The seismicity of east Africa

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THE SEISMICITY OF EAST AFRICA

by

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A THESIS SUBMITTED TO THE UNIVERSITY OF DURHAM
FOR THE DEGREE OF MASTER OF SCIENCE

February
1972

SUMMARY

A preliminary investigation of seismic recordings made by the Durham University array station at Kaptagat, Kenya, for the period July 1970 until May 1971 has been made.

A map has been produced indicating the epicentral location of earthquakes recorded within East Africa during this period. The work has shown that the majority of earthquakes recorded occurred associated with the major tectonic features of the East African rifts. The areas from which the largest numbers of earthquakes have been recorded were the western rift and the Kavirondo rift.

An empirical preliminary magnitude scale has been devised for earthquakes recorded at Kaptagat. Magnitude frequency plots for areas within East Africa indicate a difference in the strength between the underlying rocks of the western and eastern rifts.

The western rift exhibits the large bursts of energy release normally associated with more elastic material failure of the crust. The eastern rift shows earthquake magnitude and frequency characteristics that are indicative of the lower maximum stresses possibly associated with semi plastic flow rather than more elastic failure. This is taken as confirmation that the western rift has a structure possibly close to normal crust with significant spreading having not yet occurred. The eastern rift on the other hand, is seen as a zone of more plastic material failure and crustal spreading, identifying it as a plate boundary.

ACKNOWLEDGEMENTS

The author wishes to express his thanks to both Professor M.H.P.Bott and Professor G.M. Brown for granting him the privilege of working in, and the facilities of, the Department of Geology.

He wishes to thank Dr. R.E. Long for his valuable suggestions throughout the period of the study.

Grateful thanks are also due to:-

R.W. Backhouse and P.K.H. Maguire who helped with the computer analysis of events.

R. Birch whose work helped in the identification of earthquake arrivals.

J. Kipsang whose loyalty and hard work within East Africa greatly facilitated the recording of high standard data.

Dr. R.W. Girdler, J.D. Fairhead and B. Darracott for their helpful discussions.

Finally the author wishes to thank his wife, Jean, whose constant encouragement has made this work possible.

TABLE OF CONTENTS

Page

LIST OF TEXT FIGURES

CHAPTER ONE

INTRODUCTION

1.1. International recommendations

In 1963 a UNESCO geophysical survey mission to Africa recommended the immediate installation of a network of seismic recording stations within East Africa to enable a detailed study of the seismicity associated with the East African rift valleys to be made.

Bellusov (1964) expected that a comprehensive survey of the East African rift faults would provide much material for the highly controversial problem of the origin of continents and oceans. Most authorities currently consider the location of the world rift system to be critical to the understanding of continental drift, and the African rift system is thought by some as an early stage in the breakup of that continent.

Heazen, in his opening remarks to the UNESCO seminar in Nairobi, Kenya, in 1965, delineated the world rift system as stretching from the Red Sea to the Gulf of Aden on to the Indian Ocean where it bifurcates westwards through the Atlantic and Arctic Oceans to Siberia, and eastwards through the Pacific Ocean then following the western margins of the North American continent. He felt that as the East African rift was a land extension of this predominantly marine feature, the detailed geophysical investigations of this area would help the understanding of the mechanism causing the phenomena of rifting. A more specific recommendation at this seminar was that a detailed regional seismic investigation should be undertaken at the earliest opportunity.

In accordance with this recommendation Durham University installed three portable, semi-permanent, seismic array stations in East Africa during the period 1968-1971.

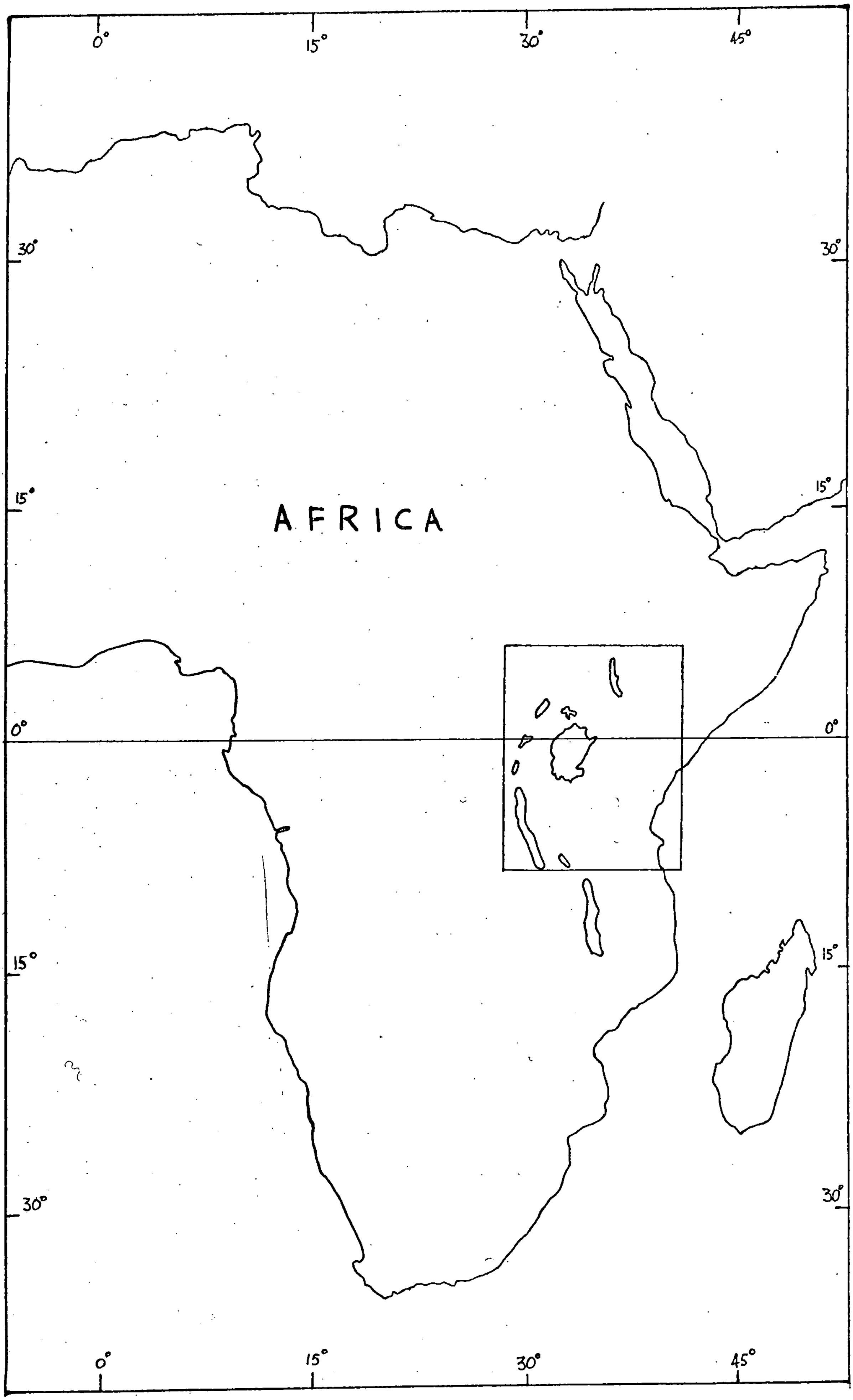


Fig. 1

The three array stations were situated at:-

Kaptagat, Kenya	0.4545°N	35.4612°E
Kakumiro, Uganda	0.787°N	31.328°E
Murchison Falls, Uganda	2.3°N	31.6°E

1.2. Geological Structure of the region

The area under consideration is bounded by the following geographical coordinates:-

Latitude 4°N to 9°S

Longitude 28°E to 42°E

and is shown in figure 1. The major tectonic features are the segments of the East African rift system (see figure 2), which can be traced from the Red Sea through Ethiopia, Uganda, Kenya and Tanzania and as far south as the lower Zambezi valley.

The western branch of the rift appears to originate north of Lake Albert and on its southward path manifests itself as the Ruwenzori mountain massif continuing southwards as the Lake Tanganyika depression, at the south end of which it joins the eastern rift a short distance north of Lake Malawi.

The eastern rift follows a more north-south path which in places exhibits more classic graben structures than does the western sector.

The rift valleys exhibit considerable variation in structure throughout both their length and breadth and range from well defined, relatively simple grabens such as the Kavirondo Gulf rift and the Lake Albert rift to the complicated pattern of block faulting and tilting found in eastern Tanzania.

The trend of the rift in Kenya appears to have been controlled to a great extent by the pre-existing mainly pre-Cambrian structural

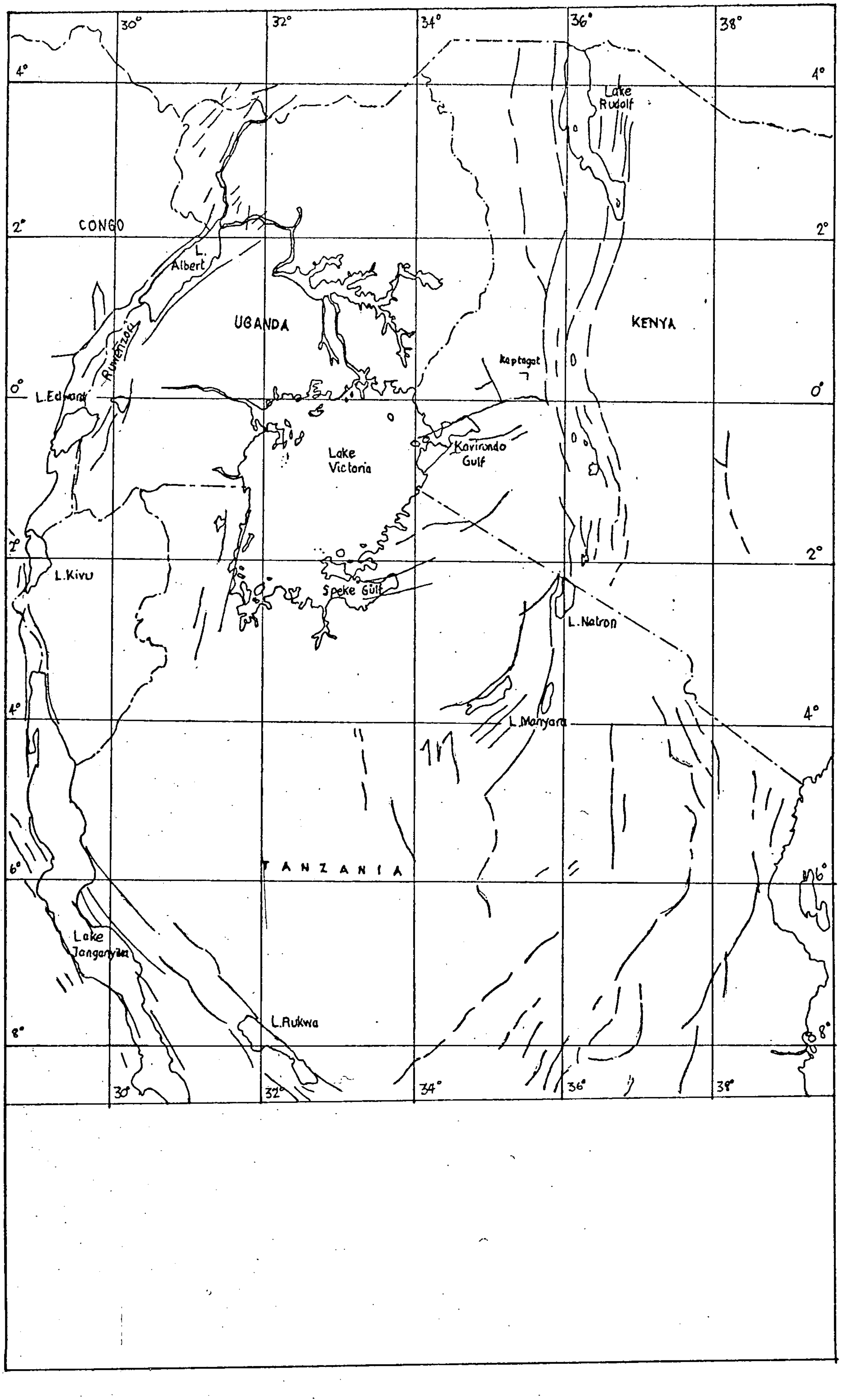

30°
32°
34°
36°
38°
4°
2°
0°
6°
8°
CONGO
L. Albert
Ruwenzori
UGANDA
KENYA
Lake Rudolf
Kaptagat
L. Edward
Lake Victoria
Kavirondo Gulf
L. Kivu
Speke Gulf
L. Natron
L. Manyara
TANZANIA
Lake Tanganyika
L. Rukwa

Fig. 2

features. The eastern rift follows the main north-south orientation associated with the Mozambique belt. The Kavirondo Gulf and the Speke Gulf alignment closely parallels that of the Dodomian and Nyanzian systems. The relationship between the pre-Cambrian basement structure and the rift trends was noted in southern Tanzania by Maconnel (1951). He noted that the structures in the pre-existing Ubendian and pre-Cambrian mylonite zones were seen to control closely the Lake Tanganyika - Lake Rukwa - Lake Malawi rift system. The Musoma and Eyasi rifts follow the general trends of the Mozambequian system.

In northern Tanzania, where there is no obvious graben structure, James (1956) suggested block and tilt faulting rather than through faulting. The horsts of Uluguru, Usambara and the Pare mountains, and grabens such as that which extends from the east of Mahenge northwards to north west Morogoro in which over 3000 M of Karoo sediments had been recorded had resulted from the convergence of parallel sets of opposed faults.

The geological evidence to account for the depression associated with Lake Tanganyika appears not to have been extensively worked upon but in contrast to the Rukwa - Malawi graben has no karoo or cretaceous sediments. It was therefore thought to be a post cretaceous feature due wholly to trough faulting of either Tertiary or Quaternary age (Dundas, 1965).

Three main phases of epeirogeny and crustal flexuring have been identified by Baker and Wohlenberg (1971) as associated with the eastern rift.

In the Miocene the eastern rift passed through a period of broad crustal flexing and local faulting accompanied by basic volcanism.

This was succeeded in the late Miocene by voluminous flood phonolite eruptions.

The first faulting to affect the whole length of the rift took place in the early Pliocene creating an asymmetrical graben. This was followed by dominantly basaltic volcanism in the floor of the rift depression. Towards the end of the Pliocene the last and major uplift of the plateaux began accompanied by massive trachytic volcanism and phases of graben faulting which lasted until the middle Pleistocene.

1.3. Previous Geophysical Studies

The frequent occurrence of earthquakes in East Africa prompted the writings of many of the early travellers and explorers but the difficulties of travelling without accurate maps resulted in the eroneous appraisal of the seismicity within the region.

In 1906, De Balladore reported isolated tremors in East Africa but attached no seismic importance to them in his map of the seismicity of Africa. Gregory (1921), after whom the Kenya section of the eastern rift valley was named, considered the East African coast to be one of the most seismic areas of the world.

Krenkel in 1921-2, however, established the first positive correlation between the rift valleys and earthquake occurrence by collecting reports of felt earthquakes and producing a map indicating event frequencies. A map delineating the seismic zones of Africa was produced by Sieburg (1932) who was the first worker to indicate the activity of the Kavirondo Gulf as well as the other rift zones. Gorshkov in 1963 published a summary of all previous reviews of seismicity of East Africa which showed not only a good correlation between obvious rift structures and earthquake locations but also a wide spread of epicentre locations in non rift areas.

Gutenberg and Richter (1949, 1954) were the first workers to use instrumental data. Their work listed events from East Africa ranging in magnitude from M=6.0 to M=7.7 but the paucity of stations used left the epicentral locations much in doubt. Sykes and Landisman (1964) give additional epicentral location information covering the period January 1955 until March 1964 using data from the then newly established World Wide Standard Seismological Station network (W.W.S.S.S.). They concluded that although a large number of earthquakes in East Africa were associated with the various branches of the rift system, though predominantly the western branch, many well recorded earthquakes were not located within the rift valleys. They suggested that compared with the narrow linear pattern of seismic activity associated with the mid-oceanic ridge system, East Africa exhibited seismicity of a relatively widespread areal extent.

The western rift was studied by Sutton and Berg (1958) who attempted fault plane solutions and by De Braemacher (1964) who worked predominantly on events located to the west of the western rift. He found some correlation between the epicentres and structure within the Congo. Both parties used data from the I.R.S.A.C. network of seismological stations situated within the Congo.

Wohlenberg (1968, 1970) gave event locations for earthquakes with magnitudes greater than M=4 occurring within East Africa during the period 1957-1964 and studied the focal depth of some western rift earthquakes concluding that the area was one exhibiting predominantly shallow focal depths. Epicentres were located to a maximum focal depth of 55 Km with maxima within their distribution at approximately 15 Km and 27 Km. Positive correlation was found between the areas of highest seismicity and those of strongest negative Bouguer anomaly, particularly areas with the steepest gradients of this anomaly. The strongest gravity anomalies were observed in the Lake Albert region and near to the Rukwa rift as well as in an area to the east and south east of Lake Victoria. An extremely steep gradient of this anomaly was observed in the northern part of Lake Tanganyika. The large negative Bouguer anomalies associated with the Semliki - Lake Albert and the Rukwa rifts were seen to be associated with the deep sedimentary troughs in those areas.

Fairhead (1968) studied the accuracy of the epicentre location of events of magnitude Mb=4.8 and greater within East Africa. He used the method of Joint Epicentre Determination (J.E.D.), described by Douglas (1967) and Lilwall and Douglas (1970) which involves the estimation of both source and station corrections and regional variations in the travel-time curve as well as the best estimates of the parameters of earthquakes. In order to solve the equations of conditions, it is necessary to restrain one event in the region of study. The epicentres for a group of earthquakes within that region, together with the travel time corrections could then be estimated. In this and later work Fairhead and Girdler, (in press) found good correlation between the relocated events and geological features in areas where detailed mapping data was available. In particular a relocation of the events that formed part of the aftershock sequence following the Toro event of 20 March 1966 was undertaken with the redistribution showing that instead of occurring around the main event the epicentres were located between the Kitimbi — Semliki fault and the western wall of the main rift valley. Correlation was presumed between the events and movement of the western fault escarpment, the main event having triggered adjustment along this fault zone. Girdler et.al. (1969) indicated good correlation between the location of teleseismically recorded earthquakes and faults mapped as having been active during the period Pleistocene-Recent. Gravity data was used to support their attenuation theories. Short wavelength positive Bouguer anomalies over the rift floor – thought to be indicative of the presence of an intrusive zone under the axis of the rift - and long wavelength negative anomalies were interpreted in terms of thinning of the lithospheric plate.

The micro-seismic activity associated with parts of the rift system within East Africa has also been studied by Tobin et.al. (1969) and Mclnar and Aggerwal (1971). Portable equipment located at many sites within East Africa was used for periods seldom exceeding a few days at any site. Their conclusions indicated that there was, during the periods of study little or no activity occurring outside the accepted rift valley structures and that the activity was mainly concentrated within the Kavirondo Gulf rift and the eastern rift south of Lake Baringo with a relatively high level of activity around Lake Magadi in southern Kenya and northern Tanzania.

Gumper and Pomeroy (1970) derived a mean crustal model for Africa from the dispersion of Rayleigh waves along paths outside the rift zone. This model, referred to as the AFRIC model, is a minor modification of the CANSD model for the Canadian Shield (Brune and Dorman, 1963) and indicates that, away from the immediate rift zone, Africa has a structure typical of the stable shield areas of the world.

In contrast a refraction line in the Gregory rift reported by Griffiths et.al. (1971) shows a 20 Km thick layer of 6.4 Km/s overlying a 7.5 Km/s layer presumed to be anomalous mantle material. Such an anomaly is to be expected from the failure of Sn to propagate across the rift zone north of the Equator, which is taken by Gumper and Pomeroy to indicate 'a gap in the lithosphere'.

AFRIC MODEL PARAMETERS

Figure 6

H(km)	Vp (km/sec)	Vs (km/sec)	ρ (gm/cc)
	CRUST		
7.0	5.90	3.35	2.70
10.5	6.15	3.55	2.80
18.7	6.60	3.72	2.85
	MANTLE		
80.0	8.05	4.63	3.30
100.0	8.20	4.78	3.44
100.0	8.30	4.65	3.53
80.0	8.70	4.85	3.70
	9.20	5.25	3.76

Sundaralingam (1971) measured delays in arrival times at Addis Ababa, Nairobi and Lwiro with respect to Bulawayo using events in the distance range 25° to 90° and found, using Herrin's tables (Herrin et.al., 1968) no significant variation with azimuth or epicentral distance. Evidence from these surface wave dispersion studies indicates that Bulawayo lies on shield crust and mantle typical of Africa as a whole. The relative delays between these stations were taken as a direct measure of the difference between the upper mantle beneath the rift stations and the shield structure of Africa. The delays were concluded as evidence of a substantial low velocity zone in the upper mantle beneath the eastern branch of the rift. Results with respect to Lwiro indicated that a reduced amount of anomalous material may occur beneath the western rift.

Long et.al. (in press) note that the similar delays calculated for the eastern rift stations at Nairobi and Addis Ababa are close to the large relative upper mantle delays calculated for Iceland relative to the Swedish shield (Long and Mitchell, 1970), and imply that as the mean compressional velocities are similar beneath the eastern rift and Iceland they are therefore presumably similar to that beneath the mid-Atlantic Ridge.

Long et.al. (in press) using array data from Kaptagat, Kenya found that evidence for normal crust and topmost mantle between the two branches of the rift system was given by velocities of 7.9 Km/s associated with the propagation of Sn from regional earthquakes located west of the eastern rift. Events located to the east of the western marginal fault of the eastern rift, however, were noted as having apparent velocities of 7.1 Km/s. This velocity, taking into account local structural complexity, indicated that the wavepath of these events had passed through the anomalous structure along the axis of the eastern rift that has become evident from both refraction data (Griffiths et.al., 1971) and gravity data (Khan and Mansfield, 1971). The array data was seen to provide confirmation of the crustal modification along the eastern rift.

1.4. Explanation of the Gregory Rift

Many attempts have been made to explain the formation of the Gregory rift: Willis (1936) proposed crustal doming, Girdler (1964) inferred isostatic subsidence of a block and later, Girdler et al (1969) crustal stretching and thinning, whereas McKenzie et.al. (1970) and Searle (1970) suggested it as a zone of crustal rupture and igneous injection.

Freund (1966) considered the depth of the graben too great for the doming or isostatic models to be adequate. If the fault planes had a mean dip of 63^{o} there must have been at least 5 Km of crustal extension to account for the observed structures (Baker and Wohlenberg, 1971). They suggest a total crustal extension of up to 10 Km taking into account the possible existence of faults concealed under the graben floor. This value is much in excess of one that could have resulted from crustal arching (Evans, 1925). Baker and Wohlenberg (1971) suggest that the doming and faulting observed must be related results of some sub-crustal process.

A maximum equatorial crustal separation of 20-30 Km, the width of the inner graben of the Gregory rift is inferred from coastline fitting of the Gulf of Aden (McKenzie et.al., 1970). This is equivalent to the whole width of the inner Gregory rift as Pre-Cambrian rocks are locally exposed on either side of the inner graben. At the northern and southern extremities of the Kenya domal uplift Pre-Cambrian rocks outcrop across the whole width of the rift-zone broken only by normal faulting, allowing a maximum crustal extension of 3 Km in those areas. The absence of cross rift faults of transform type is taken by Baker and Wohlenberg (1971) to indicate a crustal extension across the central rift of nearer 5 Km than 25 Km.

From gravity measurements made throughout the Gregory rift area it can be seen that the principal anomalies are linear and follow the north-south trend of the rift. The negative anomaly associated with this rift has a superimposed axial positive high with an amplitude of between 30 and 50 mgal. No similar positive anomaly was recorded from the Kavirondo rift.

Baker and Wohlenberg (1971) have computed a crustal model from this gravity data which allows about 10 Km of crustal separation under the central part of the Gregory graben. They suggest that the crustal separation took place concurrently with basic igneous injection and basalt eruptions that flooded much of the rift floor in mid-Pliocene or Miocene times. Much of the evidence for this was however concealed by the massive late Pliocene and Quaternary sialic volcanism of the central rift (Williams, 1969).

Recent modifications (Mohr, 1970; Freund, 1970) to the crustal separation model, suggested by plate tectonics, (McKenzie et.al., 1970) which bringing the Arabia - Nubia and Arabia - Somalia rotation poles closer together imply a crustal separation of nearer 10 Km than the previously proposed 30 Km.

1.5. Significance of previous studies

All previous work has pointed to the conclusion that the eastern rift is the site of very limited crustal separation but that it is a plate boundary. Plate boundaries throughout the world are normally associated with a high level of seismic activity. All workers, and more notably Wohlenberg (1970) have found that the eastern rift exhibited a very low level of activity.

The Kenya array station was placed at Kaptagat, 10 Km from the western wall of the eastern rift to enable a more detailed study of the area to be undertaken.

CHAPTER TWO

2.1. Location of Recording Stations

The three array stations were situated at:-

	Latitude	Longitude
Kaptagat, Kenya	0.4545°N	35.4612°E
Kakumino, Uganda	0.787°N	31.328°E
Murchison Falls, Uganda	2.3°N	31.6°E

The first two were ten instrument L-shaped arrays, with five sites at intervals of 1 Km, on each arm of the array (see figure 4). The third, at Murchison Falls, consisted of three radio linked short period instruments situated at the apeces of an approximately equilateral triangle of side 1 Km, for the Uganda Electricity Board.

In the installation of the two ten-instrument arrays, an attempt was made, as far as the local geology permitted, to align the arms respectively north-south and east-west. The north-south alignment was made to ensure the fullest analytical use of those events which had a travel path along the rift and the east-west alignment to look at events traversing the two rifts and the area between them.

The three instruments at Murchison Falls straddled the site of the then proposed hydro-electric scheme on the River Nile.

Although recordings are available from all three stations only data from Kaptagat has been used in this study.

2.2. Geology of the Kaptagat Station area

The area is characterised by the gently westward dipping Uasin Gishu phonolite lavas. The north-south arm of the array is underlain by the Upper Phonolite and the east-west arm by the Lower Phonolite. The whole array is situated at the boundary between these two lava flows.

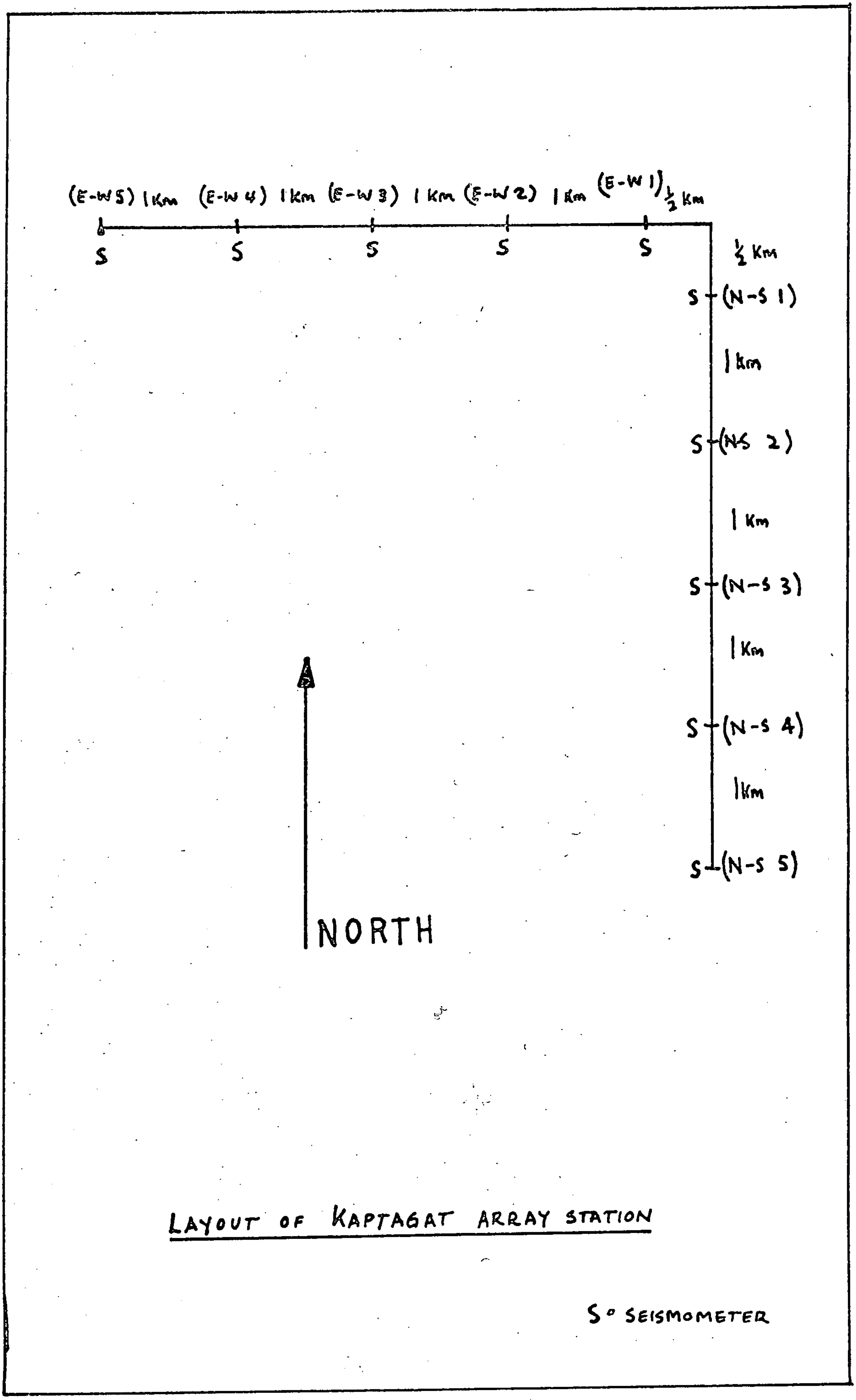

(E-W 5) 1 Km (E-W 4) 1 Km (E-W 3) 1 Km (E-W 2) 1 Km (E-W 1) ½ Km
S
S
S
S
S
½ Km
S (N-S 1)
1 Km
S (N-S 2)
1 Km
S (N-S 3)
1 Km
S (N-S 4)
1 Km
S (N-S 5)
NORTH
LAYOUT OF KAPTAGAT ARRAY STATION
S = SEISMOMETER

Fig. 4

The lower of the two phonolites is a sparsely porphyritic lava whereas the upper contains abundant large Nephelines and glassy Feldspar phenocrysts, easily distinguishable in hand specimen. The basal flow forms the greater part of the flat land of the plateau while the later flow overlies it in the east and north forming a pronounced, near continuous erosion scarp at the junction.

The lower flow apparently lies directly on Basement System gneisses in this area, though the presence of a thin intervening layer of pyroclastic material, at least locally, is indicated by the occasional blocks of tuff preserved near the Kip Karren waterfall. To the south of the station the phonolite everywhere rests directly on the lower volcanic rocks of Tinderet (Jennings, 1964).

The Basement System gneisses of the area display a regional strike generally NNW-SSE with steep dips to the north east. The rocks have been sufficiently mobilised locally to destroy the original gneissose foliation, such fabrics as occur being the result of limited plastic flow.

The area is structurally an upslanding block, bounded to east, south and west by major faults. The Elgeyo scarp runs sub-parallel to the north south line at a distance of approximately ten kilometres. There is no indication of subsidiary faulting of this trend within the area of the station.

The widespread, uniform, gently dipping nature of the plateau phonolites suggests a fissure source rather than a central cone and the general southwest trend of the contours on the lava surface indicates that such a fissure would probably run on the Elgeyo escarpment fault trend.

The major structure to the west is known as the Nandi fault. It runs NNW-SSE and has been traced from its emergence beneath the volcanic rocks of Mt. Elgon (Gibson, 1954) to its disappearance into the South Nandi Forest (Huddleston, 1954) and again from its crossing of the Nyando Scarp to its disappearance beneath the alluvium of the rift valley plain (Shackleton, 1950; Binge, 1962). Over most of this distance south of Broderick Falls a pronounced west-facing scarp, resulting from recent rejuvenation of the fault, indicates its line, but the feature dies out southwards so that the point of entry of the fault into the area is in some doubt. Shackleton (1951) indicated the south end of the Nandi fault as cutting the Nyando escarpment, and confining Nyanzian meta-volcanic rocks to the west, with Basement System gneisses on the east. More detailed mapping has shown (Jennings, 1964) that several faults of NNW-SSE trend across the scarp and the country north of it.

A major fault zone runs along the western half of the southern boundary of the area approximately 45 Km south of the station and is the continuation of the Nyando fault which forms the northern boundary of the Kavirondo Gulf rift valley.

2.3. Method of Recording

(a) Use of Arrays

Comparison has been made between array and multi-station analysis of an event by Cleary (1967) who found array data to be superior for the purposes of event recognition, focal depth determination and velocity and azimuth calculations.

Detailed accounts of arrays have been published by Birtill and Whiteway (1965) who worked on the linear type favoured and used in the United Kingdom and Backus et.al (1964) who together with Burg (1964) worked on the cluster type more favoured in America.

Cluster arrays consist of many instruments arranged in some regular pattern over a wide area. This type of array enables a reduction in the level of ambient noise by the filtering of all events with a given velocity.

Linear arrays normally consist of two mutually perpendicular lines of instruments. This configuration allows phasing of the signals from the various instruments so that the array can be turned to look in a given direction. Surface velocity is measured from the relative times at which the signal appears at the various seismometers and azimuth of the arrival is deduced from the order in which the arrival appears on the various instruments. The distance between the instruments at an array has to be smaller than the wavelength to be recorded.

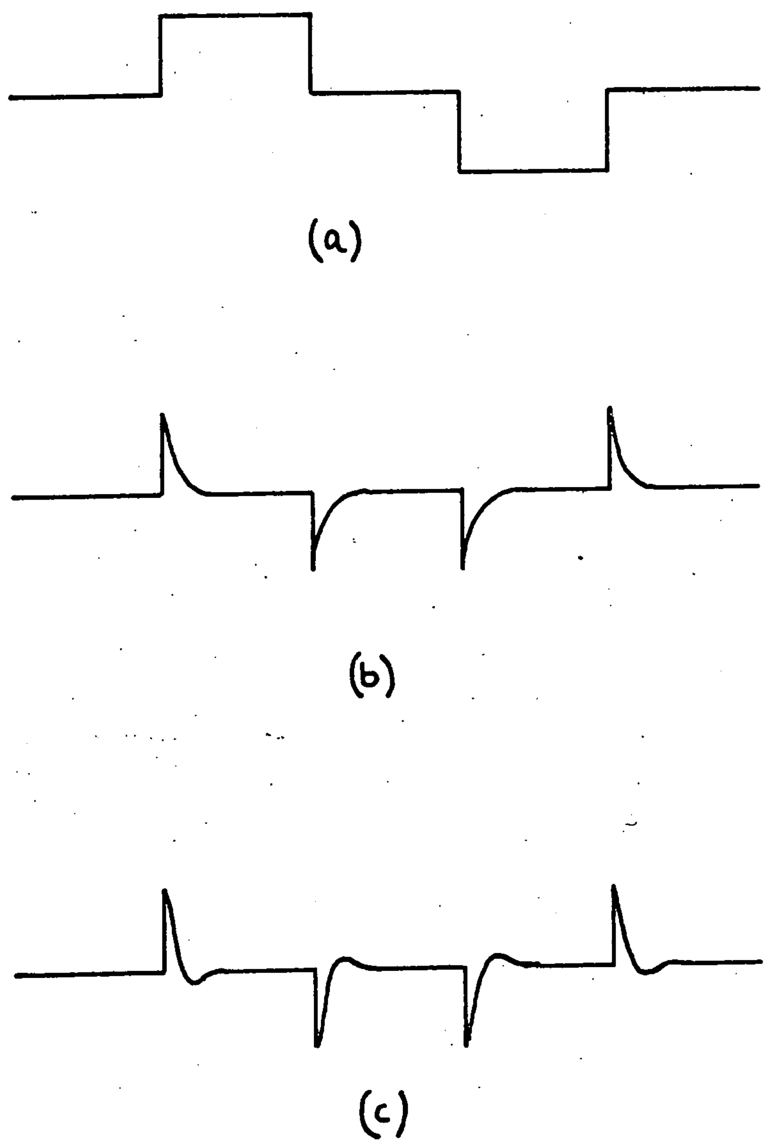
(b) Choice of Array

To obtain the maximum amount of useful data from East Africa arrays of the linear type, see figure 4, for the configuration, of Willmore MKII seismometers, set vertically to a period of 2 seconds, spaced at an interval of 1 Km were used. The design of the array allowed the recording of seismic waves in the frequency range 0.5 Hz to 20Hz which included teleseismic P-waves and all types of waves from regional events.

(c) Method of Recording

The signal, which represents the ground velocity at the seismometer was communicated to the central recording station, where it was recorded together with time from a quartz crystal clock onto multi-channel magnetic tape. The communication was by twin field telephone cable at Kaptagat.

Power was fed to the amplifier packages at the seismometers from the central recording station on the same pair of wires that



CALIBRATION CHARACTERISTICS

Fig. 5

carried the seismic signal. The whole station was powered by a single set of accumulators at the recording station. This method of communication allowed the remote, periodic calibration of the seismometer once again using the same pair of wires. Both the seismometer amplifier and recording electronics were described by Long (1968).

Signals from the seismometers were amplified and frequency modulated before being sent to the recording station. There was obvious need for remote calibration facilities to ensure that both the seismometers and seismic amplifiers were functioning correctly. The calibration was generated in the seismometer package by a remote calibration unit triggered by a pulse sent down the line from the recording station. The calibration sequence from the amplifier was so designed that four pulses were generated from a square wave form sequence (see figure 5a), and the display of these pulses from each seismometer package (figure 5b) ensured that the recording electronics were satisfactory.

To check that seismometers were functioning correctly a similar calibration technique was used. Figure 5c shows the characteristics of the seismometer that were returned in response to the square wave input. The output, which contains details of the characteristics of the whole direction system, was recorded on the seismic channel to which it referred and was therefore available at the time of subsequent playback. The amplitude of the pulses recorded was particularly necessary as a control for the event amplitude measurements required for magnitude determinations.

(d) Central Recording Station

A fourteen channel 'Geotech-Teledyne' recorder fitted with both recording and playback heads was used. Recordings on 1 inch wide magnetic tape using fourteen inch spools at a tape speed of 15/160 inches per second allowed an uninterrupted recording time of about eleven days.

A single channel hot wire 'Cambridge' pen recorder provided both display and monitoring facilities. This instrument could be driven either from the playback facilities of the Geotech recorder or direct from the land lines via a demodulator.

Binary encoded time was recorded on one channel from a high stability crystal oscillation clock and for absolute time comparative purposes GMT pips, transmitted by the B.B.C., were recorded on another.

(e) Power Supplies

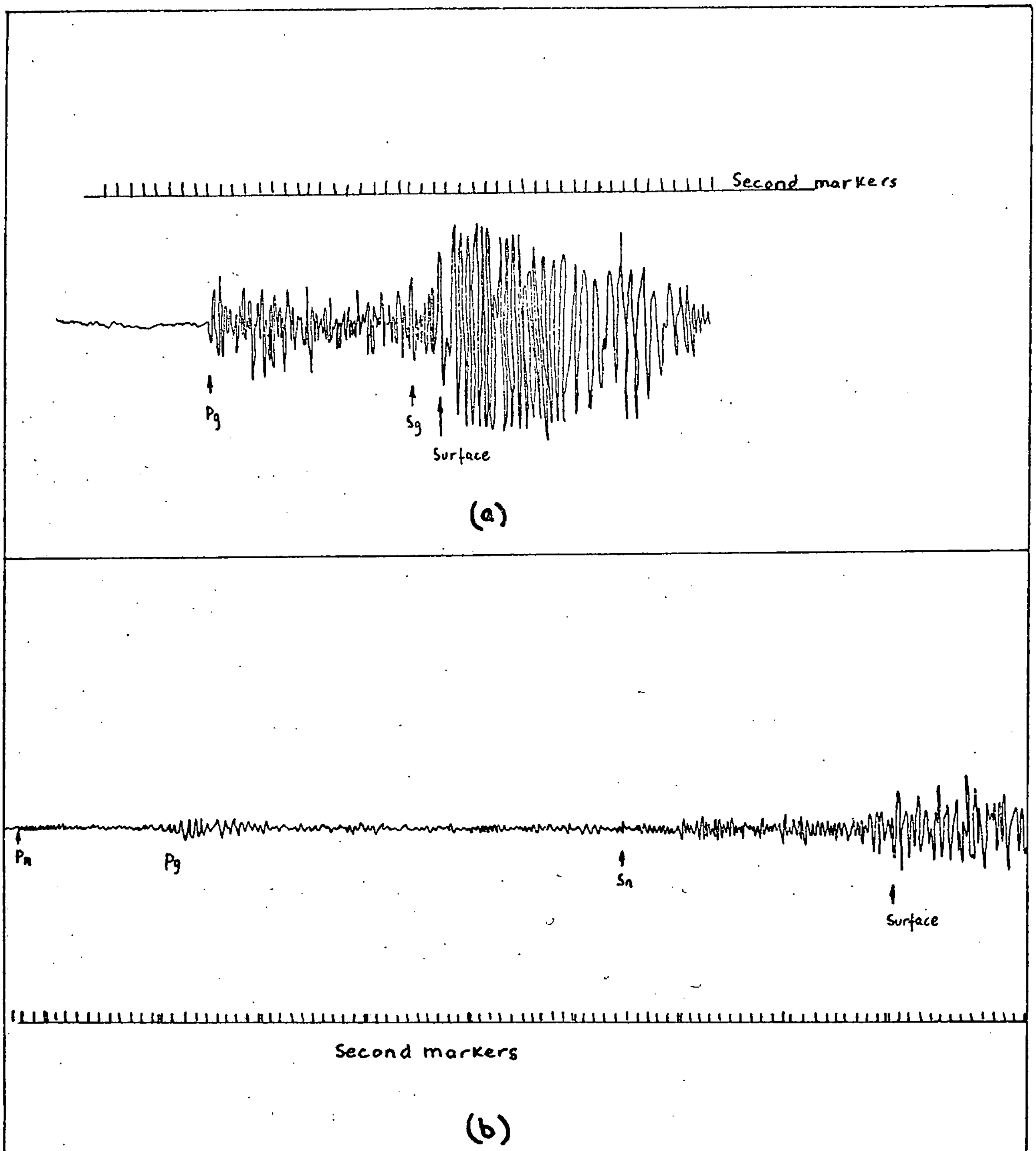
Power for the whole array station was taken from a single group of twelve, six-volt, accumulators located at the central recording station. It was necessary to charge these accumulators at least once per day if continuous recording were to be maintained.

2.4. Analysis of Recordings

(a) In East Africa

Although periodic calibrations and voltage checks were made to ensure that the equipment was functioning satisfactorily, during the duration of the tape, each tape was subsequently played out onto paper for more detailed quality checks and an initial appraisal of the seismic data recorded on it.

At Kaptagat the tapes were played back on an E.M.I. deck at a speed of 7.5 inches per second, eighty times the recording speed. This playback speed allowed the economic playout of the data onto paper on a six channel jet pen recorder with the encoded time track sufficiently readable to enable the determination of the onset time of each event with respect to G.M.T.



EXAMPLES OF ARRIVAL DATA PLAYED BACK AT KAPTAGAT.

(a) KAVIRONDO RIFT EVENT.

(b) WESTERN RIFT EVENT.

fig. 3.

The initial playout on paper allowed the different event types to be readily identified i.e. local, regional or teleseismic. The bulletins produced from this initial analysis together with the station log sheets and the individual event playouts were sent together with the magnetic tape to Durham.

The detailed checks on the tapes made both during recording and immediately afterwards allowed any system faults to be speedily diagnosed and rectified.

(b) Durham Playback Equipment

The playback equipment at Durham comprises E.M.I. decks with a speed of 15/16 inches per second, replay electronics to demodulate the signal and Krohn Hite band pass filters. Display can be effected either on paper, produced by a 16 channel jet pen recorder or on a 12 channel oscilloscope or both.

By means of a multichannel switching panel it is possible to display the recorded channels on the jet pen recorder, in any required order. Recordings, for ease of identification, were played back with respect to their geographical position as indicated in figure 7[illegible].

In order to maintain the correct timing across the playout time code is printed out on the two edge tracks of the paper allowing the arrival traces from the individual instruments to be measured with respect to G.M.T.

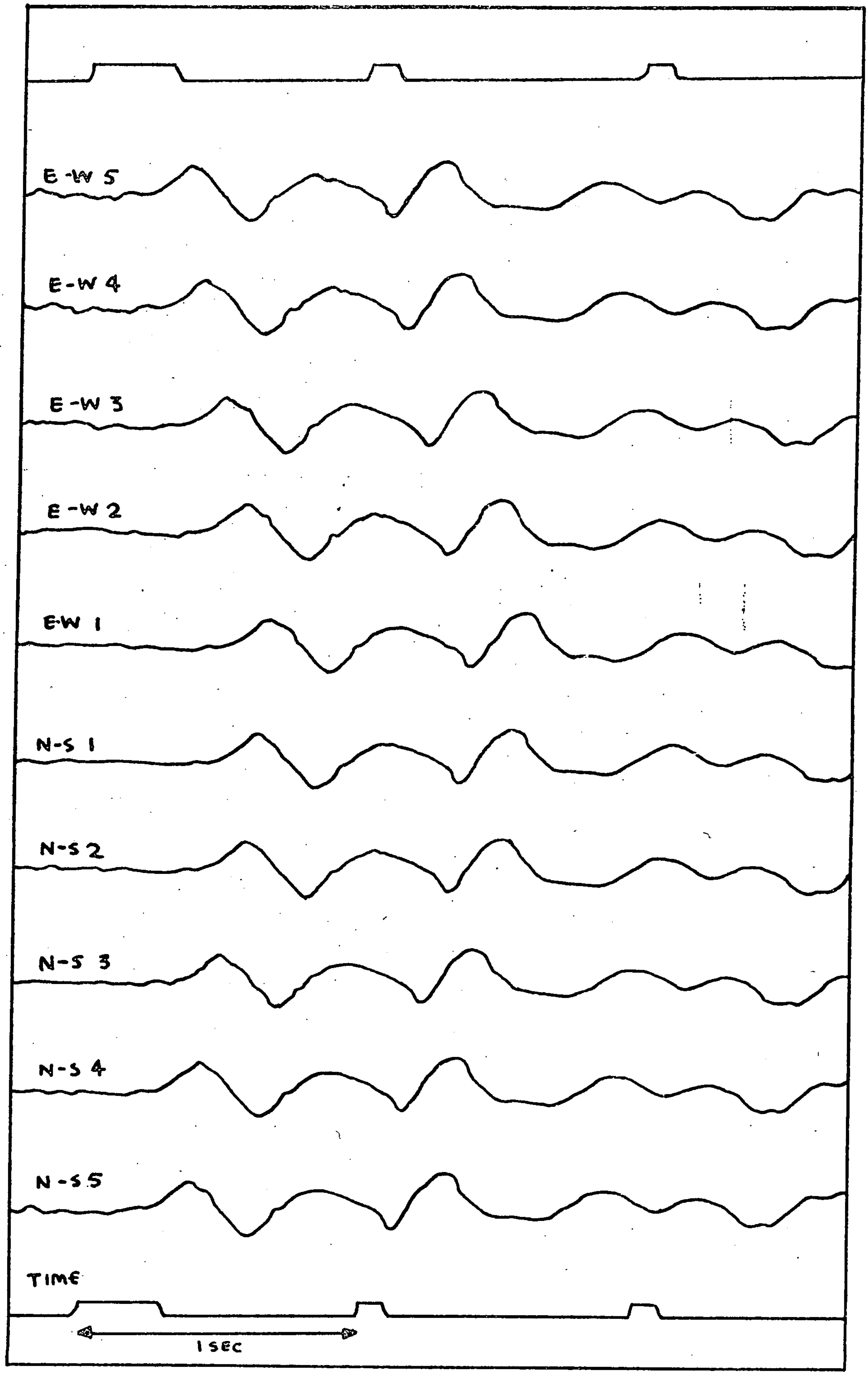


fig. 7.

CHAPTER THREE

EPICENTRAL DETERMINATION

3.1. Method

From the printout made at Durham as described in chapter two, it was possible to isolate the arrivals, representing the various wave forms i.e. P, S and surface wave, recorded at the seismometers functioning at the time of the event.

As many of the events investigated were of small magnitude, the first motion was frequently found to be indistinct due to the relatively high level of background noise on the recording, greater accuracy was therefore achieved by the consideration of the first three cycles of each wave form rather than the first motion.

Although only the first P-wave arrival for each event was normally investigated for velocity and azimuth purposes, it was found useful, in some cases where the first few cycles of the P-wave arrival were seen to be emergent from the background noise, to consider the S wave or surface wave arrivals provided that the phase concerned could be correctly identified.

In order to correlate the channels from the various seismometers a tracing of the best channel on the printout was made on transparent paper, with an arbitrary reference point marked on it, so that by superimposing this tracing on the other channels the same reference point could be marked on each channel.

The measurement of these reference points, relative to each other, was made from a straight line drawn between second markers on each side of the printout which represented the same period in time. Measurement was then made in millimetres from this line to the reference point marked on each channel. To enable this measured distance to represent time, the distance between second markers was measured. The measurement was generally made over a distance of five seconds and the average distance per second divided into the previous measurement for greater accuracy. By this method the stepout between the instruments could be measured to an accuracy of 0.01 sec.

(1) <u>Coordinates of Instrument Sites</u>

Although an initial ground survey had been carried out using tape and compass in conjunction with Ordnance Survey maps, the precise locations of these sites must remain with some element of doubt. The difficulty arises because the scale of the only maps available was about 1 inch to represent 1 mile, and at this scale an error in location of between 50 and 100 metres is not thought impossible. The dense undergrowth that exists over a great part of the area was hitherto precluded a detailed instrumental survey.

It is, however, hoped that any inaccuracies due to this problem will be shortly eliminated as an aerial survey of the area is proposed. It is not expected that this survey will substantially alter the coordinates of any instrument site as comparison of events recorded both at Kaptagat and other stations in the world have shown good azimuthal correlation.

(2) Data Processing

A computer programme that has been developed at the University of Durham by R.W. Backhouse, will give a printout of the surface velocity of the event at the recording station, the azimuth from the recording station of the epicentre, and provided that the P-wave to S wave or P—wave to surface wave time interval is given, the geographical coordinates of each event.

The process is one involving the best least squares fit between the arrival times and their respective instrument coordinates which enables a computation of the apparent velocity of the phase along each arm of the array.

If Vx and Vy are the apparent velocities recorded along the arms of the array, x and y, then ϕ, the angle of approach relative to the line x is given by the relationship:-

$$\phi = \sin^{-1} \frac{Vy}{Vx}$$

The azimuth of the epicentre from the station is measured as the angle between due north and the angle of approach, measured anti-clockwise.

From the measurement of the P—wave to S—wave time interval, using the following relationship, the epicentral distance can be evaluated:—

$$\text{Distance} = \text{Time (S wave} - \text{P wave} \left(\frac{1}{Vs} - \frac{1}{Vp} \right)^{-1}$$

where Vs and Vp are the respective S and P wave velocities.

In the absence of an identifiable S phase it was found necessary to evaluate the epicentral distance by using the time difference between the P wave arrival and the surface wave arrival. The relationship:-

$$\text{Distance} = \text{Time (Surface wave} - \text{P wave)}\left\{\frac{1}{V\ \text{surface}} - \frac{1}{Vp}\right\}^{-1}$$

was used where V surface was the surface wave velocity.

In order to locate earthquakes within East Africa using data from the Kaptagat array three different relationships were used according to the location of the event.

(a) <u>Events occurring within 180 Km of Kaptagat</u>

At epicentral distances from Kaptagat of less than 180 Km the first arrival was Pg which was followed by Sg just before the surface wave arrival. For these results mean upper crustal velocities of Pg = 6.1 Km/s and Sg = 3.5 Km/s were assumed. The velocity of 6.1 Km/s has been confirmed by Long et.al. (in press) by the velocities measured across the Kaptagat array from events located to the west of the station within this distance range.

Events located to the east of Kaptagat were recorded associated with a variety of velocities indicating that normal crustal conditions did not exist within that region.

It has been possible to establish the focal depth of events located within a distance of about 40 Km from Kaptagat by considering the Pg arrival velocity. The velocity observed across the array is related to the angle of emergence from the hypocentre.

The focal depth, h, has been calculated using the following relationship:-

$$h = x \sin (\cos^{-1} \frac{v}{V})$$

where x is the distance from the hypocentre to the recording station, v is the recorded velocity and V the mean crustal velocity. The distance, y, between the epicentre and the recording station is given by:-

$$y = x \frac{v}{V}$$

Focal depths, see figure 11, were initially calculated using a mean upper crustal velocity of 6.0 km/s and events which exhibited a focal depth greater than 17.5 Km, the base of the first layer in the AFRIC model, were then recalculated considering velocity of 6.6 Km/s for the lower crust.

From this data it is evident that at distances of less than about 40 Km from Kaptagat the epicentral distance is greatly affected by the focal depth, but at greater distance this is not significant.

The accuracy of focal depth determination, with a maximum reading error of Pg across the array was ± 2 Km at a focal depth of 20 Km.

(b) <u>Large amplitude events located at a distance greater than 180 km from Kaptagat</u>

Events located west of Kaptagat at distances of greater than 180 Km were recorded with Pn as the first arrival. Very great

Focal depth calculations for events near to Kaptagat

Event	Azimuth (^{o})	Distance Km	Epicentral Distance Km	Focal Depth Km	Velocity Km/s
53-1870	164^{o}	28.2	12.1	25.6	15.95
53-1884	73^{o}	38.5	22.0	31.6	11.53
54-1926	124^{o}	30.1	23.8	18.4	8.35
56-2125	177^{o}	44.0	40.4	17.5	7.19
38-1173	177^{o}	45.0	42.4	15.7	7.0
38-1125	280^{o}	32.0	28.6	14.3	7.38

Figure 11

difficulty was experienced in identifying the Pg onset and for this reason it has not been possible to assign focal depth to earthquakes from this area.

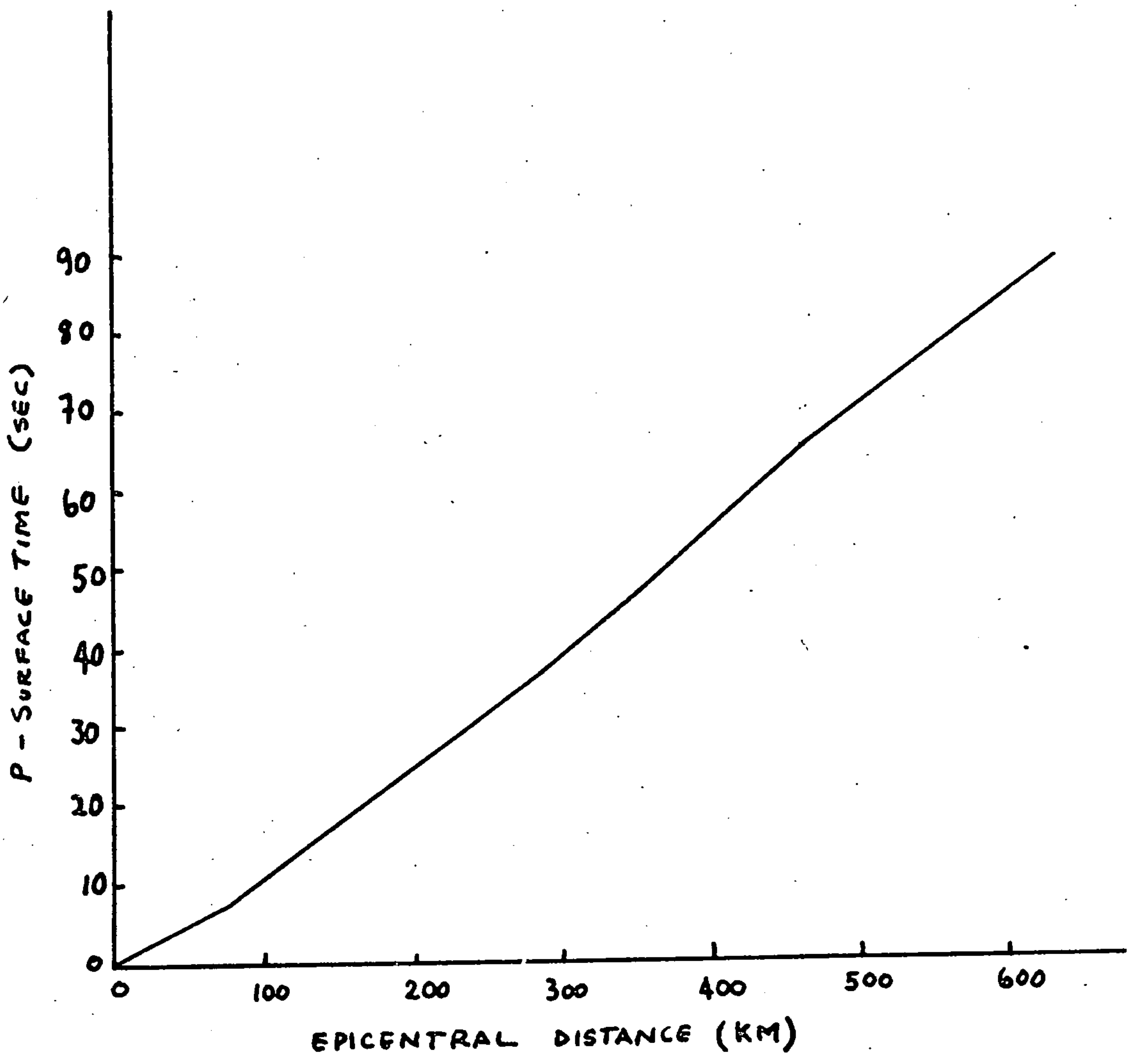
Velocity filtering of events is proposed by other workers at Durham but at the time of this work had not been undertaken. It is hoped that such filtering will allow phases to be much more easily identified thereby enabling focal depth determinations to be made. Another difficulty in focal depth determination is the lack of data on the crustal thickness beneath East Africa.

The epicentral distance of these events was calculated using arrival time data for the phases Pn and Sn. A velocity of 7.9 Km/s confirmed by Long et.al. (in press) was used for Pn. Sn was assigned the corresponding velocity of 4.5 Km/s.

(c) Small amplitude events located at a distance greater than 180 Km from Kaptagat

Small magnitude events located to the west of Kaptagat at a distance greater than 180 Km were frequently recorded without any discernable S wave onset. For these events it was necessary to consider the relative onset times for Pn and the surface waves. The surface waves recorded were Rayleigh waves since Love waves have no vertical component and only vertical seismometers were used at Kaptagat.

The arrival time relationship between P, S and Surface waves has been noted and a graph drawn, see figure 10, showing the P wave to surface wave time in seconds against the distance from the recording station calculated using the P-S wave relationship



PWAVE - SURFACE WAVE TIME AGAINST DISTANCE PLOT

(USED FOR EVENTS WITHOUT CLEAR SWAVE ONSET)

fig. 10

described earlier. This calculation was made for events where all three phases were readily identifiable. This has been confirmed by the distances calculated from the U.S.G.S. location of earthquakes recorded at Kaptagat.

It was, therefore, possible knowing only the P wave arrival time and the surface wave arrival time for the other events to assign an epicentral distance.

(3) Azimuth measurements

Unlike epicentral distance, azimuth was not dependent upon the identification of a specific phase within the arrival pattern. Provided that it was possible to identify the same phase across the array then the azimuth of the event to the recording station was independent of velocity.

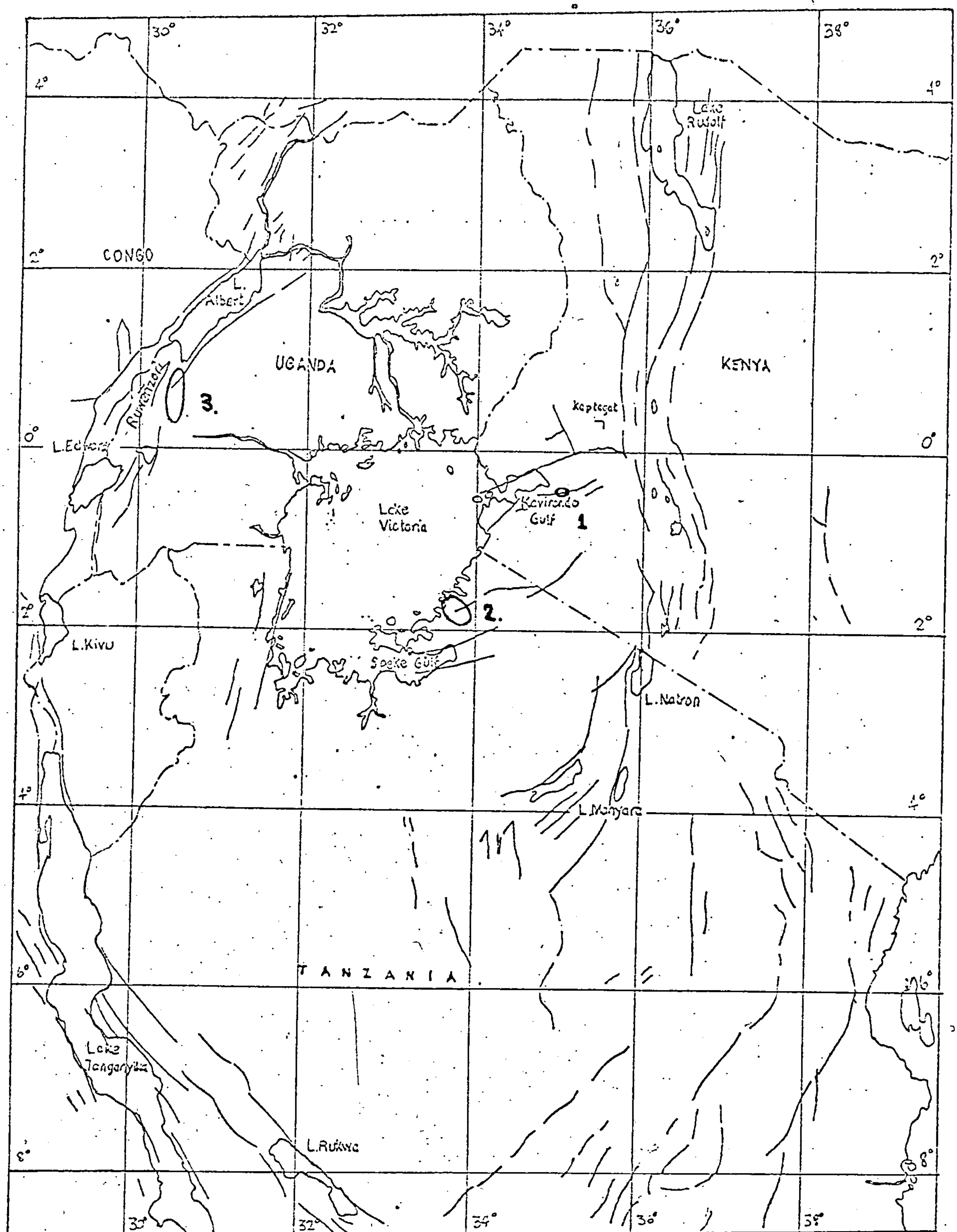
The computer programme that used a least squares fit method to determine apparent velocities on each arm of the array and subsequently the azimuth, made use of the errors in the least squares fit to assign errors to each azimuth.

3.2. Estimated accuracy of epicentre location

The errors given by the least squares fit to the arrival data may be subdivided into three groups.

(a) Errors in picking. These are due to the difficulty experienced, from time to time, in establishing exact correlation between the channels on the printouts.

(b) Error due to the innacuracy in instrument site location. Any consistently large error in the results from one or more instruments could indicate such innacuracy.



MAP INDICATING ACCURACY OF EARTHQUAKE LOCATION

ELLIPSE INDICATING EXPECTED ACCURACY IN:-

1. KAVIRONDO GULF
2. SOUTHEASTERN LAKE VICTORIA AREA
3. WESTERN RIFT

Fig. 8

(c) Any variation in the immediate subsurface geology across the array will affect the ground velocity. It therefore followed that if an instrument were situated on weathered material or soil, or even a boulder which itself was resting on weathered material then the characteristics of the recording from that site would be at variance with those from instruments situated on rock.

There were no consistent delays attributable to any one or more sites, therefore any error from these last two possibilities was negligible and taken into account by the total error given for the azimuth and velocity of each event by the least squares programme.

Comparisons have been made between the location of the major events recorded within the region by the W.W.S.S.S. network and the Kaptagat array. There appeared to be good correlation between the epicentres of events located to the west of Kaptagat. An azimuthal error of approximately 10^{o} was noted when comparing results for the Tanzania event of 9 August 1970, located at a distance of 700 Km from Kaptagat, at the southern end of the eastern rift.

The eastern rift appears to follow the alignment of some structure affecting the velocity of events located both within it and to the east of it when recorded by Kaptagat.

This observation was consistent with the results of the seismic refraction experiment carried out in the Gregory rift by Griffiths et.al. (1971) who suggested a velocity of 6.4 Km/s within 3 Km of the surface, underlain by material of velocity 7.5 Km/s at 20 Km. This was further complimented by gravity measurements made by Khan and Mansfield (1971) who postulated material of density ρ = 3.15 gm/cc within 20 Km of the surface over this area.

Because of the difficulty of establishing what happened to the paths of events recorded from this zone, considerable doubt existed concerning event locations within and to the east of the eastern rift.

The possibility existed that events recorded as lying to the east of the Kaptagat array would be refracted in the horizontal as well as vertical plane when they crossed the boundaries of the anomalous material. This could have led to the incorrect azimuth being assigned to an event.

Uncertainty about the boundary conditions between normal and anomalous material precluded the direct application of a refraction relationship such as Snell's Law.

Errors in epicentral distance calculations were thought to be of the same order as those for events located to the west of Kaptagat, as the α / β ratio remained almost constant in both anomalous and normal areas (Sundaralingam, 1971).

The area to the west of the Kaptagat array may be divided into two parts for the consideration of the accuracy of location of events. At distances less than 180 Km from Kaptagat the first arrival was Pg which was followed by Sg, both of which could be clearly identified and the accuracy of their onsets measured respectively to 0.2 seconds. Taking the worst case where the innacuracy was a combination of the errors then the epicentral location would be to an accuracy better than ± 5 Km.

At distances greater than 180 Km from Kaptagat the first arrival was Pn recorded with relatively small amplitude followed by much larger amplitude Pg which had indistinct onset. For an earthquake located at an epicentral distance of 600 Km, such as those located within the western rift, this meant a delay between the recording of Pn and the arrival of Pg of about 16 seconds. It was therefore thought most unlikely that the incorrect phase had been picked. Similarly, with the S phase, following the arrival of Sn there was a delay of approximately 25 seconds before the arrival of Sg which was closely followed by the surface waves. Misidentification of these phases was therefore ruled out as a possible source of error.

It was however most difficult to establish the exact onset time for Sn due to the relatively small amplitude arrivals which were frequently emergent from the P wave coda. A maximum picking error of 2sec.was thought attributable giving rise to a location accuracy of ± 12 Km for an event located within the western rift, located at an epicentral distance of 600 Km from Kaptagat. For an event situated at a distance of 300 Km the maximum possible error would be $\pm$ 20 Km.

Errors in azimuth have been taken direct from the accuracy of the least squares fit of the arrival times and are better than $\pm 4^{\circ}$ for an event located within the western rift. For nearer events located within the Kavirondo Guld the accuracy was better than $\pm 3^{o}$. The better accuracy was the result of better onset determinations from the clear Pg phase arrival.

The ellipses shown in figure 8 indicate the expected accuracy of event locations occurring in these areas.

The focal depth of earthquakes occurring within East Africa has been considered very shallow (Wohlenberg, 1968; 1970) occurring at depths of less than 55 Km with maxima within their depth distribution at 15 and 27 Km.

A comparison was therefore made between the expected arrival data for an event of focal depth of 40 Km and one of 20 Km. The effect was to alter the Pn to Sn time by an amount of the order of 1 second. As this amount was within the accuracy of picking it was not considered significant.

3.3. Observations on the map of epicentres

The map of epicentres, figure 14 is located in the back pocket of the thesis binding.

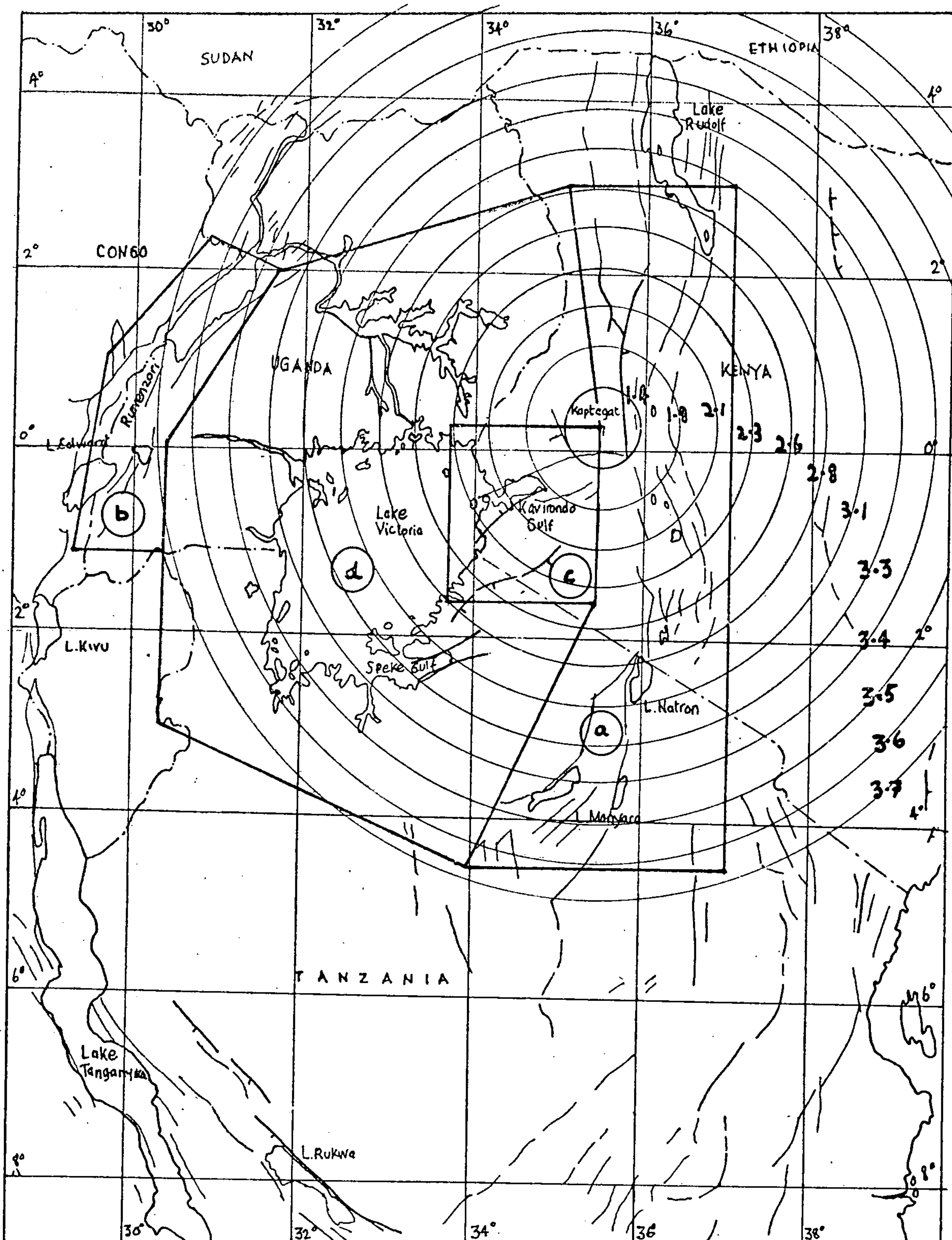
It can be readily seen that the majority of earthquakes recorded occur along or near the major tectonic features of the East African Rifts.

The areas from which the highest number of earthquakes have been recorded are:-

(a) The western rift between latitudes 1^{o}S and 1^{o}N

(b) The Kavirondo rift

The eastern rift in general shows a much lower level of activity than either the western or Kavirondo rifts and with



GEOGRAPHICAL AREAS USED FOR COMPARISON OF ACTIVITY

(a) EASTERN RIFT (b) WESTERN RIFT

(c) KAVIRONDO RIFT (d) LAKE VICTORIA AREA

MAGNITUDE CUT-OFF VALUES FOR KAPTAGAT

Fig. 13

the number of events recorded very similar to that of the Lake Victoria area. See figure 13 for an indication of the geographical extent of the areas considered for the purposes of this comparison.

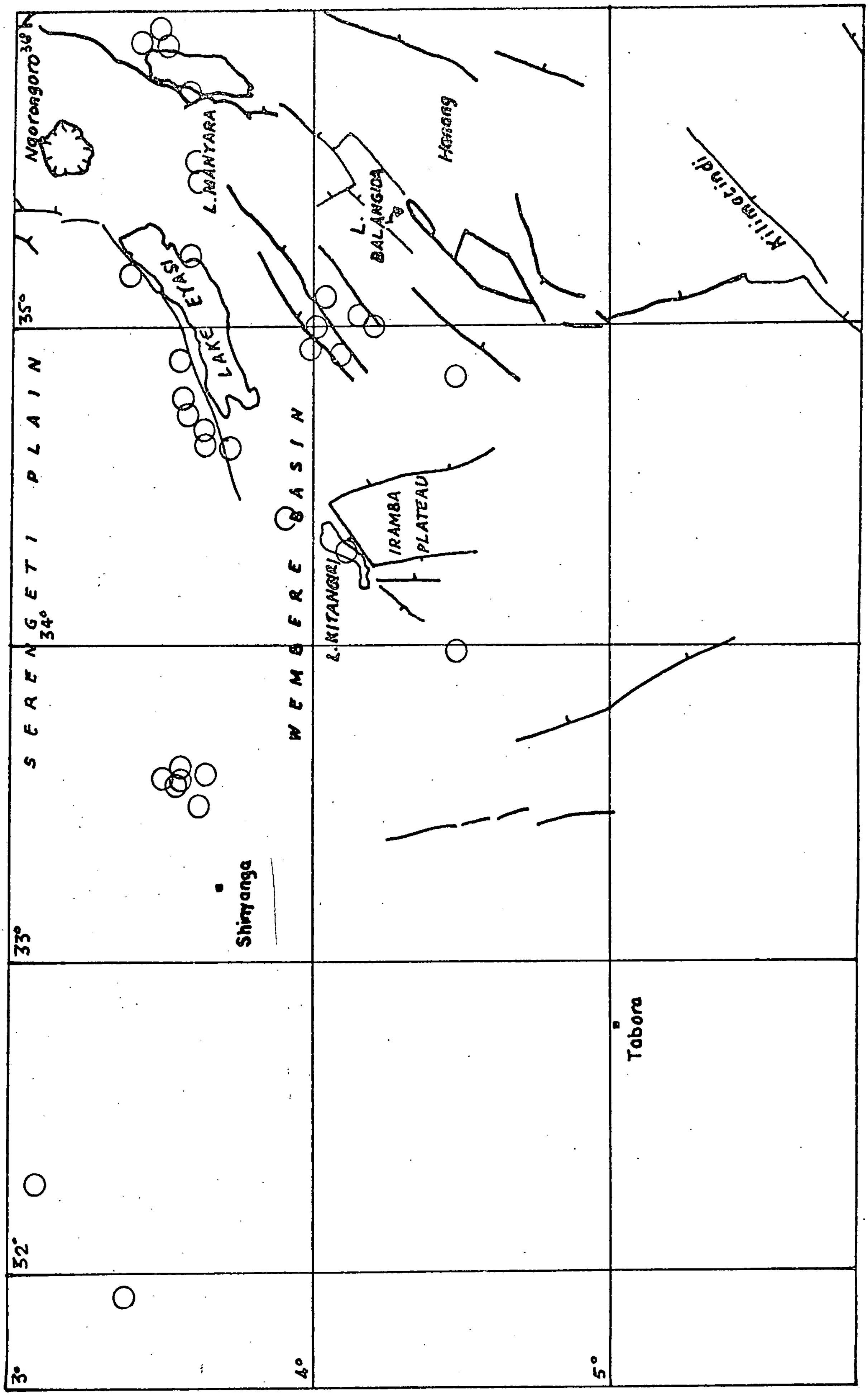
(a) The Eastern Rift

From the map, figure 14 it can be seen that the northern end of the eastern rift exhibits a very low level of activity. This observation is in agreement with all previous work (Sykes and Landisman, 1964; Fairhead, 1968; Wohlenberg, 1968; Rodrigues, 1970). The results of micro-earthquake studies (Molnar and Aggerwal, 1970) conducted within Kenya show that there is an increase in activity from north to south and that the only area of the eastern rift that was very active at the time of the survey (December 1969 to March 1970) was the area around Lake Magadi in the south, near the Tanzania - Kenya border.

Molnar and Aggerwal (1971) argue that the lack of events recorded from this area is a product of the lack of nearby recording stations. The current availability of records from Kaptagat precludes any similar argument.

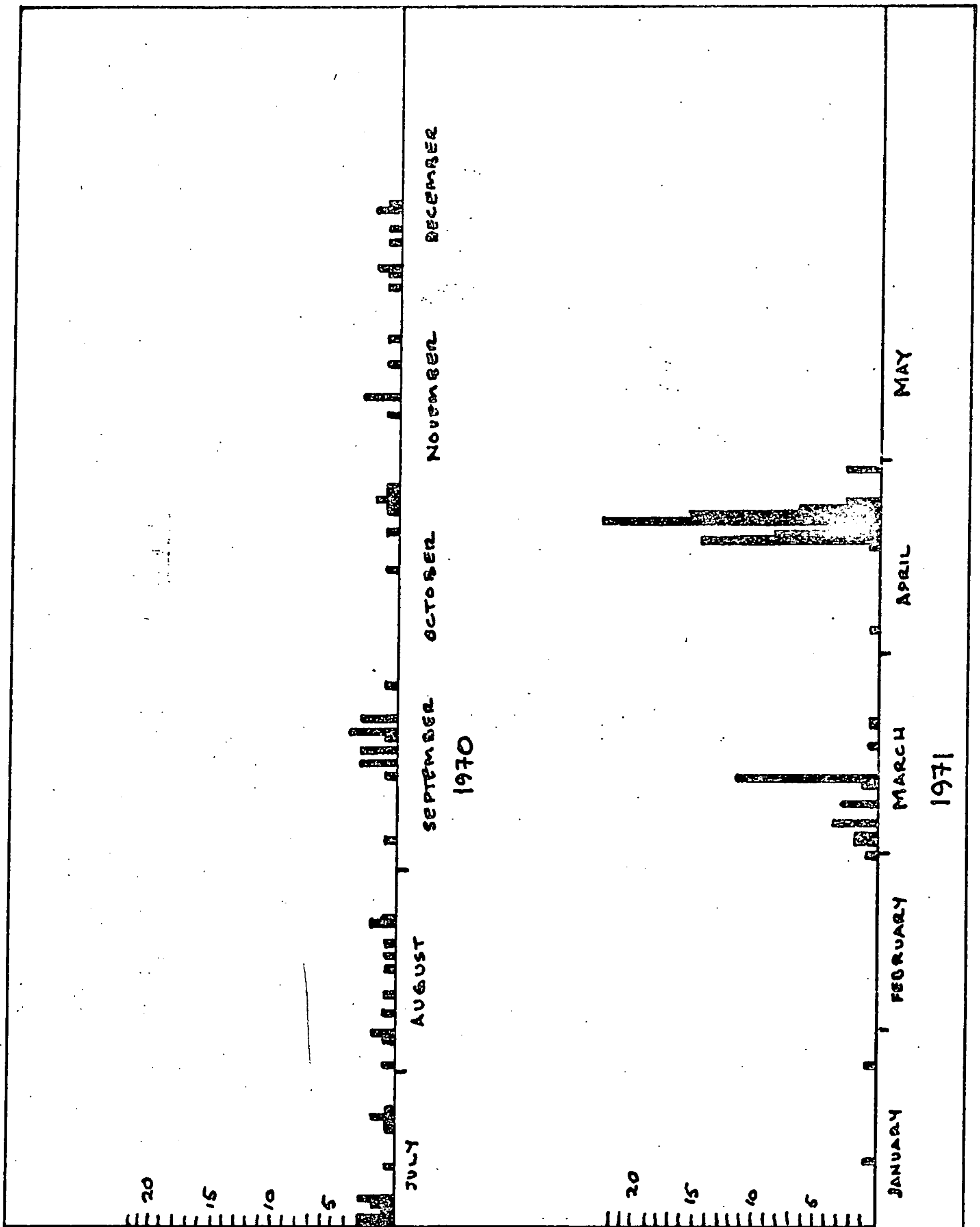
During the period of recording from July 1970 to May 1971, one large earthquake of Mb=4.8 occurred on 9 August 1970 at the southern end of this zone in Tanzania. A number of smaller events were recorded from northern Tanzania, these being generally associated with some known fault or tectonic feature, see figure 15.

The Lake Eyasi - Lake Natron section is the most active part of this area with the earthquakes generally of moderate size.



Ngorongoro
36°
L. MANYARA
LAKE EYASI
L. BALANGIDA
Kilimatindi
35°
SERENGETI PLAIN
34°
WEMBERE BASIN
IRAMBA PLATEAU
L. KITANGIRI
Shinyanga
33°
Tabora
32°
3°
4°
5°

Fig. 15

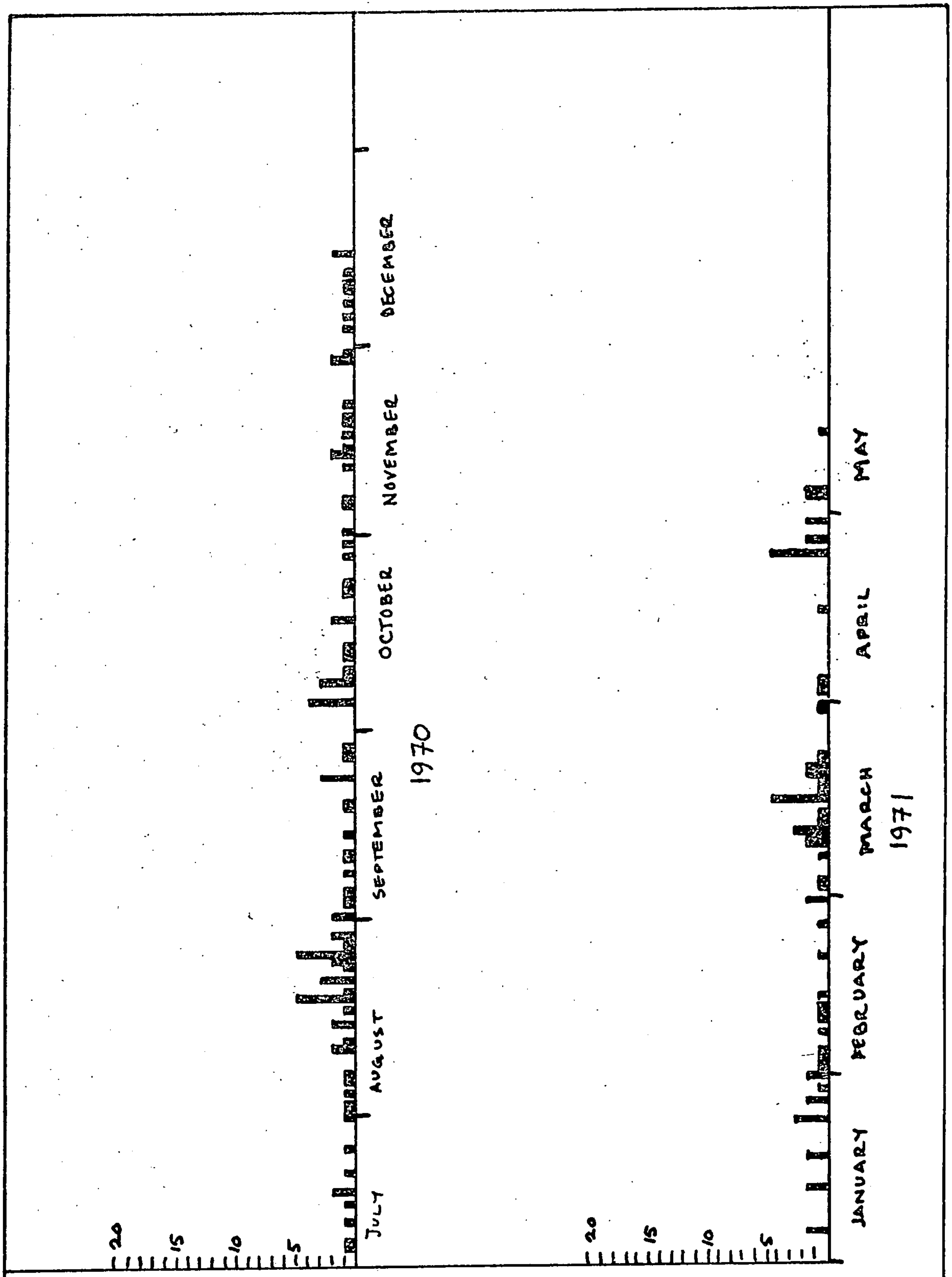


NUMBER OF EVENTS RECORDED PER DAY - WESTERN RIFT

Fig. 20

Previous work compliments the present evidence that events rarely exceed magnitude Mb=5.0. In 1964, however, an earthquake of magnitude Mb=6.4 occurred near the town of Babati (4^{o}S, 35^{o}E) causing small loss of life and much damage to property, with the zone of greatest intensity running NNW, from Kondoa to Oldeani and beyond (Whittingham, 1964). Studies of the after-shocks (Rodrigues, 1971) from this event indicated that the position of the epicentres showed a general parallelism with the trend of the rift faults rather than random location.

In direct comparison with events recorded from the western rift which are generally followed by a high level of after shock activity (see figure 20), events recorded from this region generally appear in isolation rather than as a part of a sequence (see figure 19). Only three events were recorded from the area of the 9 August 1970 event. The evidence cannot, however, be conclusive as any small magnitude after shocks might not be recorded due to the magnitude cut off characteristics at that distance from the recording station though Rodrigues (1970) reached a similar conclusion by a comparative study of the Babati earthquake of 1964 in the eastern rift and the series of earthquakes from western Uganda following the event of 20 March 1966. The differences are thought to represent the different states of stress between the two areas. The Tanzania events (see figure 19) showed relatively localised stress release compared with the Uganda events (see figure 20) which exhibited widespread release of energy in more than fifty events occurring over a period of a few days, with magnitudes varying between M=4.0 and M=4.5 and with activity spread over a large area.



NUMBER OF EVENTS RECORDED PER DAY - EASTERN RIFT

Fig. 19

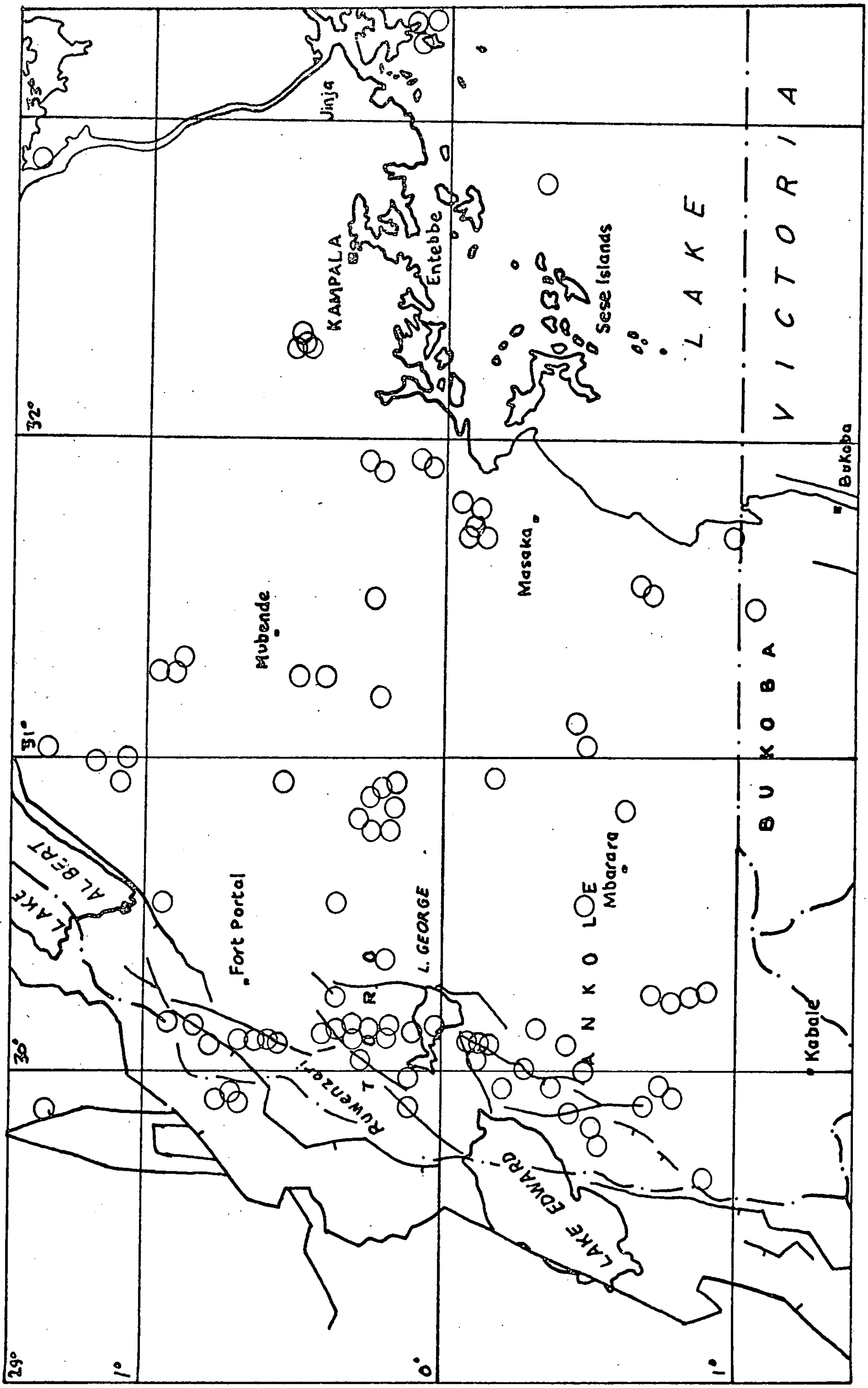
The energy release from this area may be divided into two types. Firstly, the background almost constant 2-3 events per week. Four small bursts of energy occurred in August, October, March and April.

(b) The Western Rift

The Albert Rift, Ruwenzori Massif and the area to the south has traditionally been reported as having a high level of seismic activity. This is the site of the well documented Toro earthquake of 20 March 1966 (Loupebine et.al. 1966; Fairhead, 1968; Lahr and Pomeroy, 1970; Fairhead and Girdler, in press). For the period January 1965 to December 1970 Fairhead showed the larger events to correlate closely with the rift structures with the exception of a few events to the west of the area at about 3°S.

Wohlenberg (1968) using I.R.S.A.C. data was able to study low magnitude events for the period January 1958 to December 1963. His locations correlate well with the main rift trough. Outside the main rift valley only small magnitude events were seen to occur with, once again, the exception of a seismic zone west of Lake Kivu where he located a small number of events.

Fairhead and Girdler (in press) indicate that the activity recorded included a swarm of events located at the northern end of Lake Albert and suggest their relation to normal faulting. In their study of the sequence of events associated with the event of 20 March 1966 they show a concentration of after shocks between the Kitimbi - Semliki fault and the western wall of the main rift valley, not around the main event as had been previously thought. It was suggested that many of the events in this sequence show a trend similar to that of the rift faults and one possibly related



LAKE VICTORIA
KAMPALA
Entebbe
Jinja
Sese Islands
Masaka
Mubende
BUKOBA
Bukoba
Fort Portal
L. GEORGE
LAKE ALBERT
Ruwenzori
TORO
ANKOLE
Mbarara
Kabale
LAKE EDWARD
29°
30°
31°
32°
33°
1°
0°
1°

Fig. 17

to the movement of the western fault escarpment. This major earthquake is thus seen to have triggered adjustment along the western boundary fault.

This type of activity is in good agreement with that observed during the present study where earthquakes were recorded from an area closely associated with, but a little further south than that of the Toro earthquake of 20 March 1966.

Three events:-

Date	Time	Latitude	Longitude	Mb
18 April 1971	00 34 34.1	0.238N	30.142E	4.6
18 April 1971	05 49 49.0	0.169N	30.196E	4.7
21 April 1971	18 40 54.0	0.205N	29.899E	4.3

reported by the U.S. Coast and Gedetic Survey (now N.O.A.A.R.) were also recorded by Kaptagat. These events were part of a sequence numbering over seventy, located with wide latitudinal scatter but generally restrained to the narrow zone of the western flanks of the Ruwenzori massif first north of Lake George, see figure 17.

There is a notable lack of events from the Lake Albert area, with only three located on the eastern boundary fault, during the period of the present study but several events were recorded from southern Sudan - northern Uganda area. The largest of these events was reported by N.O.A.A.R. with the following details

Date	Time	Latitude	Longitude	Mb
04 January 1971	15 14 35.4	3.641	32.450	4.4

This event was not recorded by Kaptagat as having any large magnitude aftershocks. Smaller after shocks may have occurred, the Kaptagat station magnitude cut off for this distance is Mb=3.7.

It will be seen from figure 13 that the southern part of the western rift exceeds a distance of 600 Km from Kaptagat. This effectively means that events occurring with a magnitude less than Mb=3.7 will not be recorded from this area. It would appear from a scrutiny of the bulletins produced by the organisations which correlate data on a world wide basis that no events greater than that magnitude have been registered from this area during the period of time covered by the present study.

Earthquakes have been previously recorded from the southern part of the western rift (Fairhead and Girdler, in press) indicating that the seismic activity has in the past been fairly widespread but generally located within the rift structures. Sykes and Landisman (1964) as well as Fairhead and Girdler (in press) locate concentrations of activity around Lake Tanganyika, Lake Rukwa and further south to the northern end of Lake Malawi.

The activity of the western rift, see figure 19, appears to be sporadic with small bursts of energy released during the second half of 1970 with the largest earthquake in each sequence of magnitude between Mb=4.0 and 4.5. Only three events were recorded during January and February 1971 but this period of quiescence was followed by two large bursts of energy release, the greater of which occurred during the second half of April 1971 with the largest magnitude recorded of Mb=4.7.

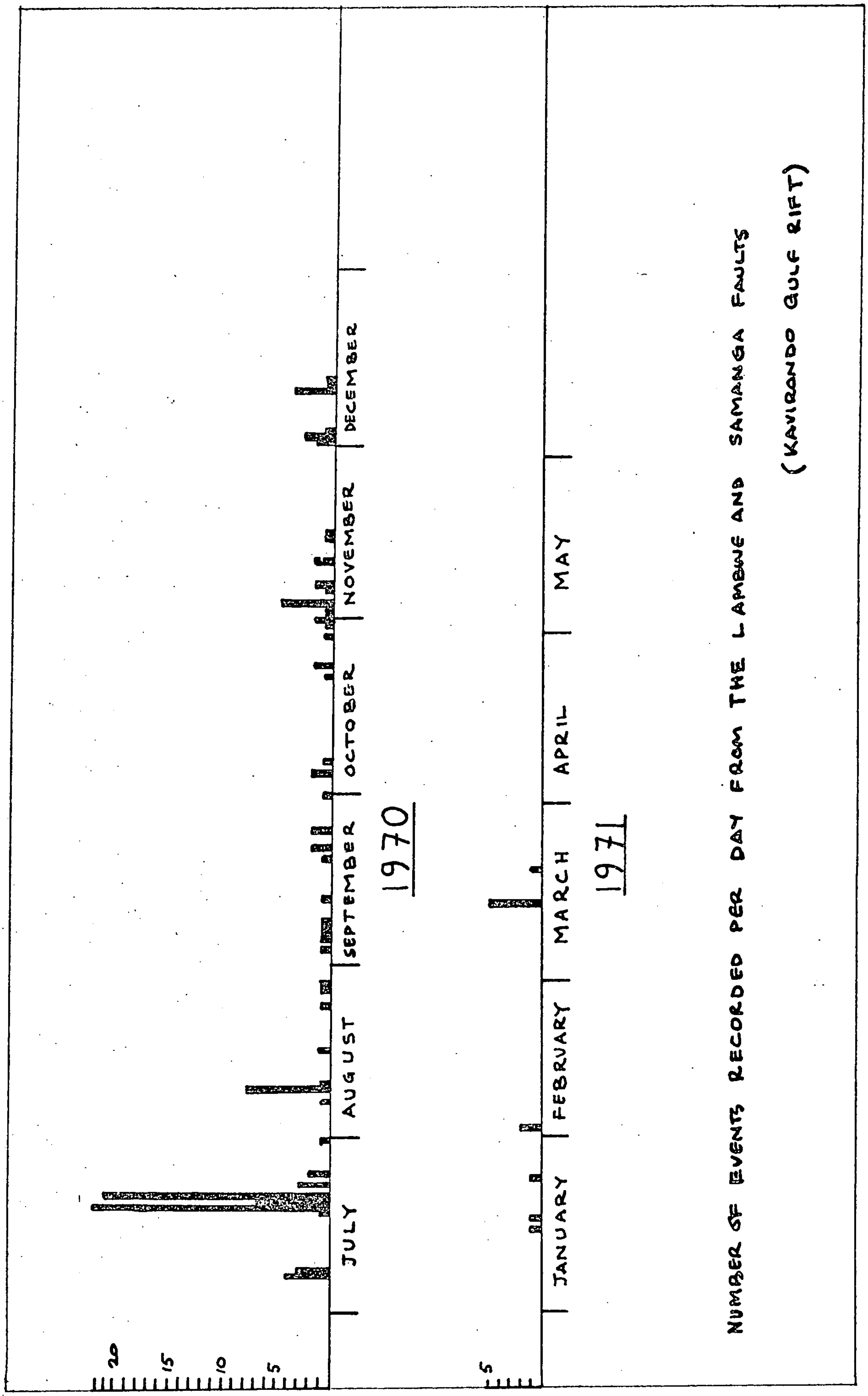
It would appear that as long as there is a reasonably frequent occurrance of earthquakes of magnitude up to Mb=4.5 then stress does not build up to give large bursts of energy release but as soon as there is a quiescent period then a fairly concentrated subsequent activity may be expected.

The present continuation of recording within East Africa should allow more evidence for the development of this pattern of stress release behaviour which is not wholly conclusive due to the relatively short period of recordings currently analysed.

(c) The Kavirondo Gulf Rift

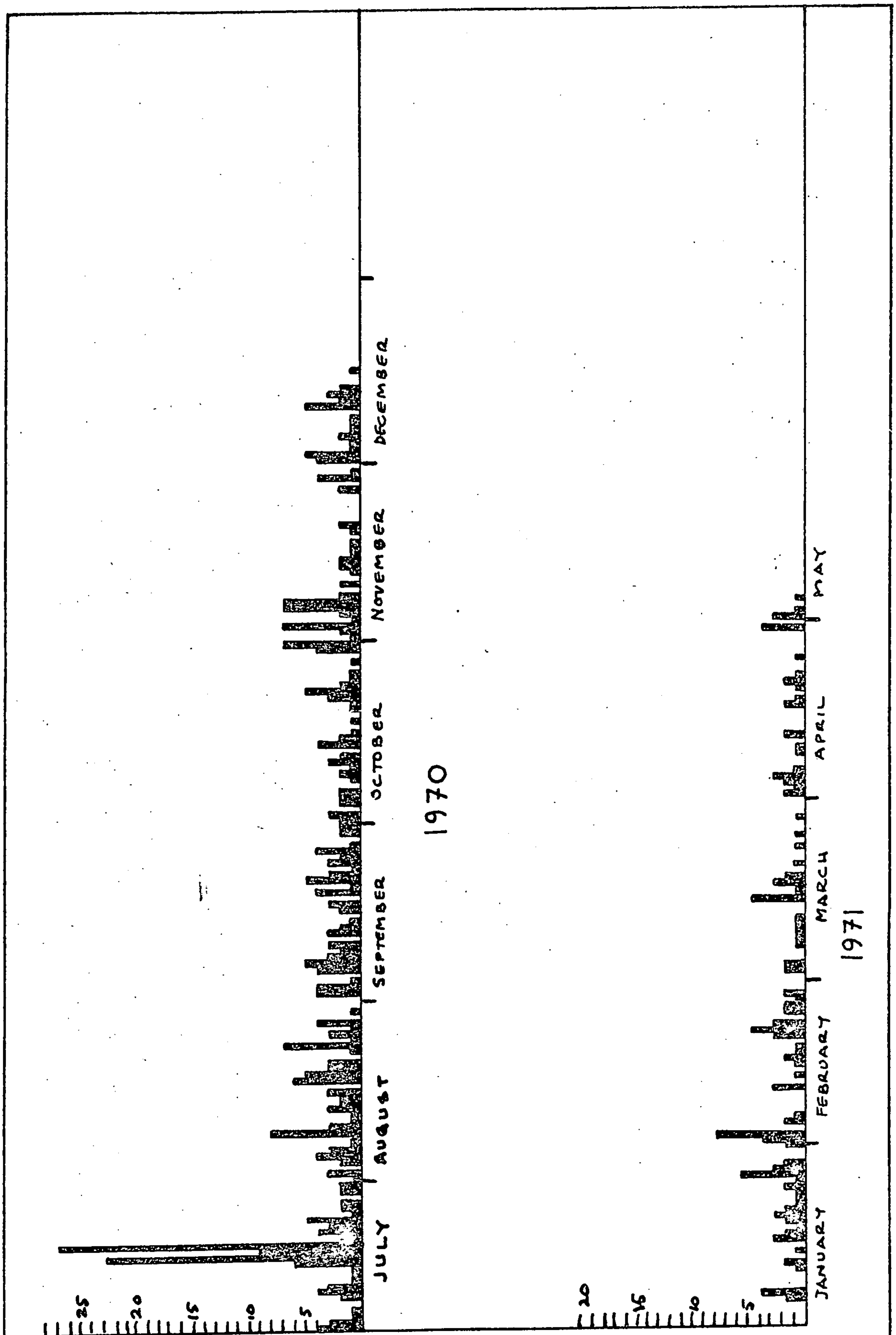
Previously little seismic importance has been attached to this rift. In March 1968 a series of earthquakes shook the town of Homa Bay, located on the southern shore of the Kavirondo Gulf, causing one death and extensive minor damage. Instrumental data, as well as the distribution of isoseismals (Loupekine, 1968) showed that the earthquakes were due to movements along sectors of the Lambwe and Samanga faults. At the time of these earthquakes there had been abundant rainfall and the level of Lake Victoria was high. Loupekine (1968) postulated that the added weight of water, as well as the lubrication of the fault plane by the water, could have triggered these earthquakes. Such a mechanism has been postulated for the Indian earthquake of 10 December 1967 (Rothe, 1968) and the events within the artificially created Lake Kariba in southern Africa (Gough et.al. 1970).

The present work shows extremely good correlation between events and the Lambwe and Samanga faults but does not restrict the events to the area near the town of Homa Bay as do previous workers (Loupekine, 1968; Molnar and Aggerwal, 1971) but indicates that the activity has occurred throughout the length of these faults which trend SW-NE. The relatively short periods over which recordings were made by previous workers probably resulted in this mis-identification of the level and extent of activity in this area.



NUMBER OF EVENTS RECORDED PER DAY FROM THE LAMBWE AND SAMANGA FAULTS
(KAVIRONDO GULF RIFT)
1970
JULY
AUGUST
SEPTEMBER
OCTOBER
NOVEMBER
DECEMBER
1971
JANUARY
FEBRUARY
MARCH
APRIL
MAY
5
10
15
20
5

Fig. 16



NUMBER OF EVENTS RECORDED PER DAY - KAVIRONDO RIFT

Fig. 21

Activity associated with these faults extends into Lake Victoria. The largest magnitude recorded from the Kavirondo rift area was Mb=3.4 and the majority of events have a magnitude between M=1.8 and Mb=2.4. The cut off for events of that size lies on the western boundary of the area, see figure 13. If there were events located further to the west, within Lake Victoria, they would not have been recorded at Kaptagat. The zone of activity associated with this rift may therefore extend further into Lake Victoria.

Although no data is available for variations in the level of Lake Victoria during the period of recordings represented by the present study, it is noteworthy that the plot of events that occurred along the Lambwe/Samanga fault with respect to time, see figure 16 shows that the majority of events occurred at times of heavy rainfall and that during the dry months the seismic activity remained at a relatively low level. The period of recordings is, however, too short to give well documented correlation between these two phenomena.

The large series of shocks that occurred during the period July 18-20 1970 included earthquakes of magnitude up to Mb=3.1.

The northern boundary faults of the Kavirondo Gulf rift have also been active during the period with activity being recorded from a relatively wide range of locations ranging from the shores of Lake Victoria to the junction with the eastern rift.

Other earthquakes have been recorded from the faults north of and parallel to the northern boundary fault. Activity has also been recorded from the Siria and Uitumbara faults which run sub-parallel to and some eighty kilometres to the south of the main rift.

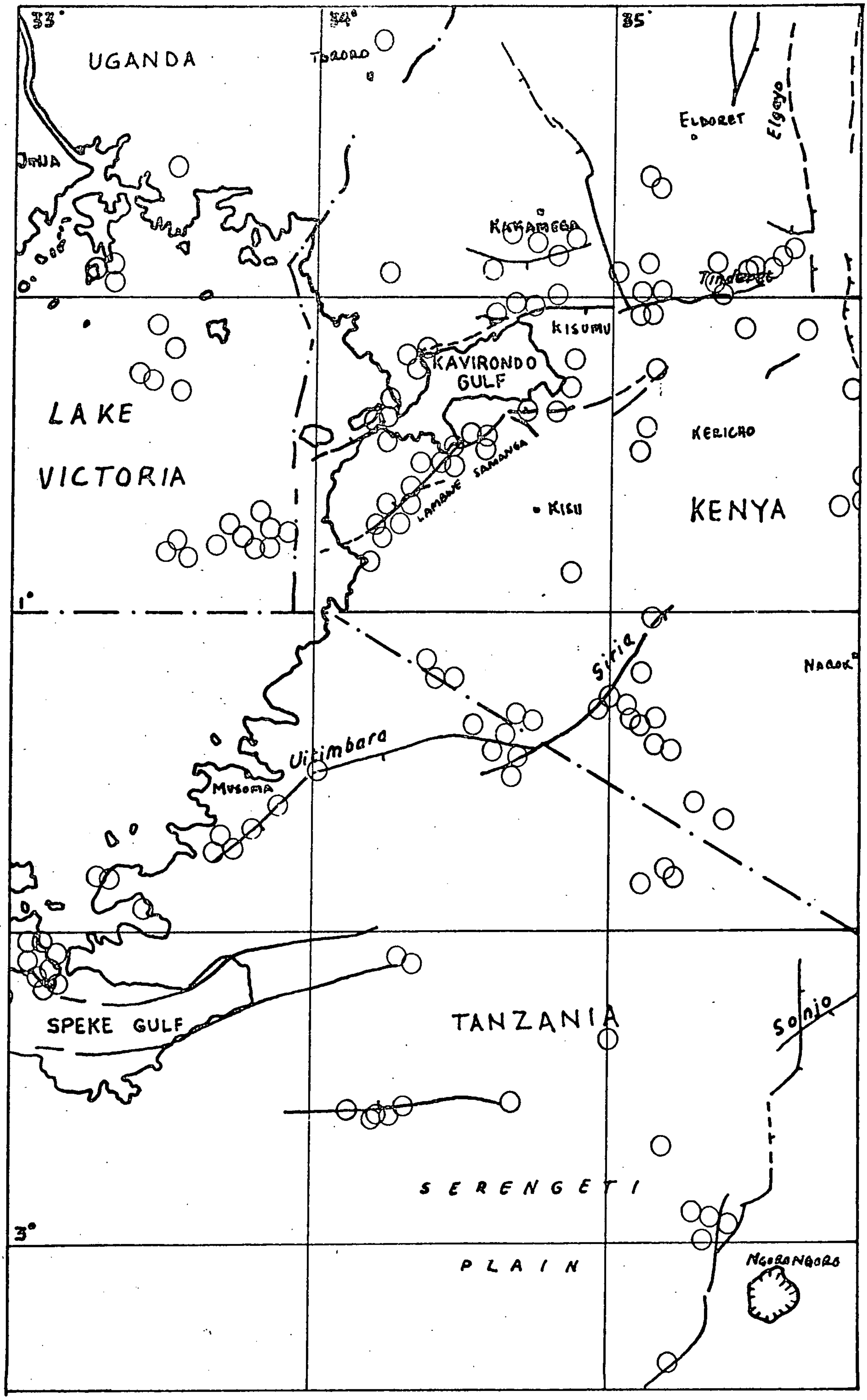

33°
34°
35°
UGANDA
TORORO
ELDORET
Elgeyo
KAKAMEGA
Tinderet
KISUMU
KAVIRONDO
GULF
LAKE
VICTORIA
KERICHO
LAMBWE
SAMANGA
KISII
KENYA
1°
Siria
NAROK
Utimbara
MUSOMA
SPEKE GULF
TANZANIA
Sonjo
SERENGETI
PLAIN
3°
NGORONGORO

Fig. 23

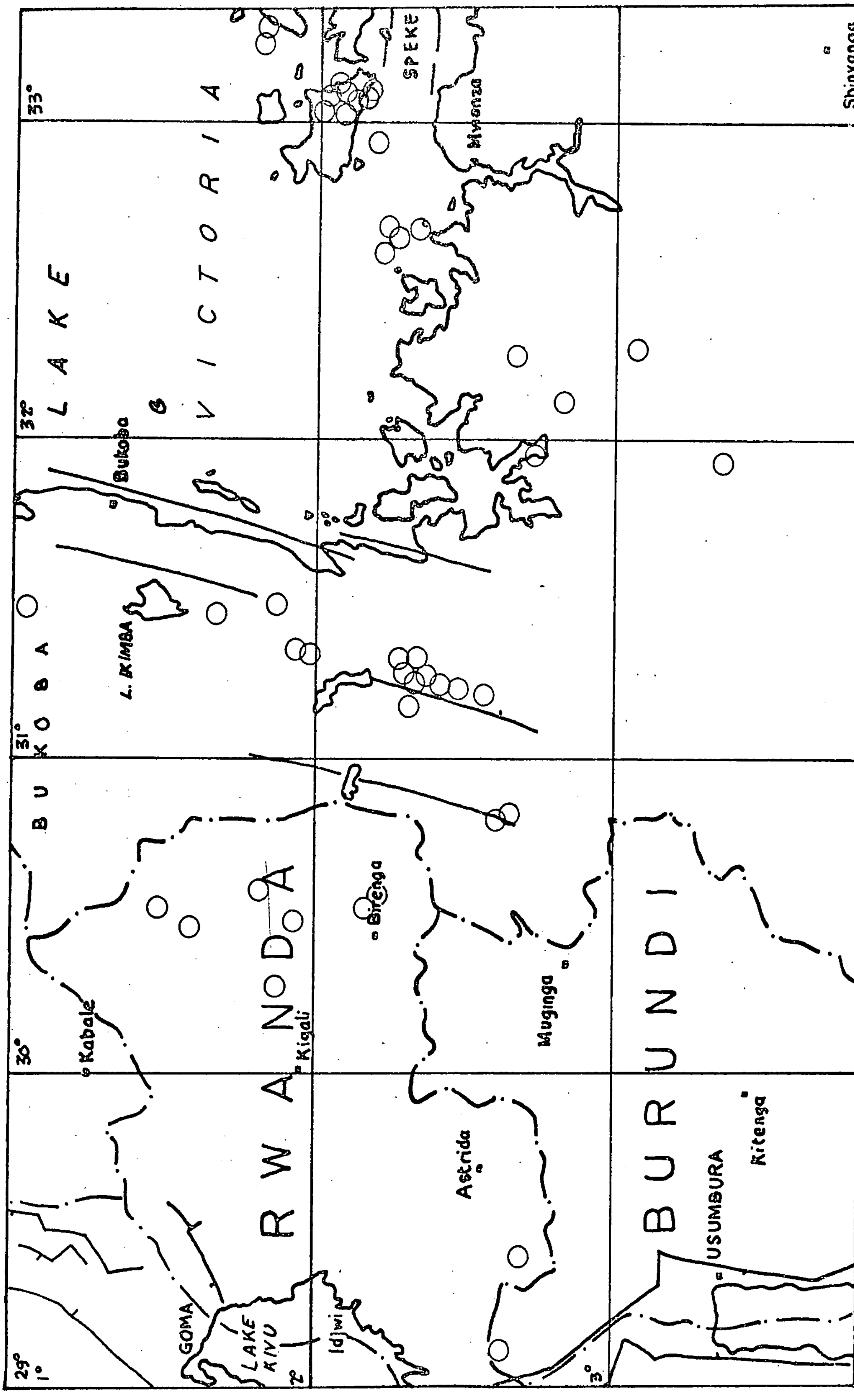

LAKE VICTORIA
SPEKE
Mwanza
Shinyanga
Bukoba
L. IKIMBA
BUKOBA
RWANDA
Kabale
Kigali
Birenga
Muginga
BURUNDI
Astrida
Kitenga
USUMBURA
GOMA
LAKE KIVU
Idjwi
29°
30°
31°
32°
33°
1°
2°
3°

Fig. 24

The event occurrance/time plot, see figure 21, for this area by comparison with that for the Lambwe/Samanga fault, see figure 16, shows fairly continuous and sustained activity. By subtracting the events that occurred on the Lambwe-Samanga fault it can be seen that the level of activity alters little during the period July to December 1970 but declines slightly during the early part of 1971.

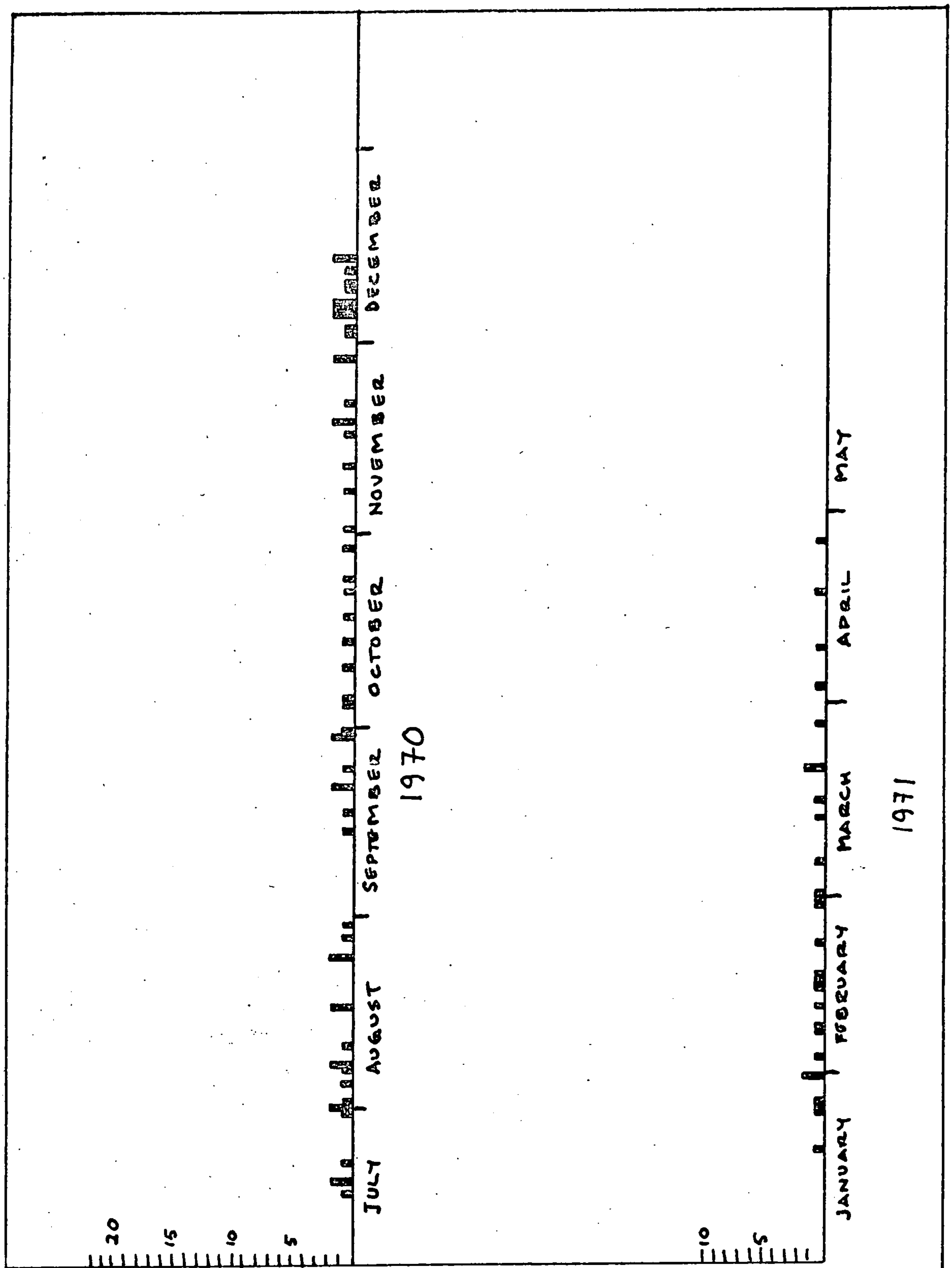
(d) The Lake Victoria Area

Outside the rift zones, earthquake epicentres are widely scattered and frequently occur in areas where there are no apparent faults, see figure 14. Earthquakes have been relatively common, both in areas surrounding Lake Victoria and within the Lake.

Earthquakes have been recorded at Kaptagat from the rift like structure of the Speke Gulf in the south east of Lake Victoria, and its westward extension into the lake, see figure 23. Scattered events occur within the southern margins of the lake, and at its south western corner appear to be aligned along faults running sub-parallel to the western rift in a north-south direction, see figure 24. The general north-south alignment of the structure of this area controls to a great extent, the physical shape of this part of Lake Victoria.

There are scattered events located within Rwanda but with no correlation with structure. South western Uganda is the location of further scattered events but further north there would appear to be a zone of weakness, occasioning several events, which closely follows the Katonga River.

Events located to the north of Mubende have been recorded, which appear to come from a zone of structural weakness within the northern part of the Mubende granite batholith. As previously noted, the north eastern sector of Lake Victoria exhibits activity



NUMBER OF EVENTS PER DAY - LAKE VICTORIA REGION

Fig. 22

which could possibly be the lakeward extension of the zone of structural weakness associated with the Kavirondo rift.

The event occurrance/time plot, figure 22, shows that the distribution of earthquakes within the period of study was almost constant. The pattern indicates that there is a general equalisation of stress taking place over the whole area. The largest magnitude recorded was Mb=4.2 and within the distribution of magnitudes from this area a peak occurred at Mb=3.5.

(e) Other Areas

Occasional events were recorded from an area of southern Ethiopia, to the east of Lake Rudolph, and from other scattered locations within eastern Kenya. No correlation was found between these isolated events and any mapped geological feature.

CHAPTER FOUR

4.1 Magnitude determinations of earthquakes at small epicentral distances

The first widely accepted instrumental magnitude scale was developed by Richter in 1935. In this work he defined a quantity M, to be called the magnitude, given by the relationship:-

$$M = \text{Log } A - \text{Log } Ao \quad (\Delta)$$

where A was the recorded trace amplitude for a given earthquake at a given distance as written by a standard type of instrument and Ao as that for a particular earthquake selected as standard.

The following quantities entered into the definition of M;-

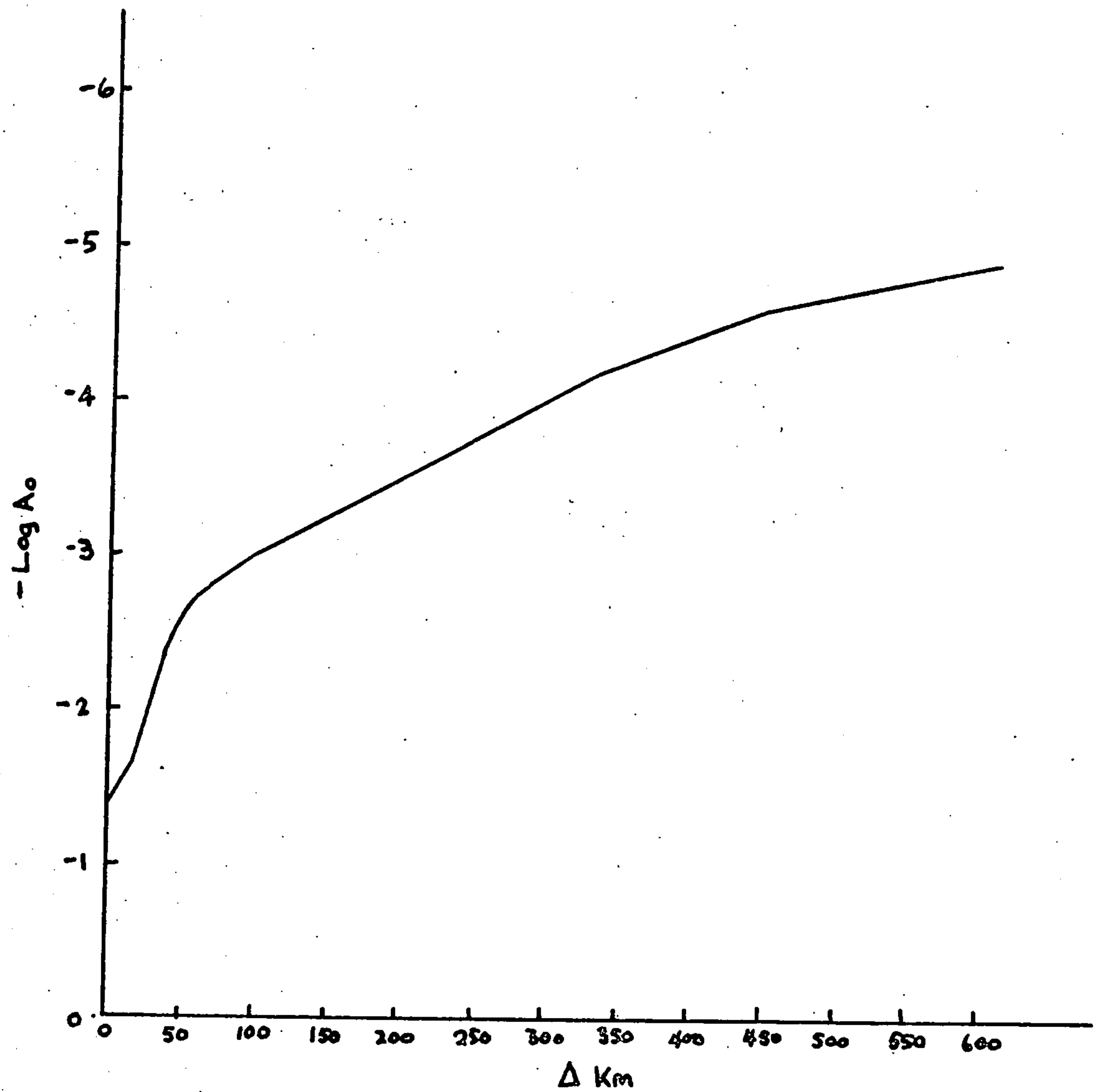
(a) The use of a particular type of seismometer with specified constants.

(b) The selection of a standard shock whose amplitude was represented by Ao.

(c) The variation with epicentral distance of the value of Ao.

The magnitude M, is a number characteristic of the earthquake and independent of the location of the recording station. The standard shock was defined as a shock of zero magnitude since if A=Ao then M=O.

A small earthquake could conceivably be recorded with an amplitude smaller than that of the standard shock; this would allow the calculation of a negative magnitude. The zero level value for Ao can be fixed by stating its value at a particular



LOG Ao - Δ CURVE FOR SOUTHERN CALIFORNIA

(RICHTER, 1935)

CONFIRMED EVERNDON (1967)

fig. 18

distance between epicentre and recording distance. Richter stated this value as one thousandth of a millimetre of ground motion at a distance of 100 Km from the earthquake epicentre. See figure 18 for the Ao–Δ curve given by Richter (1935).

Gutenberg and Richter (1956) suggested a revision of the term Log A to include the period T in the form Log $\frac{A}{T}$ as earthquake magnitudes over the distance range 0–2000 Km had been seen to be period dependent. They also suggested the necessity for both regional and station corrections. The revised relationship can be written thus:-

$$M = \text{Log} \left\{\frac{A}{T}\right\} - \text{Log} \left\{\frac{A}{T}\right\}_o + C_1 + C_2 \quad - (1)$$

where C_1 is the regional correction and C_2 the station correction.

Evernden (1967) has shown in his work on near regional magnitude calculations within the United States that shield and cratonal areas of low mean elevation display a simple Log $\left\{\frac{A}{T}\right\}$ against Δ relationship. He confirmed the shape of the curve calculated by Richter (1935) for -Log A_o over the epicentral distance range 0-800 Km (see figure 19).

4.2 Applicability of the Richter magnitude scale to East Africa

The applicability of Richter's magnitude scale depends upon the correct establishment of standard values of -Log $\left\{\frac{A}{T}\right\}_o$ and Δ.

From a consideration of the term Log $\left\{\frac{A}{T}\right\}$ it is obvious that any significant change in the value of T, the period of the P-wave arrival, will greatly affect the value of the term.

TABLE 9

Events recorded jointly by Kaptagat and W.W.S. Stations

Date	Time	Latitude	Longitude	Mb	Dist (Km)	Amplitude (A)mm	Mb-Log A
09 Aug 1970	00 26 54.9	6.001°S	34.766°E	4.8	710	8.0	3.9
04 Jan 1971	15 14 35.4	3.641°N	32.450°E	4.4	600	5.5	3.7
18 Apr 1971	00 34 34.1	0.238°N	30.142°E	4.6	600	8.0	3.7
18 Apr 1971	05 49 49.0	0.169°N	30.196°E	4.7	597	10.0	3.7
21 Apr 1971	18 40 54.0	0.205°N	29.899°E	4.3	630	4.5	3.7

The variation in period for events recorded at the Kaptagat array was negligible. For the purpose of this study T was taken to be constant leaving a simple relationship between amplitude A and distance Δ.

The Richter (1935) curve was devised for events occurring within southern California. The mean focal depth for these events is about 16 Km (Richter, 1935) which is about the same as the focal depth within East Africa. No data is available from East Africa to make a firm comparison of crustal structure. It is however not likely to differ significantly from that of California to significantly affect the Ao - Δ, except as described later to adjust the knee of the curve.

4.3 Log $\left\{\frac{A}{T}\right\}_o$ - Δ curve for Kaptagat

The basic shape of Richter's (1935) Log Ao - Δ curve is maintained in the present work. The values of Log $\left\{\frac{A}{T}\right\}_o$ have however been modified to take into account the instruments in use and the station and regional corrections.

Five regional earthquakes that occurred within an epicentral distance of about 600 Km from Kaptagat and whose location and magnitude were reported by the U.S.C.G.S. were used for correlation purposes. Details of these earthquakes, also recorded at Kaptagat are shown in Table 9.

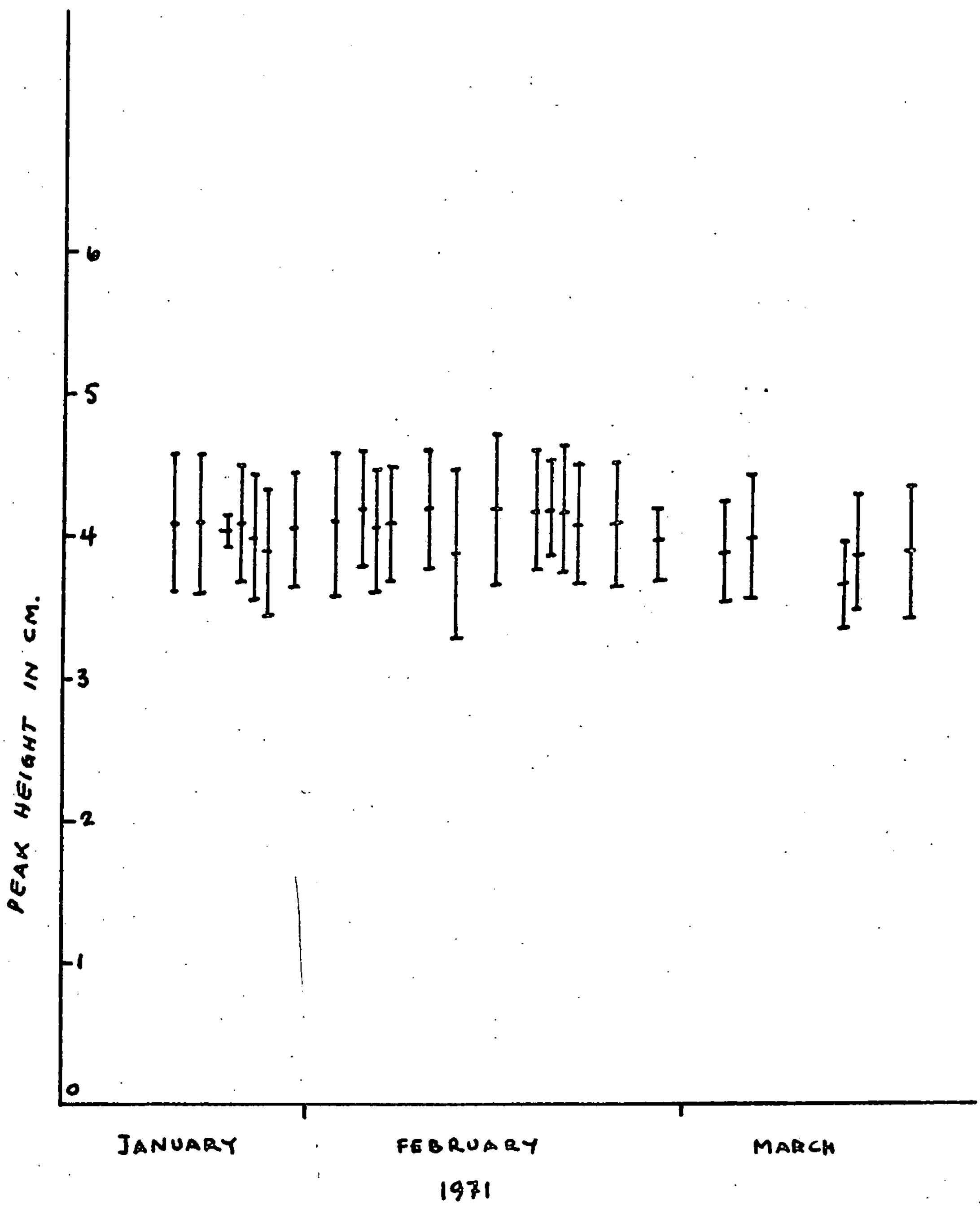
The magnitudes of these events established by U.S.C.G.S. were therefore available for the creation of a meaningful magnitude scale for earthquakes recorded at Kaptagat. As the magnitudes of these events were already known and as it was possible to measure the trace amplitude A, on the printout of each event, the quantity $-\text{Log}\left\{\frac{A}{T}\right\}_o + C_1 + C_2$ was determined by substitution into equation (1). Calculation was made using the 18 April 1971 event of magnitude Mb=4.7 and this gave the value of $-\text{Log}\left\{\frac{A}{T}\right\}_o + C_1 + C_2$ at a distance of 600 Km as 3.7. Calculations on the other events confirmed this value. By using this method to establish the magnitude scale for Kaptagat, station and regional corrections were automatically taken into account.

As no earthquakes were recorded by the W.N.S. network, that were nearer to Kaptagat than approximately 600 Km, during the period that Kaptagat has been functioning there is no further data with which to correlate magnitudes at the present time.

Richter (1935) assigns - logA the value of 4.9 at an epicentral distance of 600 Km. As the value for an event occurring at an epicentral distance from Kaptagat of 600 Km was 3.7, the magnitude scale for Southern California was reduced by an amount of 1.2 throughout the scale giving the values shown in figure 34.

It is expected that because of the continuation of recordings at Kaptagat, larger more local events may be recorded simultaneously at that station and the W.W.S. stations and from a comparison of the magnitudes calculated by these two sources, modifications to the Kaptagat magnitude scale, should it be found necessary could be carried out.

The maximum trace amplitude of the first six cycles of the first P-wave arrival was measured on a standard amplitude printout for each event and its magnitude calculated with reference to the magnitude scale of $\text{Log}\left\{\frac{A}{T}\right\}_o + C_1 + C_2$.



VARIATION IN AMPLITUDE OF CALIBRATION PULSES

PEAK AVERAGE WITH STANDARD DEVIATION PLOTTED AS ERROR BAR.

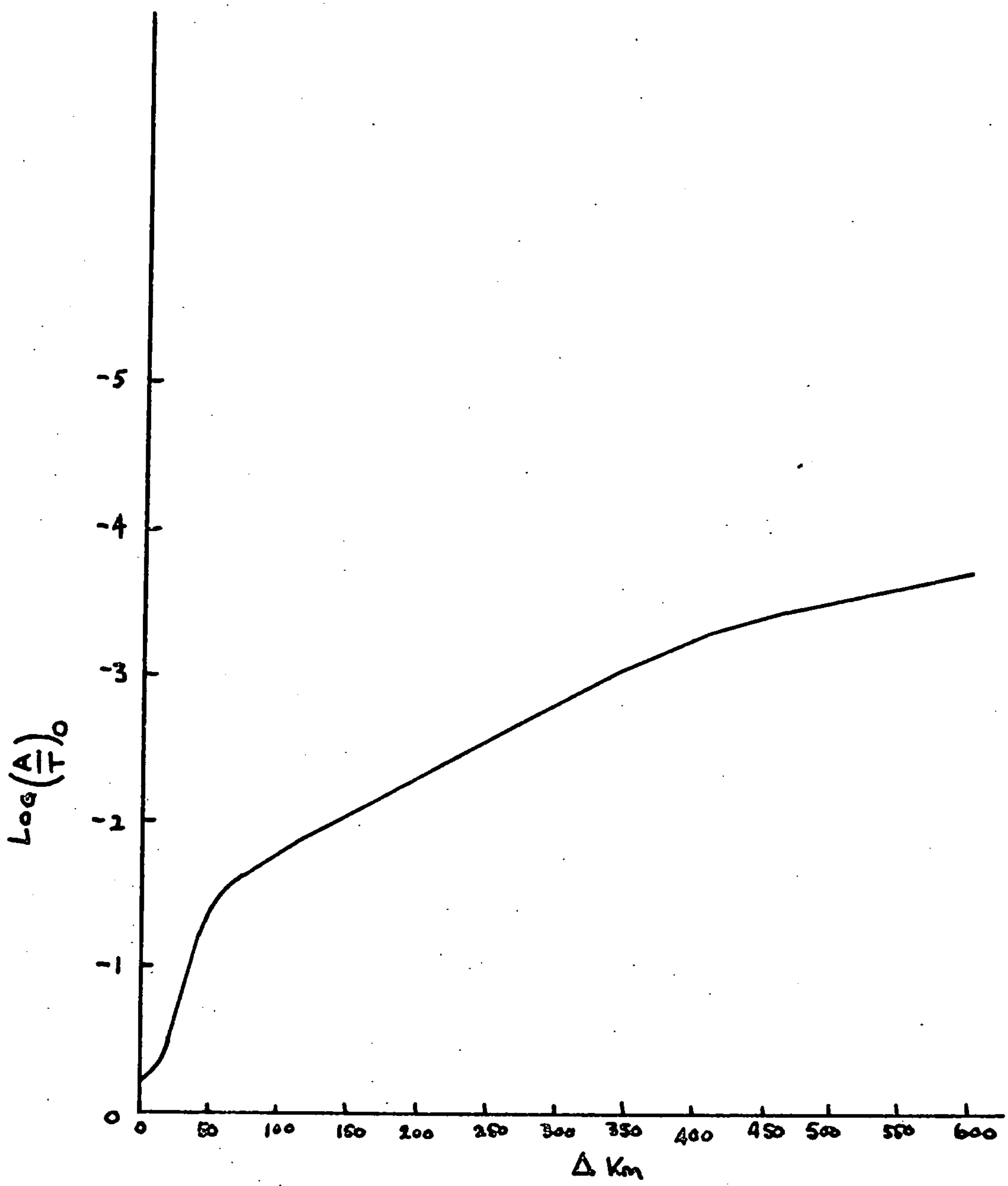
fig. 12

Because earthquakes are normally the product of fault movements, the stress pattern developed around that movement will be indicated by the relative energy levels radiated in different directions from the earthquake. Therefore for any event to be assigned a meaningful magnitude, data from several stations surrounding the epicentre is required. As the magnitude scale was erected with the use of only five correlative events, it is possible that the directional effects of those earthquakes may have led to inaccuracies on its being applied to the whole of East Africa but it is hoped, as more data becomes available, to update the scale.

4.4 Accuracy

Because of possible inaccuracies, particularly in the assignment of magnitude to very close events, where the first arrival is Pg and not Pn, the plots of magnitude/frequency are only given to the nearest half magnitude.

By the use of the daily calibration routine throughout the period of recording, it has been possible to ascertain that variations due to the recording instrumentation were negligible. The calibration pulses were generated electronically from a square wave sequence, see figure 5a, and it was observed that both the amplitude and signature of the pulses remained almost constant during the whole period of recording. Figure 12 shows the variation in pulse amplitude with time for a representative section of the results played back at Durham. The slight variation seen could be due to variations in either the recording equipment in Africa or the playback equipment in Durham, or both.



$\mathrm{Log}\left(\frac{A}{T}\right)_0 - \Delta$ CURVE FOR THE KAPTAGAT ARRAY

Fig. 26.

Figure 34

Values of Log $\left(\frac{A}{T}\right)_o$ - Δ scale for

Kaptagat

Δ Km	−log Ao	Δ Km	-log Ao
0	0.2	260	2.6
5	0.2	270	2.7
10	0.3	280	2.7
15	0.4	290	2.8
20	0.5	300	2.8
25	0.7	310	2.9
30	0.9	320	2.9
35	1.1	330	3.0
40	1.2	340	3.0
45	1.3	350	3.1
50	1.4	360	3.1
55	1.5	370	3.1
60	1.6	380	3.2
65	1.6	390	3.2
70	1.6	400	3.3
80	1.7	410	3.3
85	1.7	420	3.3
90	1.8	430	3.4
95	1.8	440	3.4
100	1.8	450	3.4
110	1.9	460	3.4
120	1.9	470	3.5
130	2.0	480	3.5
140	2.0	490	3.5
150	2.1	500	3.5
160	2.1	510	3.6
170	2.2	520	3.6
180	2.2	530	3.6
190	2.3	540	3.6
200	2.3	550	3.6
210	2.4	560	3.7
220	2.45	570	3.7
230	2.5	580	3.7
240	2.5	600	3.7
250	2.6		

One of the major problems involved in establishing a magnitude scale for an area like East Africa is the lack of information about crustal thickness, The thickness of the crust will alter the position of the change in gradient on the log Ao/ Distance curve that occurs at the distance at which the first arrival changes from Pg to Pn. The change in gradient shown in figure 26 occurs at a distance of approximately 65 Km from Kaptagat.

The slope of the curve, see figure 25, that refers to distances between 200 Km and 600 Km from Kaptagat may be slightly inaccurate but because this slope is gentle any inaccuracy will only change the calculated magnitude by a small amount.

Because of the proximity of the Kaptagat array to the structural complexity of the eastern rift and its associated crustal attenuation (Griffiths et.al., 1971), and the focal depths of events located close to Kaptagat, the position of the change in slope in the log Ao/Distance curve, originally calculated for Southern California where the crustal thickness has been given as between 22-26 Km (Mikumo, 1965), was thought to be realistic. Recordings available from other Durham stations within East Africa should allow the characteristics of the curve to be checked, but this is outside the scope of the present work.

As the northern part of the western rift was roughly equidistant along its length, from Kaptagat the magnitudes calculated for that area were not subject to any innacuracy that may exist in the Log Ao/ Distance curve. The magnitude of earthquakes recorded from this area was therefore related directly to the values given by U.S.C.G.S. and was not dependent upon the scale erected in this work for other areas of East Africa.

4.5 Magnitude Cut-Off Conditions

Because of the correlation between epicentral distance from the recording station and magnitude discernability it is important, when considering results from one recording station in isolation, to correct for the progressive cut off of increasing magnitude with epicentral distance.

In this work the region has been split into four areas:-

(i) Eastern rift

(ii) Western rift

(iii) Kavirondo Gulf rift

(iv) Lake Victoria region

These areas are indicated on figure 13.

The minimum trace amplitude that could be measured with any accuracy was a value twice the amplitude of the system noise.

To consider the seismicity of the four areas with respect to each other a recalculation of the number of events occurring at any one magnitude was necessary. It will be seen from figure 13, that the areas are situated at different distances from the recording station and that they have vastly differing shapes. This figure shows the cut off magnitudes plotted as a radius from Kaptagat. From this it can be seen that for an area such as the western rift it is not reasonable to expect a magnitude of less than Mb=3.4. It will also be seen that if earthquakes of a magnitude, which has a cut off distance that divides the area under investigation, then the recorded earthquakes of that magnitude are representative of only part of the area and their number has to be increased by a factor to enable them to be representative of the whole. The factor used is a simple percentage addition based on the relative areas included and not included by the dividing cut off line.

4.6 Magnitude-Frequency Plots

The magnitude and frequency of earthquakes satisfy the empirical relationship:-

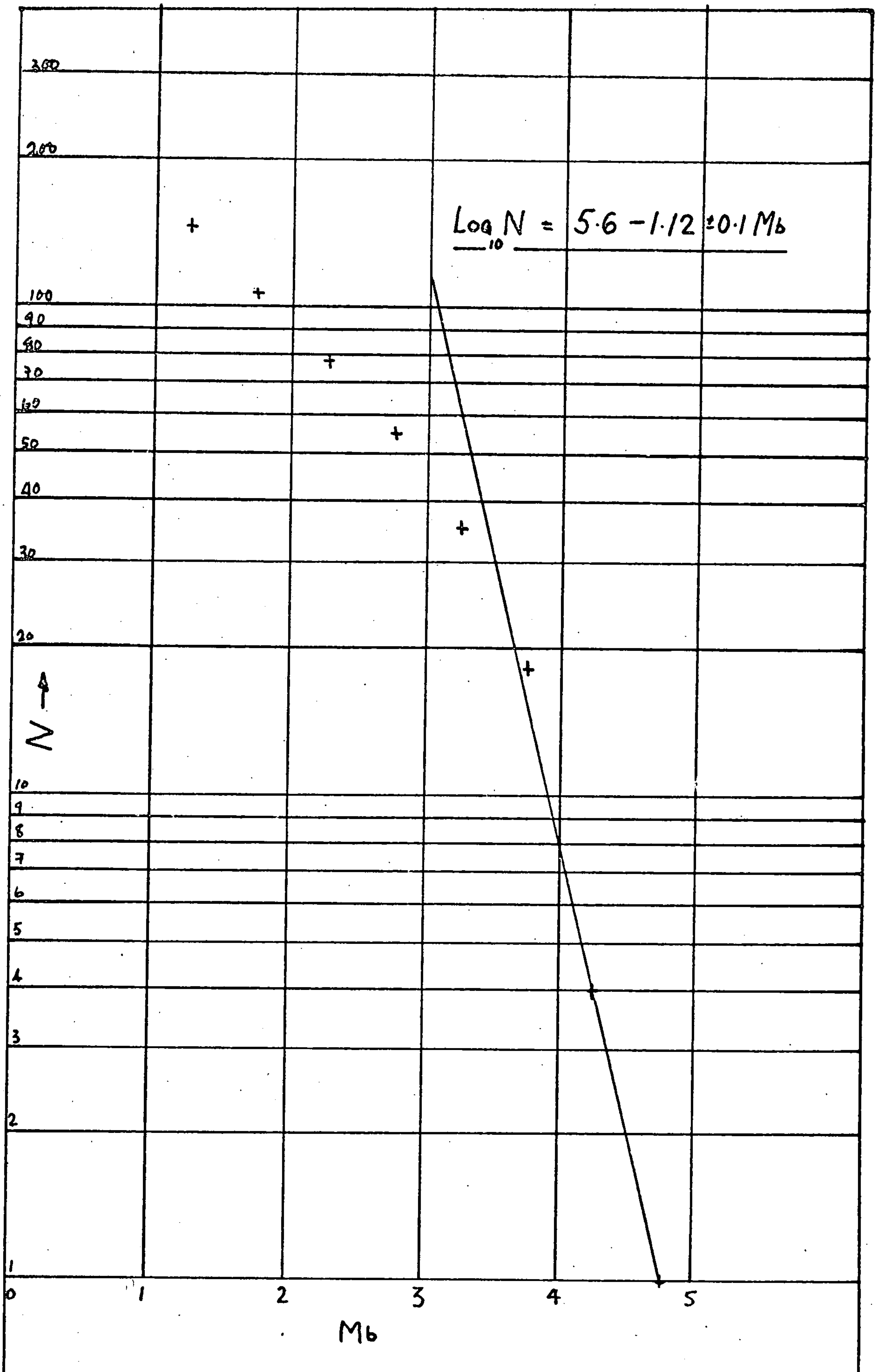
$$\text{Log } N = a - bM$$

where N is the number of events of magnitude M or greater (Gutenberg and Richter, 1954). The value b depends on the magnitude scale used, but throughout this work the magnitude determined has been Mb, the body wave magnitude.

No correlation of the section of the magnitude scale used for East Africa that refers to events with small epicentral distance, that is those occurring near to the Kaptagat array has been possible.

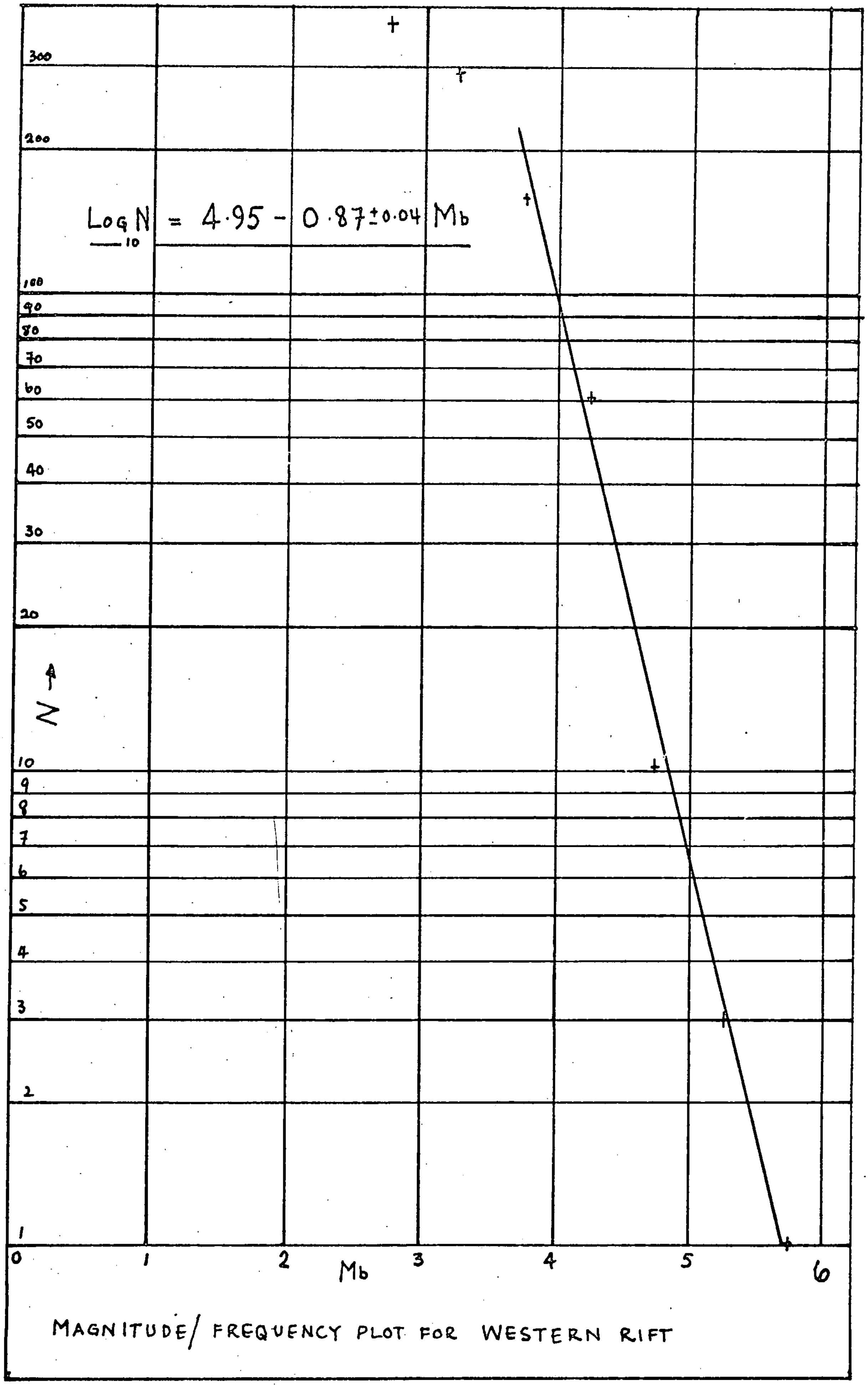
The possibility exists therefore that incorrect magnitudes have been assigned to events within this area. If this has been the case then the value of a in the above relationship will most certainly be incorrect. It is thought unlikely, however, that the value of b will be significantly altered because the relative magnitude distribution will not alter. Only the magnitudes in the magnitude/ frequency plot will therefore alter leaving the value of b, the slope of the graph, constant.

Magnitude/frequency plots have been made for the four areas under consideration together with a plot of all events that occurred within East Africa. It is recognised that the Kaptagat results represent a period of recording of only one year. In order to incorporate events of a larger magnitude, that are known to occur,



MAGNITUDE / FREQUENCY PLOT FOR EASTERN RIFT

fig.28.



$\underline{\log_{10} N = 4.95 - 0.87 \pm 0.04\ M_b}$
300
200
100
90
80
70
60
50
40
30
20
N
10
9
8
7
6
5
4
3
2
1
0
1
2
Mb
3
4
5
6
MAGNITUDE/ FREQUENCY PLOT FOR WESTERN RIFT

fig. 29.

the yearly average of events of magnitude greater than Mb=4.7, occurring within East Africa, reported by Fairhead and Girdler (in press), was added to the total events plot for East Africa given by the present study.

The following b values were obtained from these plots:-

(i)	Eastern rift	1.12 ± 0.1
(ii)	Western rift	0.87 ± 0.04
(iii)	Kavirondo rift	0.71 ± 0.15
(iv)	Lake Victoria region	0.81 ± 0.13
(v)	East Africa – this study	0.63 ± 0.06
(vi)	East Africa – this study and larger events	0.86 ± 0.03

The value of b for East Africa of 0.86 is in good agreement with that calculated by Wohlenberg (1968) of 0.85 using earthquakes located in East Africa of magnitude Mb=4.0 and greater over the period 1958 to 1963, but a little lower than the value of 0.95 given by Fairhead and Girdler (in press) for East African events, over the period January 1963 until December 1970, with magnitudes greater than Mb=4.8.

The value of 0.87 ± 0.04 for western rift events can only be compared with the specific studies made by Lahr and Pomeroy (1970) who gave the value of b as 1.05 for 108 events associated with the 20 March 1966 earthquake. Fairhead and Girdler who studied 28 events from the same sequence, gave the value of b as 0.83. Whilst neither of these values can be considered in direct comparison with the more widespread area considered in this study, their values are sufficiently near to indicate a similar tectonic regime throughout the area.

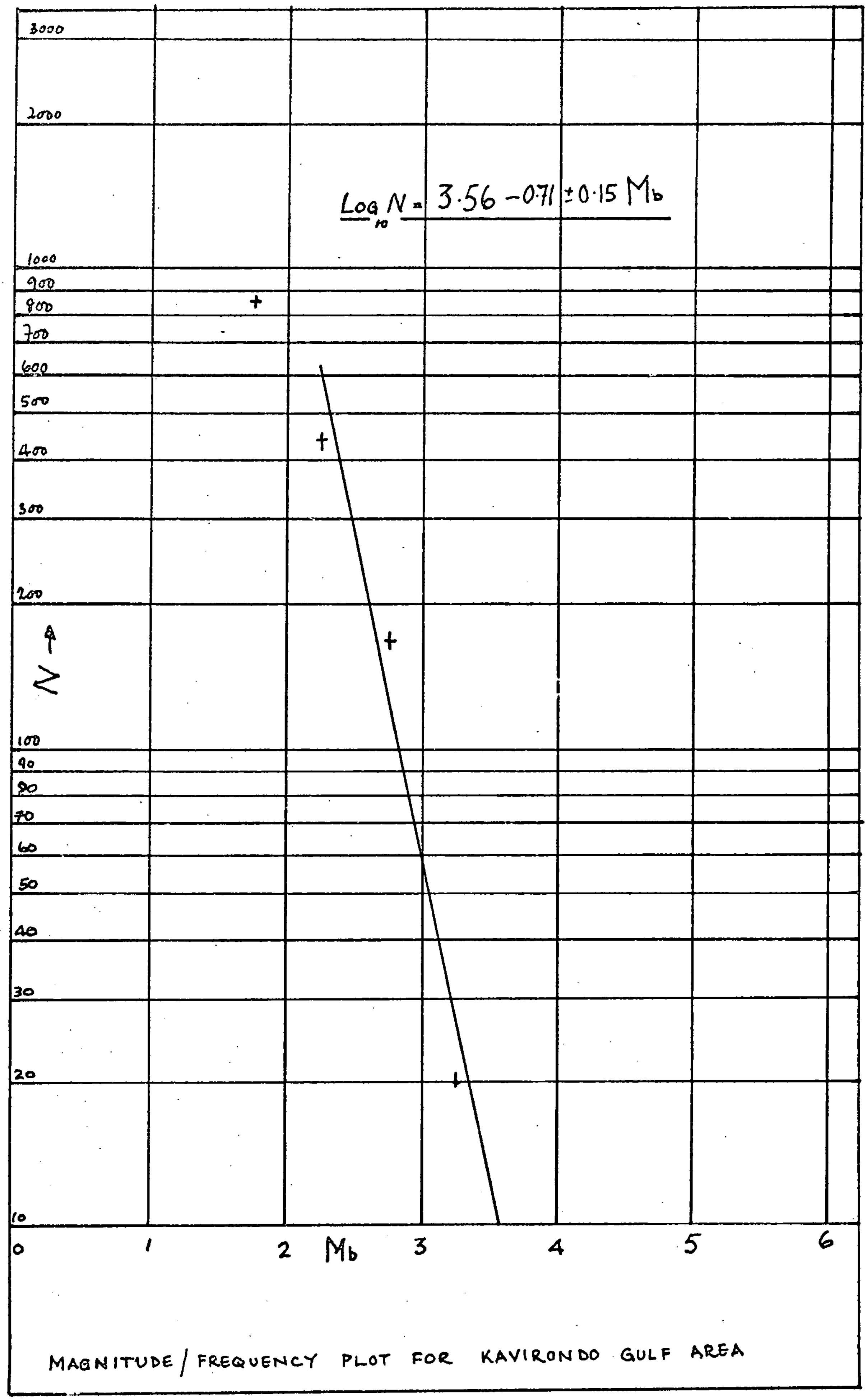

$\underline{\mathrm{Log}_{10} N = 3.56 - 0.71 \pm 0.15\ M_b}$
3000
2000
1000
900
800
700
600
500
400
300
200
N
100
90
80
70
60
50
40
30
20
10
0
1
2
Mb
3
4
5
6
MAGNITUDE / FREQUENCY PLOT FOR KAVIRONDO GULF AREA

fig. 30

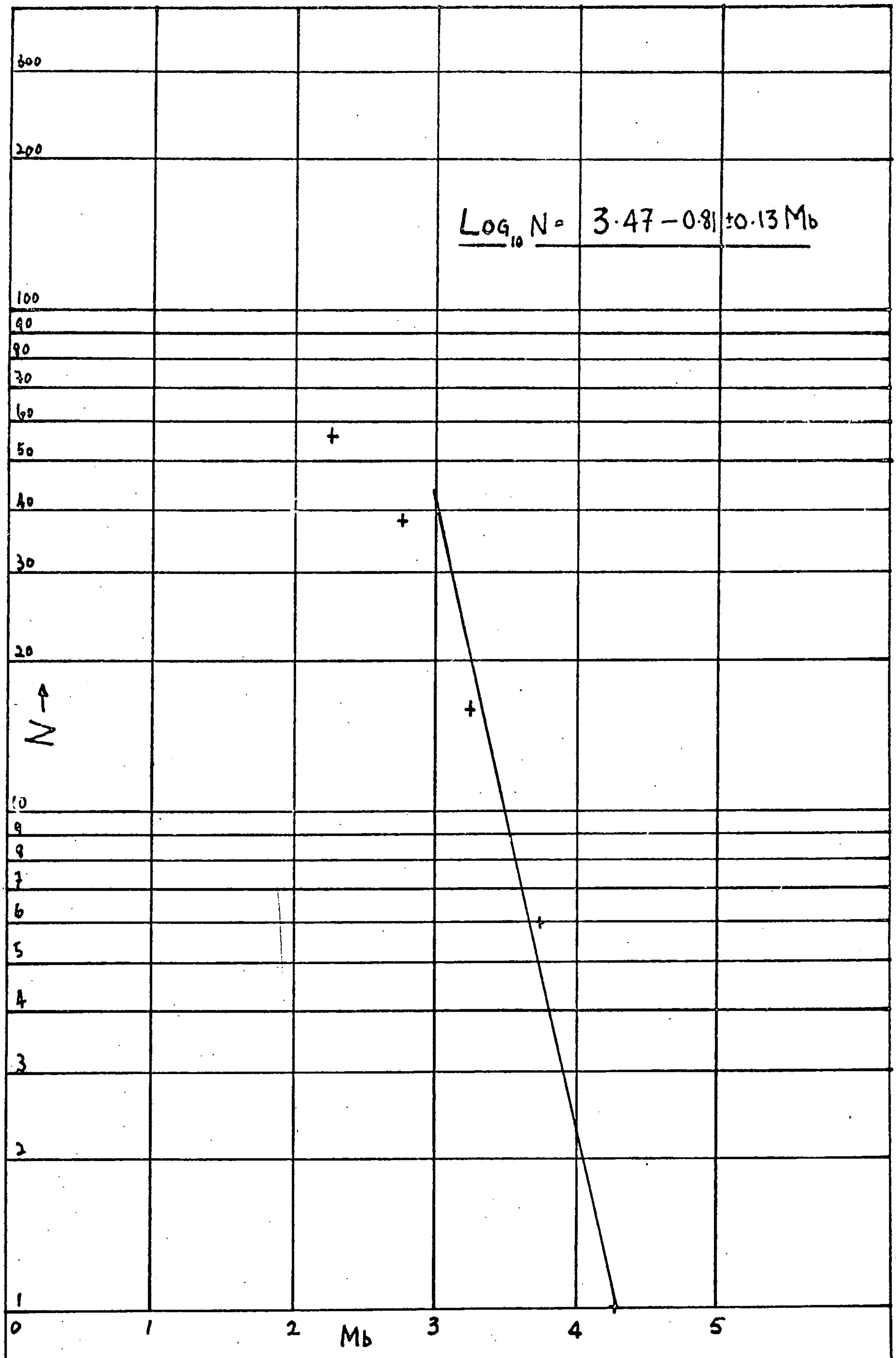
There is a significant difference between the values of b obtained for other areas within East Africa and that obtained for the eastern rift. The value of 1.12 ± 0.1 for the eastern rift is markedly higher than the average value of 0.8 for all other areas. This is thought to reflect the difference between normal shield areas and the eastern rift, which has been shown to be underlain by anomalous material (Griffiths et.al., 1971, Khan and Mansfield, 1971; Long et al., in press). It is also suggested that the slightly higher than average value of b=0.87 for the western rift may well reflect the similar but reduced anomaly reported by Wohlenberg (1970) who found the P-wave velocities at Lwiro, from western rift earthquakes to be lower than those coming from the nearby shield structure.

Table 25 lists the values of b derived for various parts of the world. It can be seen from the values given in this table that the western rift b value of 0.87 ± 0.04 is very similar to the value of 0.88 for southern California and 0.87 for New Zealand. Both of these areas are the location of very active faults but are not areas of crustal spreading. On the other hand, the higher values for the Indian Ocean of 1.3 and the Atlantic ocean of 1.4 and 1.7 (Francis, 1968) where sea floor spreading is known to be happening correlate well with the value for the eastern rift of 1.12. The eastern rift exhibits slower separation (Osmaston, 1971) than mid oceanic ridges. The reduced value of b is therefore consistant with the eastern rift being an area of crustal separation. Francis (1968) calculated the values of b for earthquakes located along the median rift of the Mid-Atlantic Ridge and found that there was a significant difference between these and the b values of these events occurring along the fracture zones which offset the ridge. He found that the effects

Figure 27

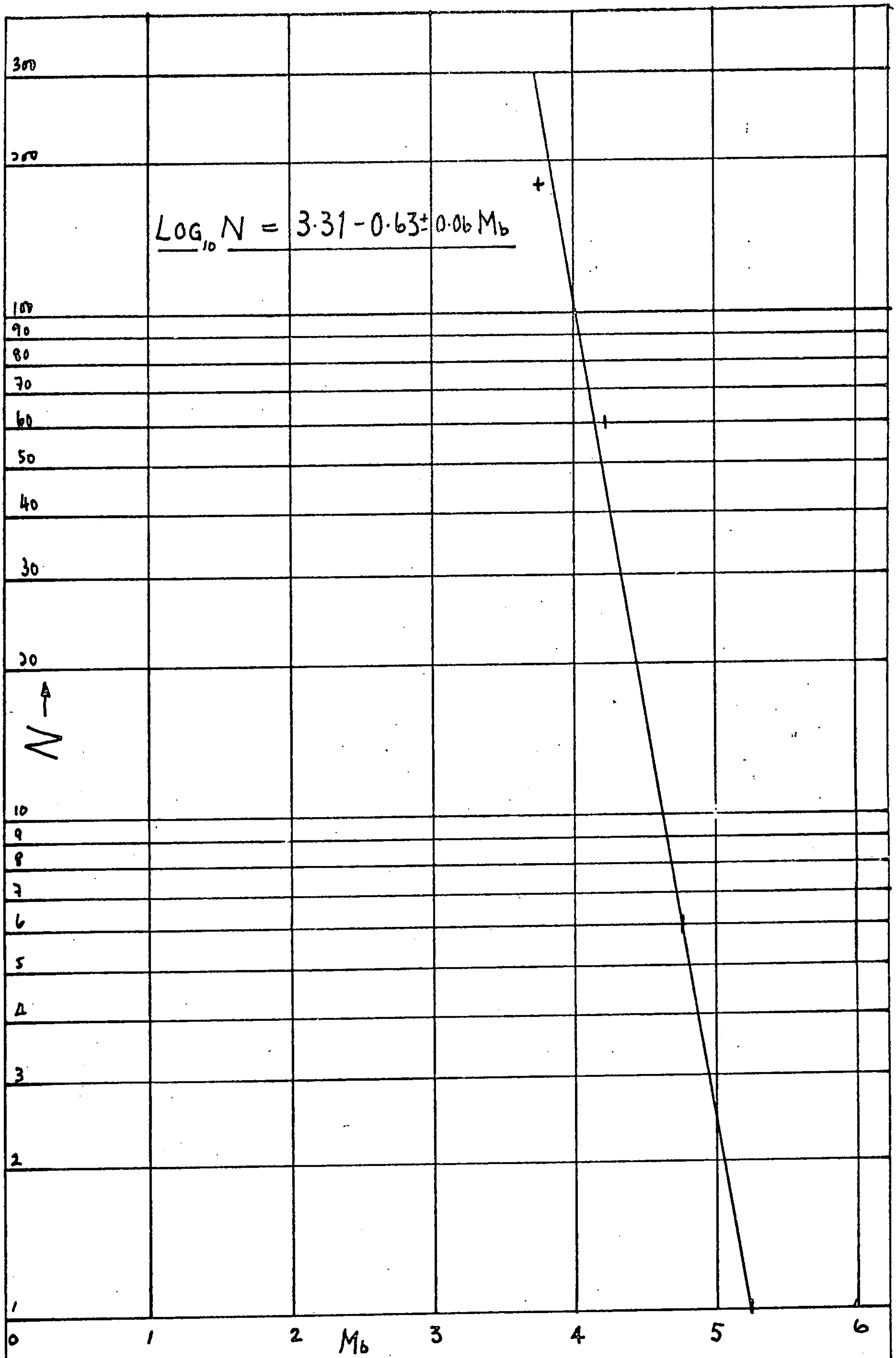
Values of b for the relationship log N= a - bM for various areas of Mid Oceanic Ridges (Francis, 1968)

Fracture zone	b	Rift Zone	b
Romanche	1.06	45^{o}-48^{o}N	1.10
St. Pauls	1.55	50^{o}-55^{o}N	1.47
'Z'	0.69	55^{o}-57^{o}N	2.63
53^{o}	0.81		
Mean	1.03 ± 0.33 SD		1.73 ± 0.65SD



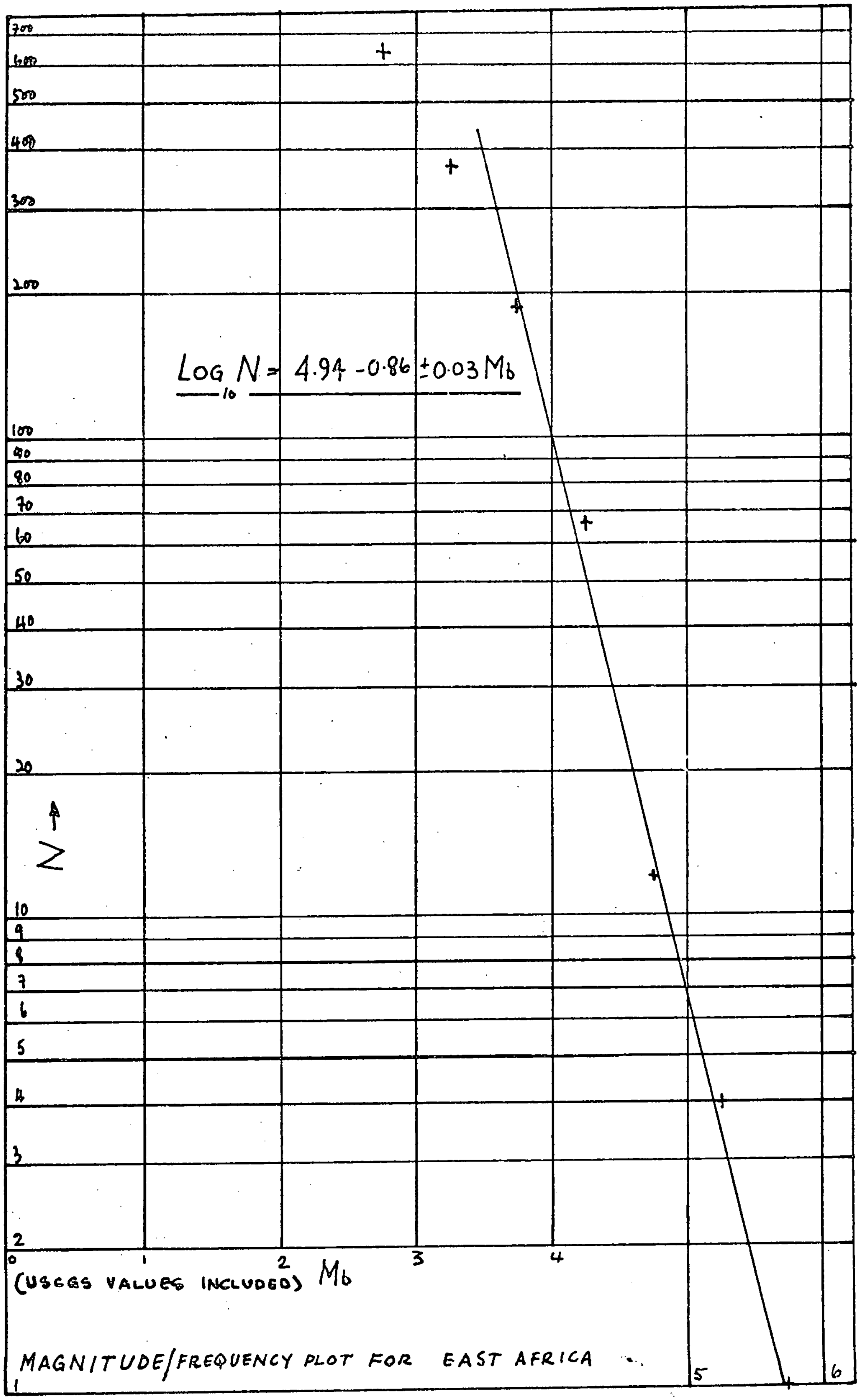
MAGNITUDE / FREQUENCY PLOT FOR LAKE VICTORIA REGION

fig. 31.



MAGNITUDE / FREQUENCY PLOT FOR EAST AFRICA (THIS STUDY)

fig. 32.



700
600
500
400
300
200
100
90
80
70
60
50
40
30
20
10
9
8
7
6
5
4
3
2
1
N
0
1
2
3
4
5
6
Log$_{10}$ N = 4.94 -0.86 ±0.03 Mb
(USCGS VALUES INCLUDED) Mb
MAGNITUDE/FREQUENCY PLOT FOR EAST AFRICA

fig. 33

TABLE 25

Values of b in the relationship Log N = a - bM for various parts of the world for earthquakes less than 100 Km deep.

Location	Value of 'b'
Alaska	1.1 ± 0.1
S. California	0.88 ± 0.03
Mexico	0.9 ± 0.1
S. America	0.45 ± 0.1
New Zealand	0.87 ± 0.04
Kermadec Islands	1.3 ± 0.2
Solomon Islands	1.01 ± 0.07
Japan	0.8 ± 0.08
Eastern Asia	0.6 ± 0.14
Turkey	0.9 ± 0.1
Indian ocean	1.3 ± 0.1
Atlantic	1.4 ± 0.2
Mean value for shallow earthquakes	0.9 ± 0.02

Reference:- Gutenberg and Richter 'Seismicity of the Earth' 2nd. ed. 1954 Princeton University Press.

of the higher heat flow under the ridge of 1.48 cal/cm^2 compared with that for the offset fracture zone of 1.13 cal/cm^2 confined the accumulation of strain under the ridge to shallow depths and that the higher temperature also set an upper limit to the magnitude of the earthquakes from that zone.

The value of b for the eastern rift shows that within the magnitudes recorded for that area there was a higher percentage of low magnitude events than was recorded from elsewhere in East Africa. This pattern of earthquake magnitude occurrance may be indicative of the strength of the crust in this area. The smaller magnitude earthquakes recorded from this area may be the direct result of a limitation of stress build up before material failure and the consequent earthquake occurrance.

The eastern rift is a zone of crustal separation and magmatic intrusion (Griffiths et.al, 1971; Khan and Mansfield, 1971; Baker and Wohlenberg, 1971). The crustal attenuation and associated high heat flow regime (Baker and Wohlenberg, 1971) may have the effect of reducing the shear strength and increasing the plasticity of crustal material thereby limiting the build up of strain before failure.

The western rift, on the other hand, exhibited energy release in large bursts of activity associated with large magnitude events, see figure 20, whereas the eastern rift activity remained at a virtually constant level throughout the period of recording, see figure 19. The b value for the western rift magnitude/frequency plot was very similar to that for the Lake Victoria area where normal crustal conditions may exist (Long et.al., in press). This was taken to indicate that the western rift exhibited material failure of more elastic type with the associated accumulation of much higher stresses giving rise to the periodic bursts of energy release already noted, see figure 20.

From this correlation it would appear that the western rift is only in the very early stages of its evolution and that little or no crustal separation or magmatic intrusion has yet taken place there.

The conspicuous lack within the western rift area of widespread vulcanicity such as that seen both within the eastern rift and on its flanks is thought to be indicative of the greater maturity of the eastern rift. It is therefore considered possible that the western rift, during some period in the future, will exhibit a more widespread vulcanicity than it currently does.

CHAPTER FIVE

CONCLUSIONS

Earthquakes that occurred within an epicentral distance of 600 Km of Kaptagat, Kenya, have been located for the inclusive period July 1970 until May 1971 and the accuracy of their preliminary location given. It has been noted that the majority of earthquakes recorded occurred along or near the major tectonic features of the East African rifts.

The areas from which the largest numbers of earthquakes have been recorded were the western rift between latitudes $1^{o}S$ and $1^{o}N$ and the Kavirondo Gulf rift where the events were generally confined to the rift walls.

In general, because of the lack of details of crustal thickness, it has not been possible to assign focal depths to events located within this area. The focal depth of events occurring within 40 Km of Kaptagat have, however, been calculated and give a mean value of about 24 Km.

Magnitudes have been calculated for all events located, using the Kaptagat array data and a preliminary magnitude scale. The accuracy of this scale will become better known when data from the other Durham stations, as yet not worked upon, is incorporated.

The preliminary data available suggests from magnitude/frequency plots that there are substantial differences between the type of energy release in the eastern and western rifts.

Evidence with respect to plate tectonics suggests almost normal crustal behaviour of stress release from the western rift.

The western rift is, therefore, seen as exhibiting a very early stage in rift valley evolution and the lack of associated widespread volcanics seen as confirmation.

The eastern rift, on the other hand, has occasioned a higher percentage of smaller events within the magnitudes recorded, than has any other area in East Africa. This is consistant with a tendency towards the more plastic floor with limited stress build up associated with rift floors spreading and consequently continental drift.

The value of b in the relationship

$$\text{Log } N = a - bM$$

for the eastern rift is consistant with the hypothesis that this region is the location of limited crustal spreading.

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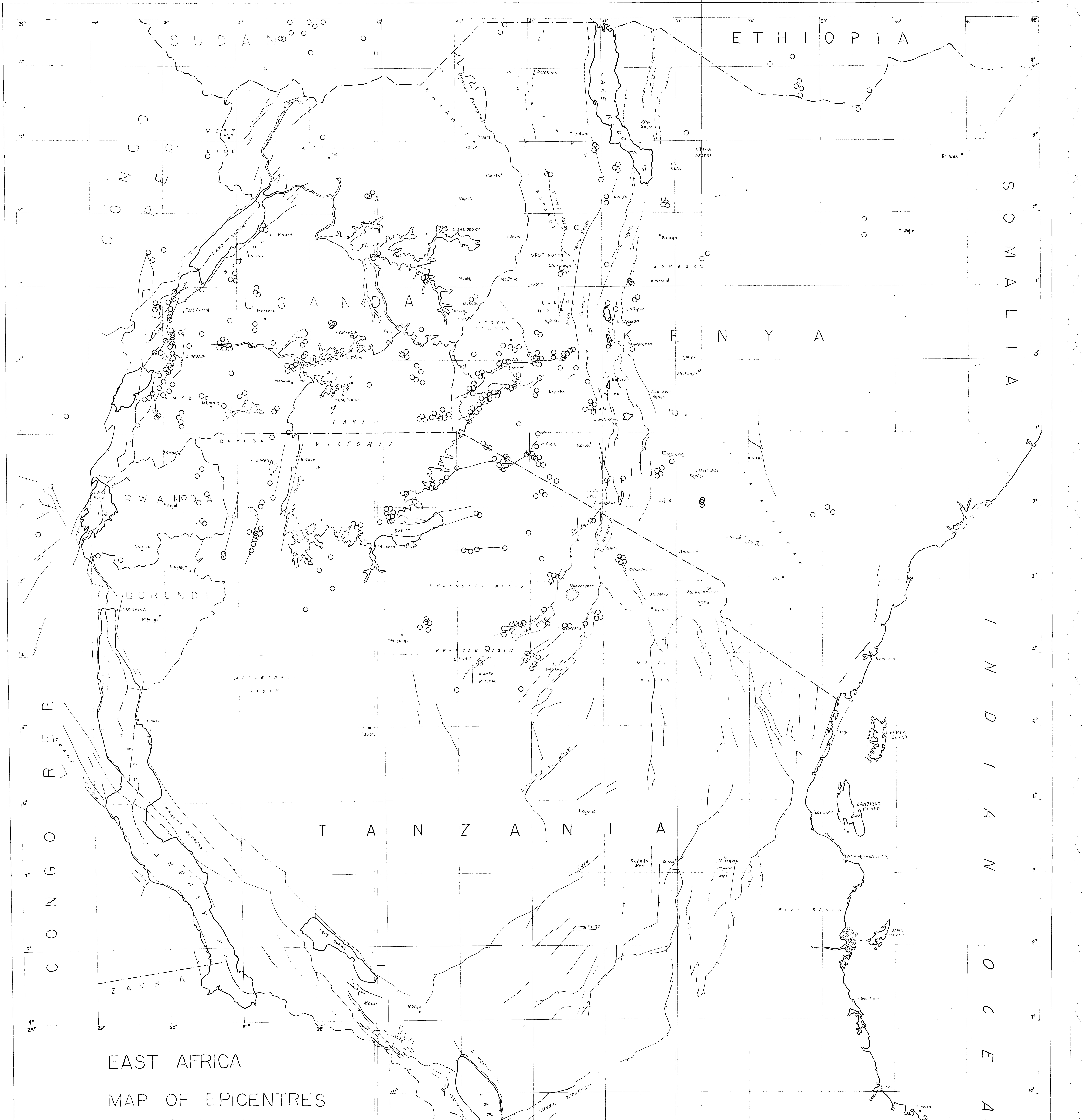
EAST AFRICA
MAP OF EPICENTRES
(JULY 1970 - MAY 1971)
COMPILED BY R.R. GILL, 1972.
faults mapped
faults inferred
Calderas
Earthquake epicentres
SCALE 1:2,000,000
25 0 25 50 75 100 125 150 175 200 Kilometers
SUDAN
ETHIOPIA
SOMALIA
CONGO REP.
UGANDA
KENYA
TANZANIA
RWANDA
BURUNDI
ZAMBIA
MALAWI
MOZAMBIQUE
INDIAN OCEAN
LAKE RUDOLF
LAKE ALBERT
LAKE VICTORIA
LAKE KIVU
LAKE TANGANYIKA
LAKE RUKWA
LAKE MALAWI
LAKE EYASI
SERENGETI PLAIN
KAMPALA
NAIROBI
DAR-ES-SALAAM
ZANZIBAR ISLAND
PEMBA ISLAND
MAFIA ISLAND
Tabora
Dodoma
Mbeya
Tanga
Mombasa
Fort Portal
Entebbe
Kisumu
Nakuru
Kericho
Mt. Kenya
Mt. Kilimanjaro
Usumbura
Goma
Kigoma
Mwanza
Bukoba
Arusha
Iringa
Songea
Mzuzu
Nkhata Bay
Kisii
Moroto
Lodwar
Marsabit
Garissa
Mtwara
Kilwa Kivinje

Fig. 14.