee), Bio Rad AG1X-8 anion exchange resin (200-400 mesh, chloride form), di(2-ethylhexyl) orthophosphoric acid (DEHPA), acetone, Teflon tape.

The Teflon powder is mixed with DEHPA in a 10:1 (powder: acid) proportion. Acetone is then added and the mixture is left on the magnetic stirrer until the Teflon powder (resin) becomes coated.

The polyethelene disc is inserted in the bottom of the column and the hole in the column is sealed with clean Teflon tape (washed with suprapure nitric acid and DDW). The coated Teflon powder is packed into the columns until it reaches a height of 10 cm; the Teflon is protected by an additional 0.5 cm of AG1X-8 resin. The column is preconditioned with 0.25N HCl.

The columns were calibrated using a tracer composed of Eriochrome Black T (EBT), methanol, and NH_4OH in equal proportions (1 mg: 1 ml: 1 ml)*. This indicator is sensitive to rare earths.

PROCEDURE

1) The dried sample from step 20 above is taken into solution with 0.5 ml of 0.25N HCl and left for at least half an hour to ensure that the solute is completely dissolved.

2) The sample is then injected through the hole of the column directly on top of the resin, and washed with 0.3 ml of 0.25N HCl (also injected).

* The column was calibrated using Nd and Sm solutions separately and EBT indicator. It is important to note that for calibration, the indicator should be added first, followed by ammonium hydroxide.

3) The hole is sealed with Teflon tape which has been cleaned with a drop of supra-pure nitric acid and deionized distilled water (DDW).

4) 10 ml of 0.1N HCl is added to the column, followed by 3 ml of 0.25N HCl and discarded

5) The Nd is collected with 3 ml of 0.25N HCl and the eluted sample is dried for use in the mass spectrometry lab.

6) The columns are washed with 2 ml of 0.25N followed by 1 ml of 0.5N HCl.

7) Sm is eluted with 3 ml of 0.5N HCl and is taken to dryness in clean small Teflon dishes for mass spectrometry.

8) The column is washed with 250 ml of 6.2N HCl followed by 150 ml of 0.25N HCl.

SUMMARY

Load Sample	0.5 ml	0.25N HCl	
wash	0.3 ml	0.25N HCl	
wash	10 ml	0.10N HCl	
wash	3 ml	0.25N HCl	
collect	3 ml	0.25N HCl	Nd
wash	2 ml	0.25N HCl	
wash	1 ml	0.50N HCl	
collect	3 ml	0.50N HCl	Sm

MASS SPECTROMETRY

In the mass spectrometry lab, the sample is put into solution with 0.1 ml of high purity DDW. This solution is loaded on the outgassed filament for analysis. Sr and Rb samples are loaded onto a single tantalum filament on which one drop of high purity phosphoric acid has been heated to dryness; when the sample has been loaded the filament is slowly heated to dryness and then quickly heated to a dull red. Nd samples are loaded onto a double rhenium filament and slowly heated to dryness.

The Sr and Nd isotopic analyses were determined using a multi-collector Finnigan MAT-261 mass spectrometer, operating in the static mode. Masses ^{86}Sr , ^{87}Sr and ^{88}Sr or masses ^{143}Nd , ^{144}Nd and ^{146}Nd were measured simultaneously on three different channels, and the observed intensities were corrected for the gain of the amplifier and the baseline readings.

Correlation coefficients for the Birunga samples:

[illegible]

	ZM	MI	KRB	R3SR	R3SRAT	Y	NB
T102	1.0000						
P205	1.0000						
MNO	1.0000						
S	1.0000	1.0000					
BA	1.0000	1.0000					
CR	1.0000	1.0000					
ZR	1.0000	1.0000					
SR	1.0000	1.0000					
RB	1.0000	1.0000					
Y	1.0000	1.0000					
NB	1.0000	1.0000					
KRB	1.0000	1.0000					
R3SR	1.0000	1.0000					
R3SRAT	1.0000	1.0000					
ALKAL	1.0000	1.0000					
KTAT10	1.0000	1.0000					
VFS102	1.0000	1.0000					
VFA1203	1.0000	1.0000					
VFF2031	1.0000	1.0000					
VFC20	1.0000	1.0000					
VFN20	1.0000	1.0000					
VFT102	1.0000	1.0000					
VFP205	1.0000	1.0000					
K	1.0000	1.0000					
IVSR	1.0000	1.0000					
VFF203	1.0000	1.0000					
VFF20	1.0000	1.0000					
T102	1.0000						
P205	1.0000						
MNO	1.0000						
S	1.0000						
BA	1.0000						
CR	1.0000						
ZR	1.0000						
SR	1.0000						
RB	1.0000						
Y	1.0000						
NB	1.0000						
KRB	1.0000						
R3SR	1.0000						
R3SRAT	1.0000						
ALKAL	1.0000						
KTAT10	1.0000						
VFS102	1.0000						
VFA1203	1.0000						
VFF2031	1.0000						
VFC20	1.0000						
VFN20	1.0000						
VFT102	1.0000						
VFP205	1.0000						
K	1.0000						
IVSR	1.0000						
VFF203	1.0000						
VFF20	1.0000						

END

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