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Earthquake Fault Parameters and Tectonics in Africa[1]

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Fault plane solutions of earthquakes in southern Africa indicate that the least compressive stress is oriented approximately E-W nearly parallel to that in the northern part of the rift system. Seismic moments, source dimensions, and stress drops were determined for eight earthquakes from body- and surface-wave spectra by using the theory of Brune (1970). Spectral estimates of these quantities for the 1966 earthquake in the Republic of Zaire agree well with those observed in the field. Relatively higher stress drops are found for events not associated with rift faulting. If higher stress drops indicate higher rock strength, these higher stress drops, with other geological and geophysical data, suggest that the northern part of the rift system is similar to ocean ridges and behaves as a plate boundary but that the southern part is different and is not a plate boundary. The tectonics associated with the northern part of the rift system appear to be extending southward.

Although the east African rift system extends from Ethiopia to Malawi, it is convenient to consider two distinct branches, the eastern and western rifts (Figure 1). Both rifts are characterized by continuous belts of normal faulting and graben structures and the occurrence of shallow earthquakes. They differ from each other in important ways. The eastern rift is thought to be contemporaneous with the Gulf of Aden and the Red Sea [e.g., *Mohr*, 1967, 1970*a*, *b*] and is often assumed to be an extension of the ocean ridge system [*Ewing and Heezen*, 1956; *Heezen*, 1960; *Rothé*, 1954], whereas much of the western rift appears to be older and is often assumed to be unrelated to the ocean ridge system [*Dixey*, 1956; *McConnell*, 1967]. In this paper, evidence is presented from fault plane solutions and source parameters of the larger earthquakes that occurred in Africa since the WWSSN (World-Wide Standardized Seismograph Network) was installed, and further differences between the two branches of the east African rift system are discussed. A case is made for the structure and tectonics of the northern part of the eastern rift being similar to ocean ridges but with a very slow spreading rate and therefore being perhaps in an infant stage of spreading. The western rift may owe its origin to other phenomena, and at present it does not appear to be a plate boundary. It is further shown that the rift system is extending southward beyond the mergence of the two branches of the rift.

GEOLOGY AND TECTONICS

The eastern rift crosses two large domal uplifts 2200 meters high, one in Ethiopia and the other in Kenya, as a simple graben 80 km wide in the north and 65 km wide in the south [*Baker and Wohlenberg*, 1971; *Mohr*, 1967]. Between the two domal uplifts and at the southern end, elevations of the graben are lower, the faulting splays out, and no simple graben exists. Although regions uplifted above the surrounding areas do exist near Lake Kivu and near the Rungwa volcanics (north of Lake Malawi), broad domal uplifts like those in Kenya and Ethiopia are not characteristic of the western rift. The elevation decreases ab-

[1] Lamont-Doherty Geological Observatory contribution 1848.

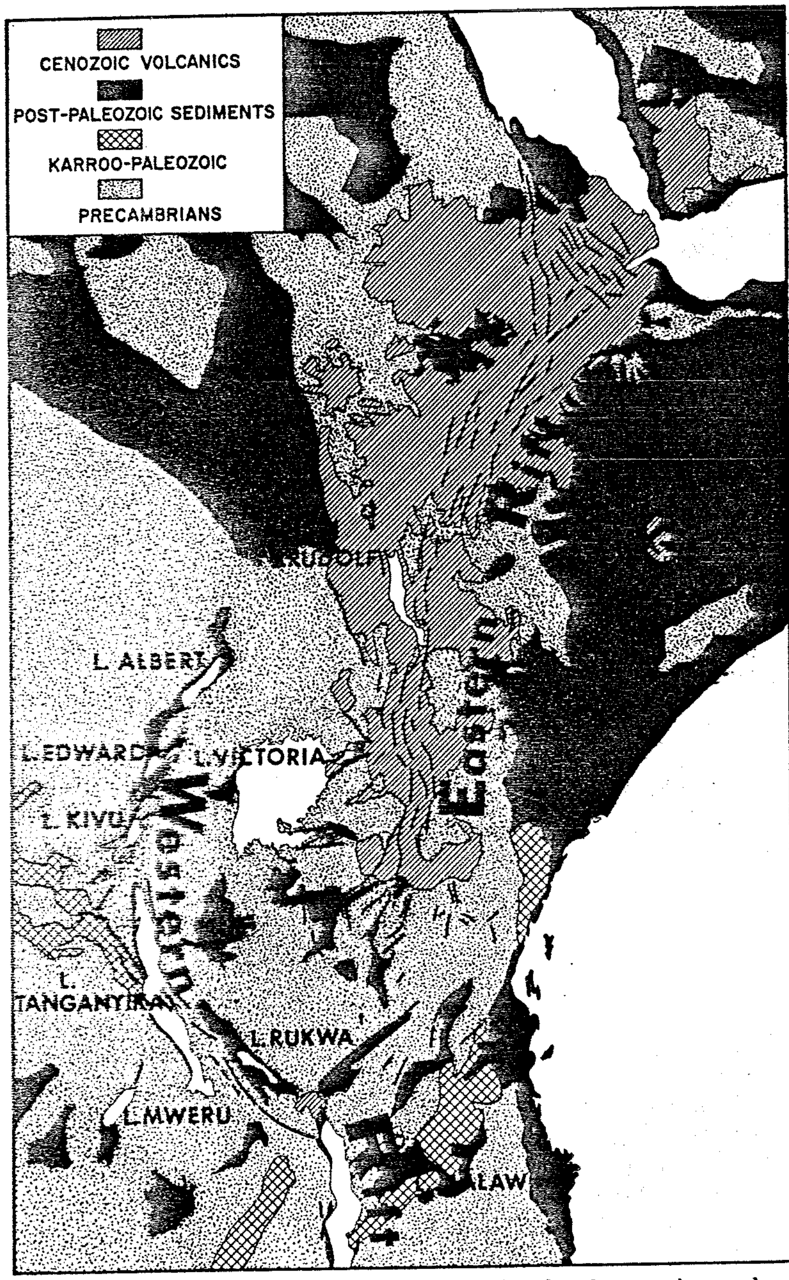

Fig. 1. Simplified geologic map of East Africa. Note the abundant tertiary volcanics in the eastern rift.

ruptly west of the rift into the Congo basin.

The extent of volcanism provides another striking contrast between the eastern and western rifts (Figure 1). The eastern rift experienced several stages of volcanism since the Eocene in Ethiopia and the Miocene in Kenya [e.g., *Baker et al.*, 1971; *Baker and Wohlenberg*, 1971; *Mohr*, 1967]. Except for the western rift, which has volcanic activity only near Lake Kivu and north of Lake Malawi, the entire rift zone and the surrounding plateau are covered by Tertiary volcanics.

The volcanism in east Africa is noted for being highly alkaline and therefore different from most volcanism emplaced at ocean ridges [e.g., *Gass,* 1970; *Harris,* 1969; *Mohr,* 1971; *Williams,* 1969]. *Gass* [1970] and *Green* [1970], among others, attributed this difference to the depths of fractionation beneath the ocean ridges and the rift system. They consider the rifts (eastern and western) to represent an early stage of continental breakup, albeit at a very slow rate [*Baker and Wohlenberg,* 1971].

Volcanism began in Ethiopia during the Eocene and in Kenya during the Miocene, but in both regions faulting and the formation of the graben occurred later, in the Miocene and in the Pliocene, respectively [*Baker and Wohlenberg,* 1971; *Mohr,* 1967]. The western rift is more difficult to date; *McConnell* [1967] suggests that faulting parallel to the present rift began as early as the Precambrian, and *Dixey* [1956] considers that well-defined grabens existed in the Jurassic. *Bloomfield and Habgood* [1966] and *Dundas* [1966] cite evidence in support of Mesozoic formation of the rift in Malawi and in Tanzania but with a significant rejuvenation in the later Tertiary. *Bishop and Trendall* [1967] find that in western Uganda the drainage changed and that sedimentation in the present rift system began in the Miocene. As in the eastern rift, faulting in the western rift was extensive in the Pleistocene. The volcanics of the western rift are all Miocene or younger. Although the formation of the two branches of the rift system may have begun at different times, both branches experienced significant tectonic activity in the late Cenozoic.

Seismicity and gravity data imply a difference in upper-mantle structure beneath the eastern and western rifts, particularly if the younger, northern part of the western rift is included with the eastern rift. Phase and group velocities of surface waves and travel times of Pn, Sn, and Lg indicate that the structure beneath most of Africa is typical of most stable continental regions [*Gumper and Pomeroy,* 1970]. Travel times for the region between the two branches of the rift system near Lake Victoria are also indicative of a stable region [*Rodrigues,* 1970; *Wohlenberg,* 1968].

A refraction survey [*Griffiths et al.,* 1971] in the eastern rift in Kenya, however, revealed a low Pn velocity (7.5 km/sec) and a thin crust (20 km). Travel times from well-located events indicate relatively low velocities beneath the eastern rift and the northern part of the western rift (i.e., north of Lake Tanganyika) [*Fairhead and Girdler,* 1971; *Rodrigues,* 1970; *Wohlenberg,* 1968]. Teleseismic *P* wave residuals at stations in Africa are latest (i.e., slowest) for stations near this part of the rift [*Fairhead and Girdler,* 1971; *Rodrigues,* 1970]. *Rodrigues* [1970] found that at NAI and LWI (Figure 2) the residuals were largest for paths from the north. *Bonjer et al.* [1970] showed that most of the delay occurred in the mantle. Thus, abundant data indicate that a zone of low-velocity mantle material similar to that at ocean ridges underlies the eastern rift and the northern part of the western rift. Data available do not suggest low-velocity material beneath the rest of the western rift.

Gross lateral variations in the character of Sn have been used to infer the presence of low-Q material beneath ocean ridges [*Molnar and Oliver,* 1969]. That Sn is highly attenuated for paths crossing the northern part of the rift system but not for paths that are distant from the rift system or that cross only the southern part of it [*Gumper and Pomeroy,* 1970; *Molnar and Oliver,* 1969; *Rodrigues,* 1970; *Searle and Gouin,* 1971], implies a discontinuity in the mantle lithosphere beneath only the northern part of the eastern and western rift system, not the southern part.

Studies of gravity anomalies indicate a long-wavelength Bouguer minimum over the eastern rift system in Ethiopia [*Gouin,* 1970], Kenya [*Baker and Wohlenberg,* 1971; *Girdler et al.,* 1970; *Khan and Mansfield,* 1971], and Tanzania [*Sowerbutts,* 1969]. This minimum is interpreted as evidence for a low-density zone and a thinning of the lithosphere beneath the rift. The less pronounced long-wavelength anomaly found over the northern part of the western rift [*Girdler et al.,* 1970; *Sowerbutts,* 1969] probably indicates low-density material there.

There is a small positive Bouguer anomaly centered over the rift in Ethiopia [*Makris et al.,* 1970; *Mohr and Gouin,* 1968] and Kenya [*Baker and Wohlenberg,* 1971; *Khan and Mansfield,* 1971; *Searle,* 1970*a*]. This anomaly is indicative of dense basic intrusive material into the crust to shallow depths in the east, and it may be a measure of separation of opposite sides of the rift [*Baker and Wohlenberg,* 1971; *Searle,* 1970*a, b*]. No such positive anomaly

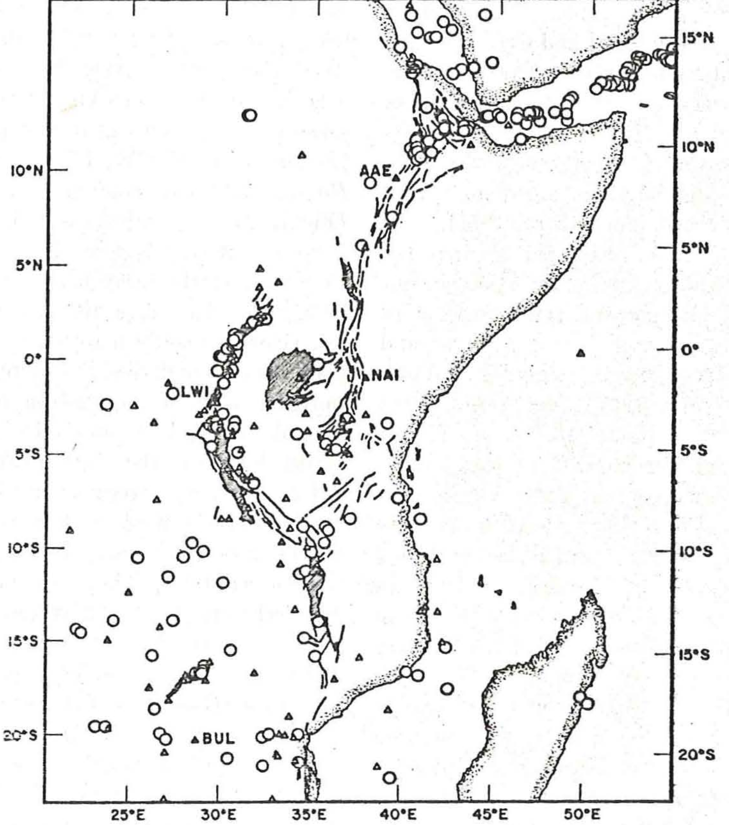

Fig. 2. Seismicity of East Africa from 1950 to 1969. Epicenters for 1950 to 1966 are from *Sykes and Landisman* [1964] and *Sykes* [1970]. More recent epicenters were located by the U.S. Coast and Geodetic Survey (NOAA). Circles are better locations and triangles are poorer locations. Three-letter codes denote seismograph stations.

is found over the western rift, and G. H. Sutton and J. A. Grow (unpublished manuscript, 1972) do not consider that the gravity data show any variation in the depth to the Mohorovicic discontinuity or such an intrusion beneath this rift.

The seismicity of Africa also indicates differences between the eastern and western rifts. Figure 2 shows epicenters relocated by *Sykes and Landisman* [1964] and *Sykes* [1970] for the period 1950–1966 and by the U.S. Coast and Geodetic Survey (NOAA) for 1967–1969. For studies by local stations, depths of focus are shallow [*Molnar and Aggarwal*, 1971; *Wohlenberg*, 1968]. Two important phenomena are illustrated. First, the activity in southern Africa is high in spite of the absence of active rifts [*Sykes*, 1970]. Although many earthquakes occur near Lakes Albert, Kivu, and Tanganyika, the southern part of the western rift is not associated with a well-defined belt of seismicity [*Fairhead and Girdler*, 1971]. Second, the activity in the eastern rift is very low [*Gouin*, 1970; *Wohlenberg*, 1968]. This pattern is observed also for microearthquakes [*Molnar and Aggarwal*, 1971; *Molnar et al.*, 1970; *Tobin et al.*, 1969], although this low activity might result in part from the short time sample of data. It is also possible that much of the deformation occurs aseismically [*Molnar et al.*, 1970]. Recent faulting that is confined to the rift attests to recent tectonic activity along a relatively narrow zone.

Thus, in many ways the eastern rift differs from the western rift and is similar to the ocean ridges. The volcanism, seismic velocities and attenuation, gravity anomalies, and seismicity suggest or are consistent with the presence of discontinuities of the lithosphere mantle beneath the eastern rift and the northern part of the western rift but not elsewhere. An application of plate tectonics to the Red Sea and Gulf of Aden enabled workers to estimate quantitatively the separation of the eastern rift [*McKenzie et al.*, 1970]. Although there has been criticism of this study [*Freund*, 1970; *Mohr*, 1970*b*] and there is uncertainty in the amounts of movement, some extension along the eastern rift is demanded by the directions of opening in the Red Sea and the Gulf of Aden. The similarity in ages of these features also implies that movements in all three regions are interrelated.

FAULT PLANE SOLUTIONS

Because of the lack of numerous large earthquakes in Africa, it is difficult to obtain many fault plane solutions. Nevertheless, 11 solutions were obtained (Table 1). Figure 3 shows schematic lower hemisphere projections of 10 of them and their locations. Two solutions in the Afar depression indicate strike-slip faulting [*McKenzie et al.*, 1970]. The NW-SE nodal plane is consistent with the surface faulting observed and is therefore the more likely fault plane. Solutions were not obtained for any other earthquakes in the main portion of the eastern rift. A large earthquake in 1928 occurred in the rift in Kenya and was accompanied by surface deformation [*Richter*, 1958]. The field evidence indicated a large component of normal faulting, but some strike-slip motion cannot be eliminated.

Solutions for the large earthquake in the Republic of Zaire in 1966 [*Sykes*, 1967] and for an aftershock [*Banghar and Sykes*, 1969] indicate normal faulting, the least compressive stress being oriented approximately perpendicular to the rift. The surface deformation for the main shock was consistent with Sykes's solution. The abundant normal faulting in the rift system implies that these solutions are typical for earthquakes occurring there.

The solution for the 1964 shock in Tanzania (Figure 3, event 5, and Figure 4*b*) is difficult to interpret. The steeply dipping plane is not parallel to the northerly striking faults in this region. The strike of the other plane is not well determined, but it also is unusual because of its very shallow dip. Moreover, the solution contains a component of thrust faulting. This earthquake is not easily reconciled with other information indicating extension along the rift.

TABLE 1. Locations and Fault Plane Solutions of Eleven African Earthquakes

Event No.	Date	Location	*P* Axis, deg		*T* Axis, deg		*B* Axis, deg		Pole 1, deg		Pole 2, deg	
			Dip	Trend	Dip	Trend	Dip	Trend	Dip	Trend	Dip	Trend
1	March 29, 1969	12.1°N, 41.1°E	14	286	12	20	70	153	27	340	90	243
2	April 5, 1969	12.1°N, 41.1°E	14	286	12	20	70	153	27	340	90	243
3	March 20, 1966	0.6°N, 30.2°E	90	0	6	117	10	207	50	104	52	302
4	May 17, 1966	0.7°N, 30.1°E	69	93	22	270	0	2	22	92	68	272
5	May 7, 1964	4.0°S, 34.9°E	35	21	49	220	14	122	14	214	73	340
6	May 6, 1966	15.7°S, 34.4°E	75	90	15	259	4	349	31	80	59	254
7	Dec. 2, 1968	13.9°S, 23.8°E	72	116	18	296	0	22	26	112	62	292
8	May 15, 1968	15.8°S, 25.9°E	65	181	10	298	23	33	50	272	32	136
9	Sept. 25, 1963	16.7°S, 28.7°E	10	283	75	138	13	19	58	270	30	115
10	Sept. 23, 1963	16.7°S, 28.7°E	10	283	75	138	13	19	58	270	30	115
11	Sept. 29, 1969	32.9°S, 19.7°E	10	90	6	181	74	353	8	134	9	223

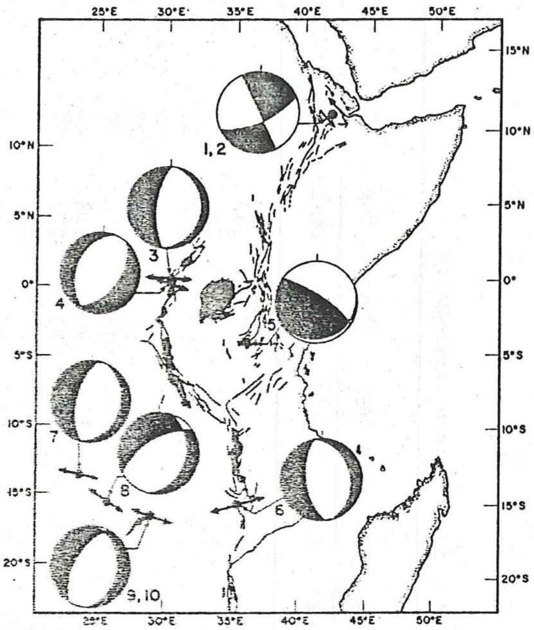

Fig. 3. Fault plane solutions in East Africa. Plots are lower hemisphere projections. Darkened areas denote compressional readings. Arrows indicate the sense of strike-slip motion or direction of least compressive stress. Numbers correspond to those in Table 1.

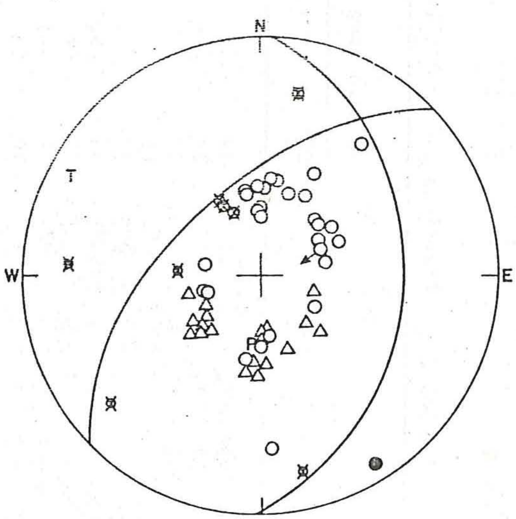

Fig. 4a. Lower hemisphere equal-area plots of data for new fault plane solutions. Large symbols are more reliable first motions. Solid symbols are compressions. Open symbols are dilatations. Triangles are upper hemisphere (pP) data. Arrows give direction of first motion of S waves. P and T give compressional and tensional axes. Cross denotes emergent signal. Data are for May 15, 1968, location is 15.9°S, 25.9°E, and distance is 30 km. This is event 11.

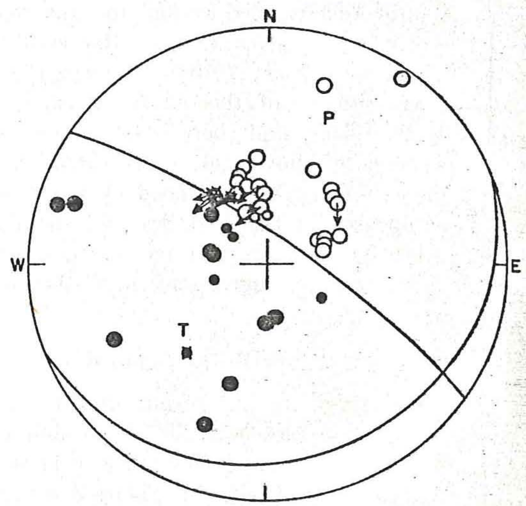

Fig. 4b. Same as Figure 4a except data are for May 7, 1964, location is 4.0°S, 34.9°E, and distance is 30 km. This is event 5.

The remaining solutions in Figure 3 indicate normal faulting, the least compressive stress being oriented approximately ESE-WNW. Only event 6, which has a poorly determined solution (Figure 4d) occurred in the rift system at the southern end of Lake Malawi. The remaining events 7, 8, 9, and 10 occurred well outside the rift (Figure 3). Two earthquakes that occurred near Lake Kariba appear to be related to the loading of the lake [*Gough and Gough*, 1970; *Sykes*, 1967]. All five events suggest that southern Africa is under ESE-WSW regional extension [*Fairhead and Girdler*, 1971]. These solutions are consistent with the idea that the lithosphere is still continuous in this region but that the directions of principal stress are similar to those that probably have existed in the eastern rift since the Miocene. This theory does not, however, rule out the possibility of the existence of a plate margin. These earthquakes reflect the earth's response to stresses that tend to extend southward the zone of active rifting now located farther north.

A fault plane solution was also determined for the Ceres earthquake in South Africa on September 29, 1969 (Figure 4e) that indicated strike-slip movement with the least compressive stress oriented approximately N-S. This earthquake may also have resulted from the same stress system that caused events 6–10, but the great distance between them casts doubts on

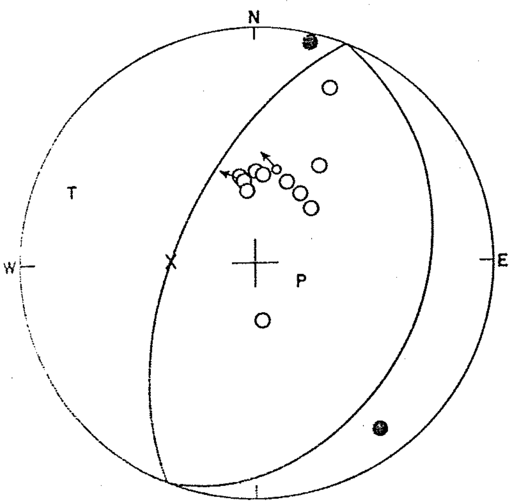

Fig. 4c. Same as Figure 4*a* except data are for December 2, 1968, location is 13.9°S, 23.8°E, and distance is 7 km. This is event 7.

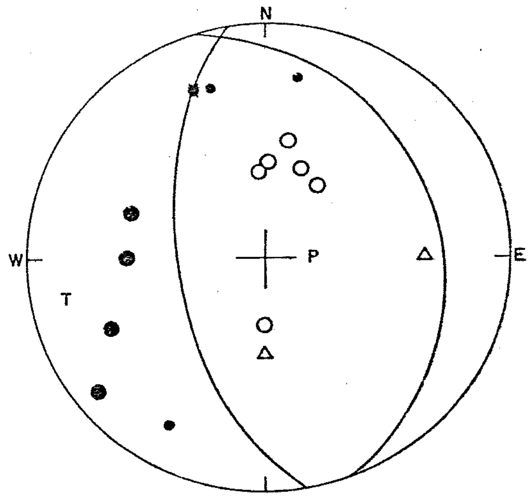

Fig. 4*d*. Same as Figure 4*a* except data are for May 6, 1966, location is 15.7°S, 34.4°E, and distance is 34 km. This is event 6.

this possibility. The solutions presented here are similar to those determined by *Fairhead and Girdler* [1971] for this region except for the new, additional event 6 and a difference in solution for event 5. The interpretations of some solutions also differ.

SOURCE PARAMETERS OF SOME EARTHQUAKES

Data analysis. Seismic moments, source dimensions, average dislocations, and stress drops were determined for eight earthquakes (Table 2) by using the theory of *Brune* [1970]. The earthquakes studied were those with several body waves clearly recorded between distances of 35° and 90° by long-period seismographs of the WWSSN. For the larger events Fourier amplitude spectra of surface waves were determined and corrected for geometric spreading by using tables from *Ben-Menahem et al.* [1970]. μ estimates of the seismic moment ($M_o = m)A\langle d\rangle$, were made, where A is the fault area and $\langle d\rangle$ the average displacement [*Aki*, 1966]. A shear modulus $\mu = 0.33 \times 10^{12}$ dynes/cm and density $\rho = 2.71$ g cm^{-3} were assumed throughout the analysis.

The analysis of body waves was essentially the same as that of *Hanks and Wyss* [1972], *Molnar and Wyss* [1972], and *Wyss and Hanks* [1972]. Long-period signals and short-period signals (where clearly recorded) for both P and S waves were Fourier analyzed.

At long periods the spectral amplitude was assumed to be constant. The value of the flat portion of the spectrum $\Omega(0)$ is proportional to the seismic moment $M_o = [4\pi\delta v^3 R\Omega(0)]/R(\sigma\phi)$ where δ and v are density and wave velocity at the source, $R(\sigma\phi)$ is the normalization factor for radiation pattern from *Ben-Menahem et al.* [1965], and R is the correction made for geometrical spreading by using the

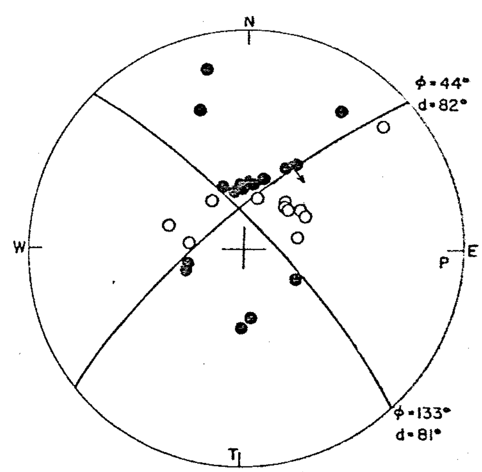

Fig. 4*e*. Same as Figure 4*a* except data are for September 29, 1969, location is 32.9°S, 19.7°E, and distance is 33 km. This is event 11.

TABLE 2. Source Parameters

(The number in parentheses beside a value indicates the number of stations used in computing the value.)

Event No.	Moment, × 10^{25} dynes cm					Corner Frequency, × 10^{-1} hz		Radii, km			Area, km^2	Dislocation, cm	Stress Drop, bars
	P	*S*	*LR*	*LQ*	Avg.	*P*	*S*	*P*	*S*	Avg.			
2	1.9(6)	1.1(8)			1.43	1.95(6)	1.1(8)	11.0	12.2	11.9	450	10	3.4
3	22.0(10)	17.0(7)	16.1(4)	16.0(4)	18.6	1.4(10)	0.81(7)	16.4	16.5	16.4	844	67	18.3
4	0.63(10)	0.75(3)			0.66	2.1(10)	1.6(3)	10.4	8.3	10.2	320	6	2.8
5	0.8(9)	2.5(7)	2.5(4)		4.8	2.0(9)	1.45(7)	10.6	9.0	9.6	290	50	23.5
7	0.58(7)				0.58	4.4(7)		4.9		4.9	75.0	23	17.7
8	2.0(10)	0.58(13)	0.85(2)		1.17	2.8(10)	1.7(13)	7.4	7.7	7.6	180	20	11.7
9	0.58(8)				0.58	3.0(8)		6.6		6.6	135	6	8.3
11	2.0(8)				2.0	2.8(8)		6.9		6.9	150	40	25.0

results of *Julian and Anderson* [1968]. The spectral amplitude is further divided by 2.5 to correct for an average crustal transfer function. Seismic moments from *P*, *S*, Rayleigh, and Love waves were averaged to give an estimate considered to be reliable within a factor of 2.

Each spectrum was corrected for attenuation by using the average *Q* values of *Julian and Anderson* [1968]; both *Q* corrected and uncorrected spectra were plotted. Although the *Q* correction affects the high frequency spectrum, the corner frequencies are not altered significantly. *Wyss and Molnar* [1972] showed that even large lateral variation in attenuation does not affect the corner frequencies very much for earthquakes with intermediate or large magnitudes (e.g., $M > 5\frac{1}{2}$).

At high frequencies the *Q* corrected spectra decrease approximately as f^{-2}. A straight line is drawn by eye through this portion of the spectrum on a log-log plot. The intersection of this line with the flat portion of the amplitude spectrum determines the corner frequency f_0 [*Berckhemer and Jacob*, 1968; *Hanks and Thatcher*, 1972]. Brune's theory for shear waves predicts that the radius r of an equivalent circular fault is given by $r = 2.3\beta/2\pi f_0$ where β (in centimeters per sec) is the shear velocity at the source. *Hanks and Wyss* [1972] and *Wyss and Hanks* [1972] showed that this formula gave source dimension estimates that were in close agreement with those observed in the field. Similar estimates have been made with *P* waves by replacing β with α in the expression for r and using a corner frequency f_0 derived from *P* spectra.

Seismic moment and corner frequency were estimated for each spectrum. The seismic moments and radii were averaged and are tabulated in Table 2. By using the moment and radius from $A = \pi r^2$, the area of the fault and the average displacement were estimated (Table 2). Then the stress drop for each earthquake was estimated; $\Delta\sigma = 7/16(M_0/r^3)$. Stress drops thus computed are estimated to be uncertain by a factor of 3 to 5 [*Molnar and Wyss*, 1972].

Results. An approximate test of this method can be made by comparing the seismic moment and source dimension estimated from spectral analysis with those measured in the field for the earthquake of March 20, 1966, in the Republic of Zaire. The computed seismic moment is 1.86×10^{26} dynes/cm and the radius is 16.4

km. The area is thus 844 km² and the average displacement is 67 cm. In his field work in the epicentral zone, J. Wohlenberg (personal communication, 1972) estimated that surface faulting extended for about 40 km and that the displacement reached a maximum of 2 to 3 meters near the center of the fault. Thus the average is about 1.5 meters. If the fault extends to a depth of 10 ± 5 km (i.e., a width downdip of 15 km), the area is 600 km or $r = 14$ km and $M_o = 2 \times 10^{25}$ dynes/cm. The agreement between these estimates and those determined by spectral analysis is quite good, if we consider the high error estimates.

The stress drops for the eight earthquakes are plotted in Figure 5. Several studies suggest that, for this range of seismic moment, the stress drop tends to increase as the moment increases [*Berckhemer and Jacob*, 1968; *King and Knopoff*, 1968; *Molnar and Wyss*, 1972; *Wyss*, 1970]. The line in Figure 5 was determined by *Wyss* [1970] as an estimate of this tendency. The data are consistent with such an increase but do not establish it.

Four of the earthquakes (events 7, 8, 9, and 11) occurred outside the rift system and a fifth (event 5) does not appear to be related to the faulting in the rift valley. The remaining three were associated with faulting in the rift system. If it is accepted that the stress drop $\Delta\sigma$ increases as moment M_o increases, the five events not associated with rifting have the highest stress drops. *Molnar and Wyss* [1972] found that earthquakes in the Tonga arc with the relatively highest stress drops occurred within one of the lithospheric plates and not on the main underthrusting zone between the plates. Thus we suggest that the pattern observed in Africa is real; the earthquakes that occur in the rift system and that reflect tectonic rifting of a plate boundary that is observed geologically tend to have relatively small stress drops in comparison with events occurring within one of the plates.

If we assume that the larger stress drops reflect greater strength of the material, the data are consistent with the idea that the lithosphere is continuous in southern Africa but not in the northern part of the rift system. We conclude that the northern part of the rift system marks a narrow zone of weakness separating two stable, aseismic plates, whereas the southern part of the rift is not such a plate boundary

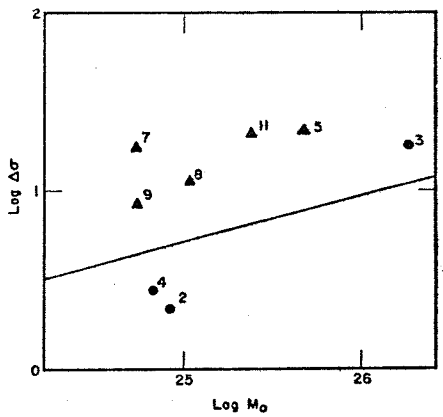

Fig. 5. Stress drop $\Delta\sigma$ versus seismic moment M_o. Numbers are from Table 1. The straight line is from *Wyss* [1970]. It is not a fit to the present data. Circles and triangles indicate earthquakes that occurred within and out of the rift system, respectively.

yet or a zone of weakness. The scatter in the earthquake locations and the higher stress drops indicate that a distinct zone of weakness does not exist there despite the seismic tectonic activity observed.

DISCUSSION

Data from many sources suggest a similarity in the structure and tectonics of oceanic ridges and those of the eastern rift and the northern part of the western rift of Africa. The contemporaneous development of these two parts of the rift system with the Gulf of Aden and the Red Sea [*Mohr*, 1967] and the directions of spreading in the last two regions suggest that the two designated parts of the rift system opened at a slow rate as the Gulf of Aden and the Red Sea formed [*McKenzie et al.*, 1970]. Most of the southern part of the western rift appears to have formed earlier, and it may not behave as a plate boundary now.

In southern Africa the lithosphere appears to be continuous, and seismicity does not define a clear, narrow zone of weakness. In the last 20 years the larger earthquakes in southern Africa did not occur within the rift system. Nevertheless the abundant earthquake activity shows that southern Africa is tectonically active. The relatively high stress drops outside the rift in southern Africa suggest higher stress there than inside the rift in the north. The consistent easterly orientation of the least compressive stress implies a similar orientation

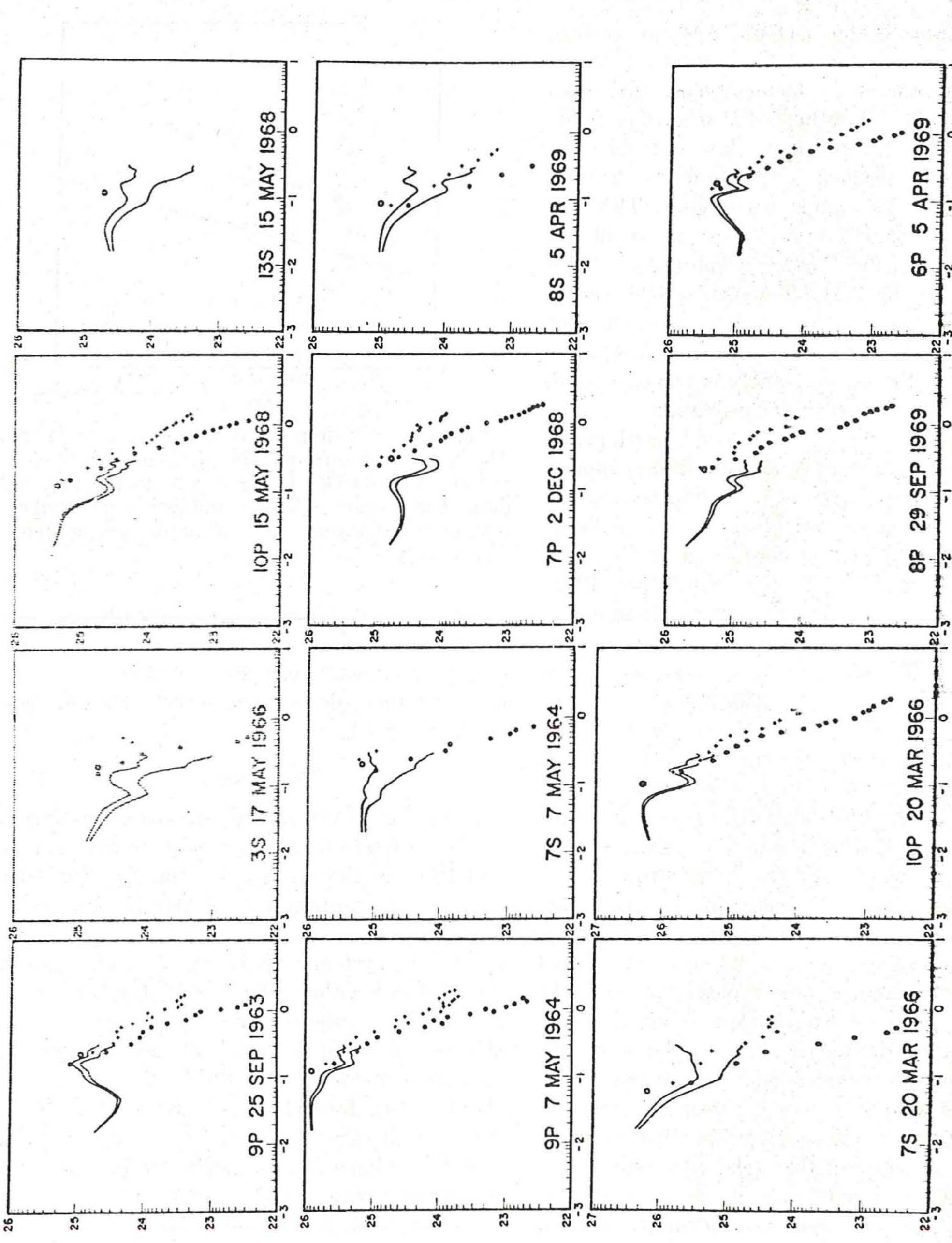

Fig. 6. Root-mean-average spectra of ground displacement for P and S waves for some African earthquakes on a log-log plot. The seismic moment is determined from the flat long-period portion of the spectrum (ordinate scale in dynes per centimeter). The abscissa is frequency in hertz. Continuous lines are long-period spectra. (The higher curves are Q corrected and the lower are Q uncorrected.) Short-period spectra are denoted by crosses and plus signs for uncorrected and Q corrected, respectively. The rms seismic moment and the corner frequency are denoted by open circles. The average moment and corner frequency on this plot are as defined by *Molnar and Wyss* [1972].

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of principal stresses throughout the rift system south of it. Thus we propose that most of the western rift is not a zone of weakness and that the zone of currently active rifting is extending or will extend farther south of Lake Tanganyika to further break up the African plate.

That the initial volcanism is older in Ethiopia than in Kenya suggests that the eastern rift may have grown southward during its formation. In other areas *McKenzie et al.* [1970] suggest that the Sheba ridge penetrated into Africa in successive stages to form the Gulf of Aden. *Swartz and Arden* [1960] claim that the opening of the Red Sea began in the south and moved northward. *Wright* [1968] suggests that the south Atlantic opened first in the south and that the break between South America and Africa moved north. Moreover, that plates break up by the growth of the zone of rifting from one end to the other is reasonable in view of the high strength of unfaulted rock and the stress concentration that might form at the tip of a growing rift.

With respect to the higher stress drops for the earthquakes in southern Africa, two additional points are worth noting [*Molnar and Wyss*, 1972]. First, because much of the destruction caused by earthquakes is due to shaking in the frequency band near 1 Hz, high stress drop earthquakes, rich in high frequencies, may tend to be more destructive than lower stress drop events. Second, the best criteria for discrimination between earthquakes and explosions is based on spectral differences, and high stress drop earthquakes radiate signals that are more similar to explosions than those radiated by lower stress events. Therefore it is important to know that, from their spectral content alone, earthquakes that are within plates anywhere on the earth may be more likely to be incorrectely identified as explosions than earthquakes on plate boundaries are.

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