PRELIMINARY RESULTS ON UPPER MANTLE STRUCTURE IN CENTRAL AFRICA FROM SPECTRAL ANALYSIS OF P WAVES

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Résumé

Résultats préliminaires sur la structure du Manteau Supérieur de l'Afrique centrale d'après l'analyse spectrale des ondes P

Les spectres d'accélération des ondes P, de séismes dont les distances épicentrales sont comprises entre 1° et 45°, ont été obtenus à partir d'enregistrements sur bandes magnétiques effectués en deux stations : l'Observatoire de Bangui, et la station temporaire de Yongossaba. La stabilité des spectres étant constatée pour des séismes de même origine, les variations de la fréquence du maximum spectral sont considérées — compte tenu de la cohérence des observations — comme une fonction de la distance épicentrale (fig. 2) qui peut être altérée par des inhomogénéités latérales du Manteau Supérieur. Le facteur de qualité Q est déterminé à partir de la pente (négative) des amplitudes spectrales vers les hautes fréquences. Les valeurs expérimentales de T/Q sont calculées en fonction de la distance, T étant le temps de propagation, (tabl. II). A la base d'une couche à faible atténuation (Q élevé) assimilée à la Lithosphère, un niveau à forte atténuation (Q faible) est mis en évidence, à la fois par les deux méthodes d'analyse (maximum spectral et T/Q), et assimilé à l'asthénosphère. L'effet de ce niveau est particulièrement net à 14-15° (fig. 2 et 8). Au-delà, entre 20° et 23-25°, les différences dans les variations du spectre global (maximum spectral) et de T/Q sont interprétées comme résultant des conditions particulières de propagation dans un milieu à forte atténuation qui correspond — en accord avec les travaux de nombreux auteurs — au système des rifts de l'Afrique orientale. Ces conditions affectent peu la valeur de T/Q, il est donc possible, indépendamment des variations latérales qui modifient le spectre global, de proposer une distribution approximative du facteur de Qualité pour le manteau supérieur correspondant à la structure du bouclier africain.

Summary

Acceleration spectra of P waves are obtained from magnetic tape recordings of two seismological stations, Bangui Observatory and Yongossaba a temporary station. Displacements of the frequency of spectral maximum are considered as a function of the epicentral distance and ray paths. From the slope *of spectra towards high frequencies, the variation of T/Q is calculated; T being propagation time and Q being mean value of Quality Factor. The existence of a low Q layer is inferred from both these types of investigations. Differences between global-spectral variations and T/Q suggest that structural particularities, as discussed by various authors, can be associated with rift areas. An approximate Q distribution is proposed, from T/Q data, for Upper Mantle structure under the African shield.*

Introduction

Bangui seismological observatory is extremely well situated for body wave studies both in the time and the frequency domain for various reasons. Its geographical situation, of a wide continental shield, ensures a very low microseismic noise. The upper mantle structure under the platform and basins may be considered as unchanging except for the Rift area. Epicentral distances being almost continuously covered from about 1°, comparison of earthquakes from widely separated regions (continental and oceanic origin) is possible. The station is equipped, among other instruments, with a system of FM recording on magnetic tape by triggering (Choudhury and Houri, 1973). This system had been operating at Bangui (4° 26.1′ N, 18° 32.8′ E, alt. 412 m) and Yongossaba, a temporary station (4° 53.5′ N, 22° 57.9′ E, alt. 600 m). Willmore seismometers, of 2.3 second natural period with a damping of 0.8 were used to carry out this study on P waves.

Earthquakes used in this paper are from various regions in and around the African continent (fig. 1), in the distance range 1 to 45 degrees. Main interest is centered on earthquakes from Western-

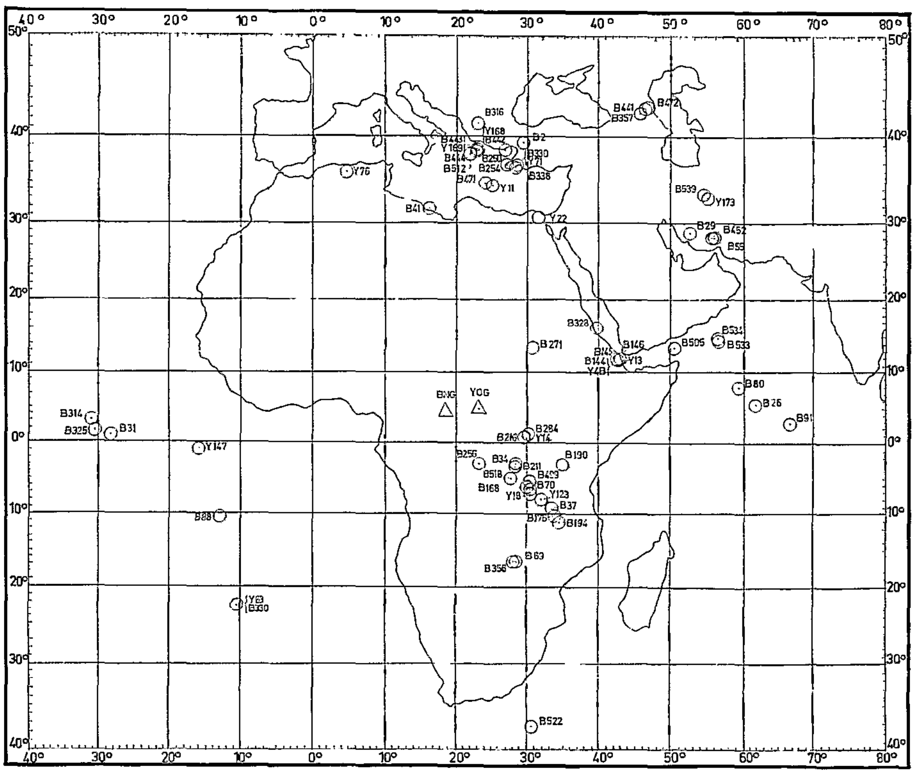


Fig. 1. — Map of epicenters. Events are indicated against the recording station B for Bangui and Y for Yongossaba temporary station.

Carte des épicentres. La station d'enregistrement est indiquée pour chaque séisme : B pour Bangui, Y pour Yongossaba.

African-Rift Valley. Different parameters of events are given in Table I; the recording station is indicated by B for Bangui and Y for Yongossaba. Some earthquakes utilised, for which no USCGS determination is available, are not reported in Table I, a number to Bangui catalogue of earthquakes is only given in the figure or in the text.

1. Spectral variations with epicentral distances

1.1. Preliminary conditions

The observed ground acceleration spectrum $\ddot{A}(f)$ can be written :

$$\ddot{A}(f)=\ddot{A}_o(f).\ddot{A}_m(f).\ddot{A}_c(f).\ddot{A}_i(f) \qquad (1)$$

where $\ddot{A}_o$, $\ddot{A}_m$, $\ddot{A}_c$ and $\ddot{A}_i$ are respectively Source function and transfert function of mantle, crust and instrument.

If we consider the differential equation of A corresponding to various earthquakes obtained at the same station with the same instrument: epicentral distance Δ and earthquake energy E are variable; but $\ddot{A}_c$ and $\ddot{A}_i$ remain constant.

We can write:

$$[d\ddot{A}=\ddot{A}_c.\ddot{A}_i[d\ddot{A}_o.\ddot{A}_m+d\ddot{A}_m.\ddot{A}_o]]$$

$$\left[d\ddot{A}=\ddot{A}_c.\ddot{A}_i\left[\ddot{A}_m\left(\frac{\delta\ddot{A}_o}{\delta\Delta}d\Delta+\frac{\delta\ddot{A}_o}{\delta E}dE\right)+\ddot{A}_o\left(\frac{\delta\ddot{A}_m}{\delta\Delta}d\Delta+\frac{\delta\ddot{A}_m}{\delta E}dE\right)\right]\right]$$

Taking the following assumptions:

— $\ddot{A}_o$ is independant of Δ for a given seismic zone (which is mainly here African Rift from 8° to 25°) where earthquakes lie at a roughly constant depth (ND)
— $\ddot{A}_m$ is independant of energy E
— $\ddot{A}_o$ is dependant on E, but neglecting this dependancy for a small magnitude range between 4.3 and 5.5 (most of earthquakes used are from 4.7 to 5.4)

we may write:

$$d\ddot{A}=\ddot{A}_o.\ddot{A}_c.\ddot{A}_i\frac{\delta\ddot{A}_m}{\delta\Delta}d\Delta \qquad (2)$$

When normalized functions $\ddot{A}_o$, $\ddot{A}_c$ and $\ddot{A}_i$ are considered to be constant, the variations of normalized spectra $\ddot{A}$ with Δ can be directly correlated to the variation of $\ddot{A}_m$.

1.2. Experimental data

Acceleration spectra are obtained by band-pass analog filtering with a Khron-Hite filter of 48 db/octave attenuation. The demodulated signal is successively replaid in the frequency range 0.1 to 8 Hz. Steps of filtering are: 0.1 Hz from 0.1 to 1 Hz, 0.2 Hz from 1 to 2 Hz and 1 Hz from 2 to 8 Hz. Measured amplitude of P waves, first arrival, for every frequency is corrected where necessary for the instrument reponse to obtain spectra. These spectra are gathered after normalisation on a map (fig. 2), where each earthquake is denoted at the corresponding epicentral distance, any spectrum being a cross section of the map along a given ordinate.

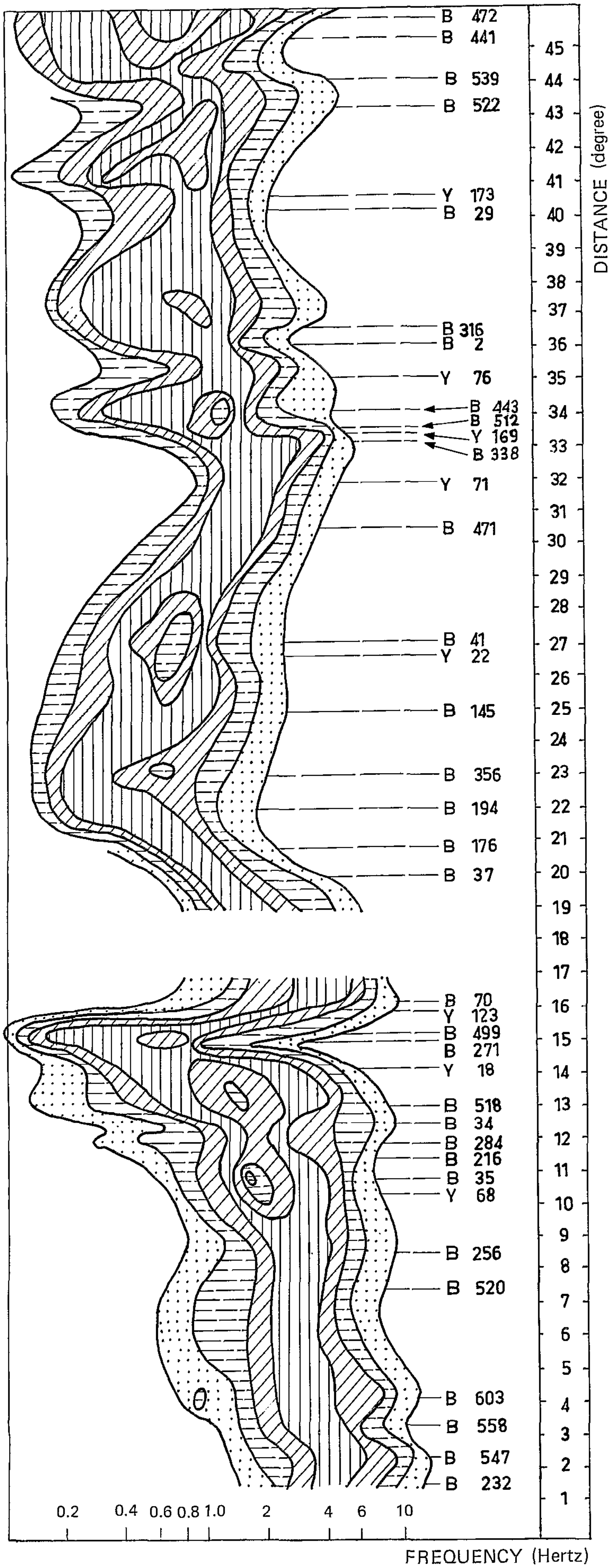
Sometimes, spectrum of a second arrival, of stronger amplitude is given (fig. 3 and 6). This second arrival does not start more than 3 seconds after the first. The time duration of the window including both the arrivals does not exceed 6 seconds.

The global map of spectra is given in fig. 2 as a function distance showing coherent variation of two consecutive spectra with respect to distance. The map shows only few of many earthquakes

TABLE I
List of earthquakes

REF. NUMBER	DATE	ORIGIN TIME	LATITUDE	LONGITUDE	DEPTH (km)	MB	Δ(DEGREE)	REGION
B2	Mar.14.72	14 05 45.79	39.277 N	29.420 E	33	5.4	36.2	Turkey
B26	Apr. 5.72	00 27 31.36	5.193 N	61.922 E	*N.D.	5.2	43.2	Carlsberg Ridge
B29	Apr.10.72	02 06 53.17	23.434 N	52.829 E	N.D.	6.1	40.4	Iran
B31	Apr.11.72	02 21 15.74	0.967 N	28.286 W	N.D.	6.0	46.9	Mid-Atlantic Ridge
B34	Apr.18.72	15 07 49.05	2.976 S	28.707 E	5.	5.4	12.6	Tanganyika lake
B37	Apr.25.72	00 59 49.76	9.133 S	33.466 E	N.D.	4.7	20.1	Tanzania
B41	Sep. 2.72	14 53 51.42	31.494 N	16.296 E	N.D.	5.0	27.1	off coast of Lybia
B69	Jan.13.73	06 05 42.38	16.808 S	28.367 E	N.D.	5.0	23.4	Zambia
B70	Jan.14.73	13 36 59.78	6.854 S	30.302 E	N.D.	4.5	16.3	Tanganyika lake
B80	Feb. 2.73	10 15 02.09	7.710 N	59.565 E	N.D.	4.4	40.9	Carlsberg Ridge
B88	Feb. 8.73	19 05 21.90	10.377 S	13.014 W	N.D.	5.3	34.7	Ascension island
B91	Feb.10.73	14 37 35.08	2.520 N	66.424 E	N.D.	5.3	47.8	Carlsberg Ridge
B144 (Y4B)	Mar.28.73	13 35 04. 5	11.673 N	42.830 E	N.D.	5.0	25.1	Djibouti
B145 (Y4C)	Mar.28.73	13 42 06.68	11.739 N	42.746 E	N.D.	5.3	25.0	Djibouti
B146 (Y4D)	Mar.28.73	14 18 52.34	11.703 N	42.927 E	N.D.	5.4	25.2	Djibouti
B162	May 11.73	13 52 31.70	33.354 N	57.376 E	50	5.1	46.3	Iran
B176	May 24.73	07 15 08. 5	10.341 S	33.764 E	N.D.	4.3	21.2	Malawi
B194	Jul.18.73	19 39 13.00	11.192 S	34.447 E	10	4.7	22.2	Malawi
B211	Nov.21.73	00 50 30.17	3.607 S	28.186 E	N.D.	5.1	12.5	Tanganyika lake
B216	Dec. 1.73	16 51 14. 0	0.647 N	29.517 E	15	4.8	11.6	Zaïre
B250	Feb. 1.74	00 01 02.37	38.552 N	27.015 E	29	4.9	34.9	Turkey
B256	Feb.10.74	16 29 26. 4	2.9 S	23.3 E	N.D.	4.7	8.7	Zaïre
B271	Mar.17.74	07 31 25. 7	13.284 N	30.761 E	N.D.	4.2	14.9	Sudan
B284 (Y14)	Apr.25.74	00 03 49.10	0.995 N	30.091 E	N.D.	5.0	11.9	Uganda
B314	Jun.20.74	02 44 19.79	3.132 N	31.303 W	N.D.	5.0	49.7	Mid-Atlantic Ridge
B316	Jun.22.74	23 30 15.02	41.262 N	23.032 E	N.D.	5.1	37.0	Greece Bulgaria
B325	Jun.27.74	18 46 25.71	1.520 N	30.841 W	N.D.	5.4	49.4	Mid-Atlantic Ridge
B328	Jun.30.74	13 26 24. 7	16.0 N	39.6 E	-	-	23.7	Ethiopia
B330	Jul. 1.74	23 11 14.54	22.642 S	10.674 W	N.D.	5.6	39.3	South-Atlantic Ridge
B338	Jul. 9.74	02 32 17. 6	36.652 N	28.449 E	69	5.0	33.5	Dodecanese
B356	Aug. 1.74	09 36 27. 0	16.7 S	28.0 E	14	5.1	23.1	Zambia
B357	Aug. 4.74	15 06 17. 1	42.3 N	45.9 E	N.D.	5.4	45.0	Caucasus
B441	Nov.13.74	02 36 25. 5	42.720 N	46.557 E	42	5.1	45.8	Caucasus
B442	Nov.14.74	13 22 33.06	38.518 N	23.123 E	19	5.0	34.3	Greece
B443	Nov.14.74	14 26 45. 8	38.5 N	23.0 E	-	5.1	34.3	Greece
B444	Nov.14.74	15 29 44. 8	38.5 N	23.1 E	-	5.0	34.3	Greece
B452	Dec. 2.74	09 05 44. 2	27.991 N	55.819 E	36	5.4	42.5	South-Iran
B471	Jan. 9.75	18 53 43.95	34.756 N	24.088 E	47	4.6	30.7	Crete
B472	Jan. 9.75	23 09 46.63	42.889 F	46.987 E	31	5.2	45.9	Caucasus
B499	Mar.26.75	03 40 48.75	5.450 S	30.207 E	N.D.	5.1	15.2	Tanganyika lake
B505	Mar.29.75	09 36 20.97	13.385 N	50.738 E	N.D.	5.4	32.9	Gulf of Aden
B512	Apr. 4.75	05 16 16.20	38.093 N	21.980 E	53	5.4	33.8	Greece
B518	Apr. 6.75	04 52 07.63	05.105 S	27.713 E	34	4.7	13.2	Zaïre
B522	Apr. 7.75	14 38 10.52	37.521 S	30.881 E	25	5.1	43.4	Off coast of South Africa
B533	Apr.19.75	13 45 50.07	14.461 N	56.515 E	N.D.	5.3	38.7	Arabian sea
B534	Apr.19.75	17 10 54.80	14.514 N	56.510 E	N.D.	5.0	38.7	Arabian sea
B539	Apr.28.75	02 01 16. 9	33.305 N	54.825 E	42	5.3	44.4	Iran
YA-4B	(cf B144)						20.7	Djibouti
YA-4C	(cf B145)						20.7	Djibouti
YA-4D	(cf B146)						20.9	Djibouti
YA-11	06. 04.73	14 13 54.16	34.431 N	25.250 E	16	5.1	29.6	Crete
YA-13	07. 04.73	17 36 42.78	11.690 N	43.021 E	N.D.	4.7	20.9	Djibouti
YA-18	15. 04.73	13 13 33.38	7.183 S	30.274 E	36	4.7	14.1	Tanganyika lake
Y -14	(cf B284)						8.0	Uganda
Y -22	29. 04.74	20 04 39.68	30.529 N	31.721 E	N.D.	4.9	26.9	Egypt
Y -63	01. 07.74	23 11 14.54	22.642 S	10.674 W	N.D.	5.6	42.9	South-Atlantic Ridge
Y -71	09. 07.74	02 32 17.60	36.652 N	28.449 E	69	5.0	32.5	Dodecanese Island
Y -76	13. 07.74	15 57 25. 2	36.0 N	4.8 E	37	4.8	35.3	Algeria
Y 123	17. 09.74	14 30 54.94	8.051 S	32.081 E	14		15.8	Tanzania
Y 147	23. 10.74	11 46 55.96	1.009 S	15.968 W	N.D.	4.9	39.3	Ascension Island
Y 168	14. 11.74	13 22 33.05	38.510 N	23.123 E	19	5.0	33.6	Greece
Y 169	14. 11.74	14 26 45.83	38.530 N	22.991 E	3	5.1	33.6	Greece
Y 173	17. 11.74	15 05 47.67	32.805 N	55.073 E	43	5.2	40.9	Iran

*N.D. (normal Depth)



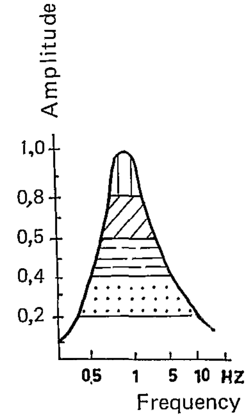


FIG. 2. — Global spectra dictribution of P waves as a function of epicentral distance. The legend is indicated by an example of a cross section.

Distribution spectrale des ondes P en fonction de la distance épicentrale. La légende est donnée par une coupe.

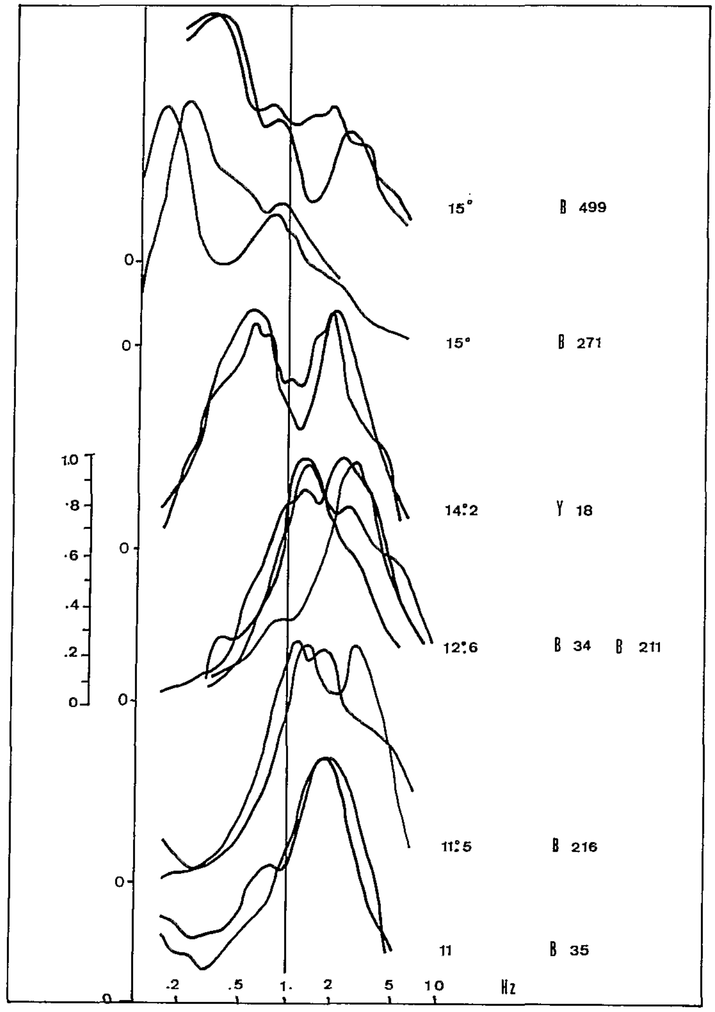


FIG. 3. — Spectra of P waves from 11° to 15°.
Spectres des ondes P de 11° à 15°.

analysed. Among them the following three groups of epicentral distance: 8° to 10° at Yongossaba, 11.7° to 12.5° and 14° to 15° at Bangui show no appreciable variation of spectrum morphology confirming the spectral stability of frequency maximum. It is generaly observed that for P waves in Bangui, for two close distances, two similar spectra are obtained even for epicentral zones sometime quite different. This is the case of spectra of two events at the epicentral distance of 15° (B499 from Tanganyika lake and 271 from Sudan). Other examples, not shown, also exist (Y4a from Djibouti and B37 from Malawi at 20°, B328 from Ethiopia and B356 from Zambia at 23°). This is not valid beyond the African domain (more than 30°) where P waves from oceanic or continental origin give very different spectra. Thus, the stability of spectra, at distances under about 30°, is uneffected by any variation arising from the choice of earthquakes or epicentral regions. This leads to the following conclusions: first, the upper mantle and the crust around the receiver can be considered as widely homogeneous; secondly, spectra of P waves seem, at the first view, independant of mechanism and magnitude of earthquakes within limits used here as well as source neighbourhood but directly correlated with epicentral distance as indicated in fig. 2.

In the first approach, if we consider the frequency of the spectral maximum which corresponds to the vertically hatched zone of the map, where normalized amplitudes are restrained between 0.8 and 1.0, we see that from 1° to 9° the maxima show a quite good stability around 3 Hz. This zone of maxima splits into two branches from 9°; one branch on the right-hand side still remains, close to 3 Hz, whereas the other drops to less than 0.4 Hz at 15°. Around 16°, both earthquakes, Y123 and B70, have a single peak which reaches again high frequencies. At distances between 17° and 19° there is a lack of recorded P waves on magnetic tape in spite of occurrence of earthquakes at these distances in Tanzania, this is probably because of too weak P waves to trigger the recording system. The earthquakes from 20° to 23°, perhaps to 25°, give decreasing frequencies of the spectral maximum from 1 Hz to 0.3 Hz. The three largest epicentral distances for purely African paths correspond to Y22 in Nile delta, B41 near the coast of Lybia and Y76 in Algeria at distances of 26°, 27° and 35°; all other events in the range 30° to 45° reported on figure 2 are located in continental regions (Greece, Turkey, Iran and Caucasus) except for B522 from south of Africa. The frequency of spectral maximum increases regularly between about 23° and 33°. From this distance up to 45° (the earthquakes are no more in the African domain), the frequency of the maximum is unstable. This instability seems to arise from the diversity of epicentral regions which entails important variations in propagation paths and source conditions. However, if we take into account only the general features, the frequency of maxima seems to remain, on the average, constant.

1.3. The 15° anomaly

Returning back to purely African paths and earthquakes, we now consider in more details, the disturbed zone of the spectral map ($11° < \Delta < 16°$) where a splitting of spectra is conspicuous. Some normalized spectra from 11° to 15° are represented on figure (3). As a control, spectrum of a second arrival, close to first onset, is also given for each earthquake. The two waves do not show any significant difference in their spectra. This leads to the conclusion that the nature of the spectra is the same in the first P wave and this later arrival. What is striking is the division of maxima into two branches from 11.5° to 15°. This splitting evolves with distance: the high-frequency branch being attenuated with increasing distances, the low-frequency tendency, on the contrary, becoming preponderant.

The high-frequency branch appears to be a prolongation, up to 15°, of the type of spectrum observed at short distances. It appears clear that this branch is representative of P waves propagating in a low attenuation medium (=high quality factor): the Lithosphere. The appearence of a second low frequency maximum up to about 15° may suggest a superposition of two waves having different spectral contents and phase velocities. In the time domain the records show in fact, a second wave group of large amplitude arriving about 3 seconds after the onset. These two groups have close phase velocities and spectra. At the present stage of analysis, it is difficult to say if these waves represent simply the source process or if they characterise propagation media. It is nevertheless important to underline that around 15° the spectral characteristics change drastically, the maximum appearing at low frequencies and high frequencies sharply attenuated. This shift of maximum towards low frequencies suggest penetration of waves in a low Q medium which is generally considered as asthenosphere.

Two remarks may be significant: Firstly, the distance variation from 9° to about 13° is obtained from earthquakes located on the western side of the Rift and propagation path lies away from the Rift domain (fig. 5). Secondly, a quite good similarity of spectra is observed at 15° corresponding to earthquakes from very different regions such as Sudan (B271) and Tanganyika lake region (B499). These two points strengthens the hypothesis that the spectral evolution between 9° and 15° is dependant on epicentral distance and not on local anomaly.

This last assessment is also supported by considering the first arrival spectra of two earthquakes which have been recorded at the couple of centrafrican stations Bangui and Yongossaba. The first earthquake (fig. 4-a) corresponds to references B284 and Y14 (see Table I), the second (fig. 4-b)

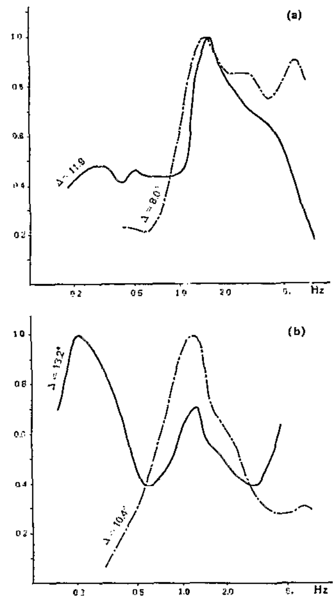


FIG. 4. — Comparison of spectra for two earthquakes (a) and (b) recorded at two stations.

Comparaison des spectres de 2 séismes (a) et (b) enregistrés en 2 stations.

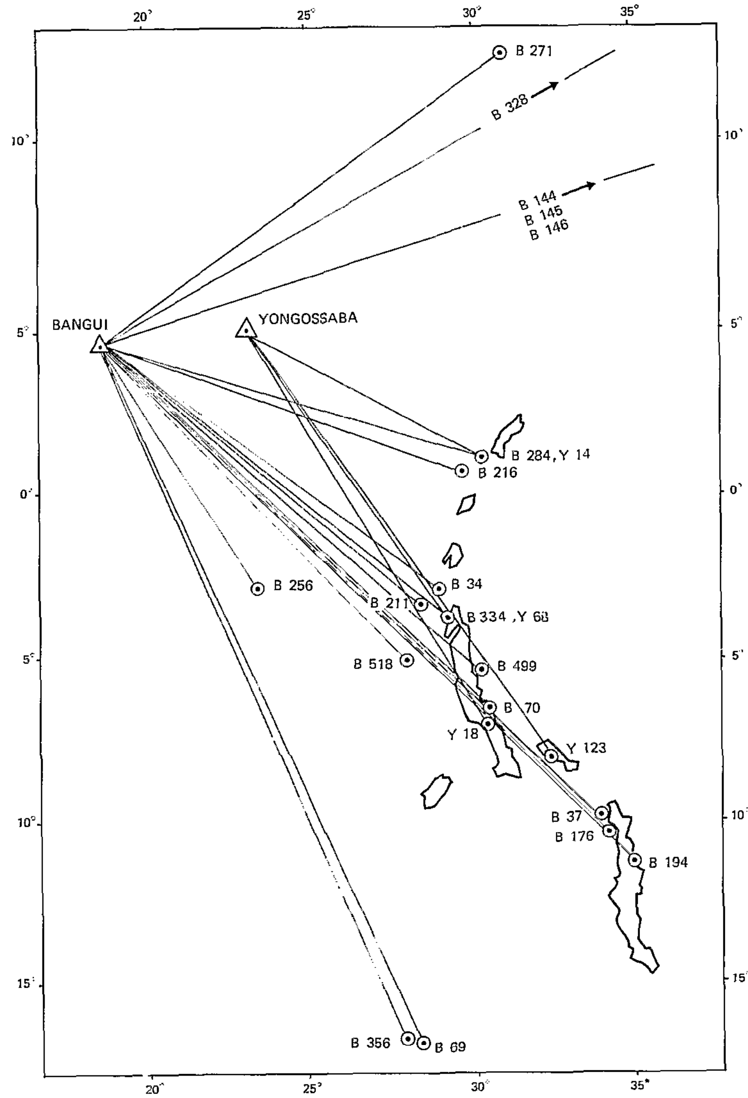


FIG. 5. — Map of propagation paths from the Rift to Bangui or Yongossaba.

Carte des trajets séismiques du Rift à Bangui ou Yougossaba.

corresponds to B334 and Y68 reported from I.S.C. determination only (Jul. 5, 1974, 05h13m10.9s - 3.64 S, 29.0 E - Mb=4.6 - h=0). Epicentral distances for Bangui, Δ_B, and for Yongossaba, Δ_Y, are respectively for (a) and (b):

$$\text{(a)}\ \Delta_B=11.9^o\quad \Delta_Y=8.0^o,\ \text{(b)}\ \Delta_B=13.2^o\quad \Delta_Y=10.4^o$$

For a given earthquake—$\ddot{A}_0$ constant—the difference between spectra can be caracterized by the importance of low frequency at larger distance. The ressemblance can be found in the stability of the high frequency maximum. This example is in good agreement with other results. However, the phenomenon of the division of spectra into two maxima for the first arrival and the close second one, which is of the same spectral type, suggest an effect of a limit, or transition zone in terms of attenuation. This can be regarded as an experimental character of the approach by seismic P waves of the Lithosphere-Asthenosphere Boundary independently of phase multiplicity. The interpretation of the process of splitting cannot be made here, but only concurrently with Velocity and Q models.

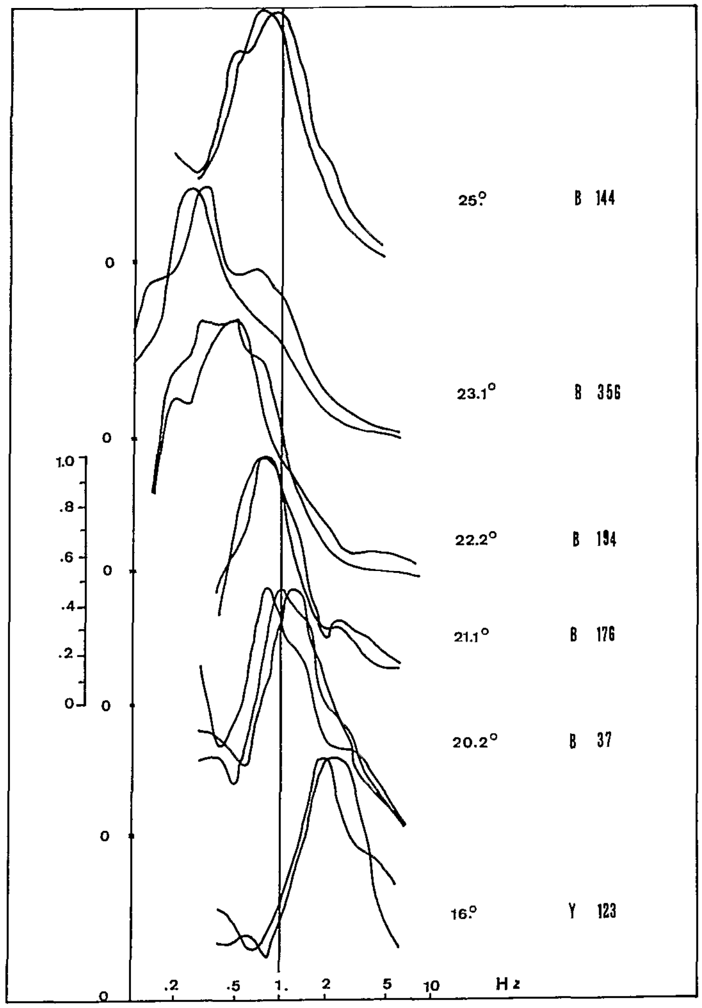


FIG. 6. — Spectra of P waves from 16° to 25°.

Spectres des ondes P de 16° à 25°.

1.4. The 23° anomaly

In fig. 6 where spectral evolution between 16° and 25° is shown; we remark that the maximum is unsplitted and moves from about 2 Hz at 16° to about 0.3 Hz at 23°. The existence of high frequency at 16° and its weekness around 15° (fig. 3) appears contradictory. Y123 being the only good observation available at this distance, we do not have any straight forward explanation. This is further complicated by the absence of records, as mentioned earlier, between 17° and 19°.

Leaving aside this particular point, we observe that the spectral map and also fig. 6 show a regular drift of maximum frequency from 1 Hz at 20° to 0.3 Hz at 23°. The simple shape—without splitting—of these spectra does not allow us to propose an interpretation similar to that for distances under 15°. At distances larger than 20°, the rays penetrate deep enough to traverse any low Q asthenosphere; the difference in the asthenosphere ray paths for $20 < \Delta < 25°$ is insufficient to explain the drift observed. This leads us to consider carefully a possible cause of attenuation on both extremities of ray paths.

For the receiver side: crust and lithosphere would be hardly concerned because earthquakes B37, B176 and B194 which represent the frequency drift, are lying nearly in the same azimuth from Bangui (fig. 5). Besides, the angle of incidence at the station decreases very little for this distance interval. On the other side of the ray paths, one can see that waves are travelling under the Rift —indicated by the contours of lakes—on more than one third of the distance. Considering the Rift as a high attenuation zone (GUMPER, POMEROY, 1970) one may admit that the drift of spectral maxima towards low frequency results from a progressively longer ray path in the Rift zone. The case of B69 and B356 is discussed below.

1.5. LATERAL VARIATIONS

The above suggestion finds many supports in works published by several authors about lateral variations of Upper Mantle in Africa. A change of velocity and attenuation between the African Rift and African shield is shown to exist from studies of wave propagations over the continent (GUMPER, POMEROY, 1970, KNOPOFF, SCHLUE, 1972). More recently impressive results were obtained from the "Durham Seismic Array in Kaptagat" (LONG, BACKHOUSE, MAGUIRE, SUNDARLINGHAM, 1972). An abnormal crust (MAGUIRE, LONG, 1976) and a complete lack of lithosphere were found under the Gregory Rift (LONG, BACKHOUSE, 1976): The authors have proposed a model—as a map of the upper surface of the anomalous low-velocity zone—deduced from slowness anomalies and relative teleseismic delay times. The identification of the shape of this boundary, away from the Rift, emphasizes the continuity of the Asthenosphere which is intrusive under the Rift and layered at the base of continental shield. The anomalous character of Upper Mantle under the Rift areas is also well correlated with a large wavelength negative Bouguer anomaly (GIRDLER, FAIRHEAD, SEARLE, SOWERBUTTS, 1969). This anomaly is extended to some adjacent regions and interpreted as a thinning of lithosphere (GIRDLER, 1975) like is observed for the Rift. The connection of this anomaly with teleseismic positive or null delay time (instead of negative for normal shield) is accounted for by a thinning of lithosphere (FAIRHEAD, REEVES, 1977).

All these results enhance the hypothesis of an extension of the anomalous Upper Mantle in Zambia and Ethiopia where earthquakes B69, B356 and B328 at a distance of 23° occur. This would explain that P waves involved in these earthquakes are propagated under approximately similar conditions to those travelling in the Rift domain.

1.6. OCEANIC AND CONTINENTAL EARTHQUAKES

As an attempt to examine the hypothesis of an effect of lateral variations on P waves, it appears to be interesting to extend our purpose to earthquakes of larger distances. Therefore, it seems significant to consider spectra of earthquakes from continental origin (Greece, Algeria, Iran and Caucasus) compared with ones of oceanic origin (Gulf of Aden, Arabian sea and Carlsberg Ridge) at equivalent epicentral distances (fig. 7). For the 33° to 48° distance range, we take advantage of the fact that P waves are not much affected by the mantle structure. So, the difference between the two sets of spectra—low and high frequency maximum for oceanic and continental origin respectively—is assigned to the Upper Mantle properties on the focal side of the ray paths. This properties include here: velocity, attenuation and focal mechanism. The low frequency of P waves of oceanic origin which have propagated a part of their paths under oceanic ridges (similar spectra were obtained from central Mid-Atlantic Ridge) can be considered as bearing some ressemblance between Mid-Oceanic Ridges and the African Rift system. Among other geological and geophysical evidences one may site the weak P velocity (7.1-7.5 km/s) in the Upper Mantle (RUEGG, 1975, LONG, BACKHOUSE, 1976) and the strong attenuation of S_n (MOLNAR, OLIVER, 1969) for both type of structures.

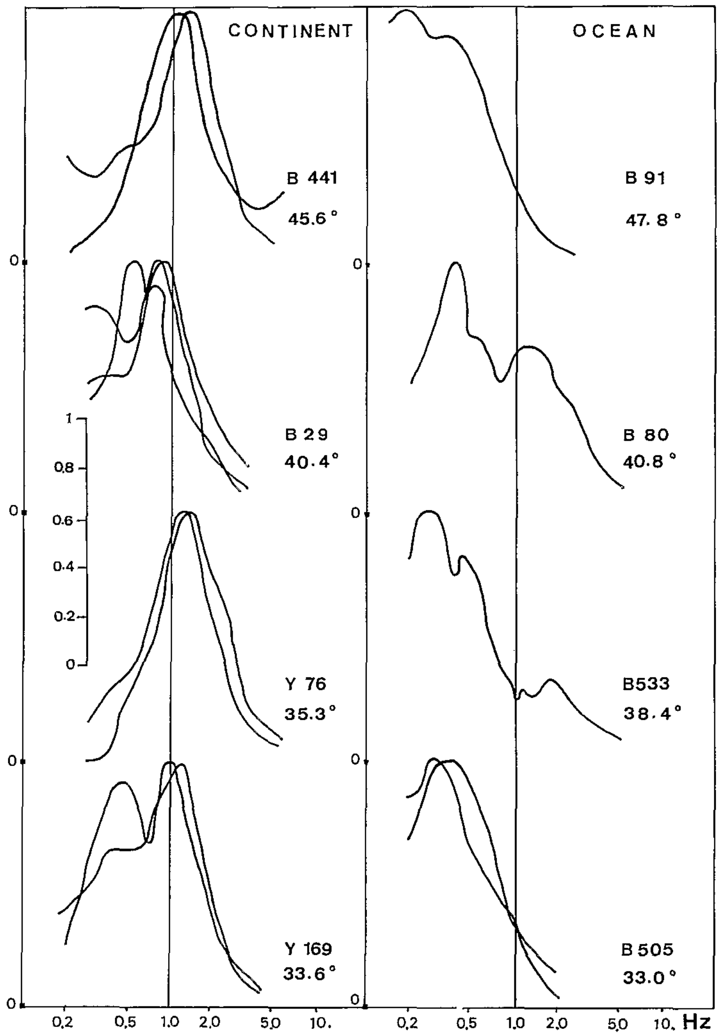


FIG. 7. — Comparison of spectra from continental and oceanic origin.
Comparaison de spectres de séismes d'origine continentale et océanique.

2. Quality factor measurements

2.1. Method

We consider the relation :

$$\ddot{A}_m(f) = G.\ \exp\left(-\Pi\ f \int_{ray} \frac{ds}{Q(z).V(z)}\right) \qquad (3)$$

Where G is frequency independant geometrical spreading factor, ds an element of ray path, Q(z) and V(z) are respectively Quality Factor and Velocity at the depth (z). Taking logarithm of (1) and differenciating with respect to frequency :

$$\frac{\ddot{A}'(f)}{\ddot{A}\,(f)} = \frac{\ddot{A}'_o(f)}{\ddot{A}_o\,(f)} + \frac{\ddot{A}'_m(f)}{\ddot{A}_m\,(f)} \qquad (4)$$

and replacing the expression of (3) in (4) :

$$\frac{\ddot{A}'(f)}{\ddot{A}(f)} = \frac{\ddot{A}'_o(f)}{\ddot{A}_o(f)} - \Pi \log e . \frac{T(\Delta)}{Q(\Delta)} \qquad (5)$$

where log e is decimal logarithm.

TABLE II

T/Q values

REF. NUMBER	Δ°	T/Q, (PHASE, SLOPE AND QUALITY OF THE SLOPE)						FREQUENCY INTERVAL
B547	2.5	0.067	(1,1,+)					6-10
B558	3.5	0.100	(1,1,+)					4-9
B603	4.3	0.058	(1,1,0)					3-5
Y26	8.5	0.064	(1,1,±)					4-8
B256	8.7	0.068	(1,1,+)	0.075	(2,1,+)			4-8
B216	11.6	0.163	(1,1,-)	0.118	(1,2,+)	0.100	(2,1,0)	1.7-3
B284	11.9	0.042	(1,1,0)					1.8-6
B34	12.6	0.084	(1,1,+)	0.091	(1,2,+)			3-8
B211	12.5	0.150	(1,1,+)	0.136	(1,2,+)			1.4-5
B518	13.2	0.120	(1,1,-)	0.120	(1,2,-)			5-7
Y18	14.1	0.192	(1,2,-)	0.212	(2,1,-)	0.147	(2,2,+)	0.6-1.3 and 3-7
B271	14.9	0.316	(1,1,+)	0.312	(2,1,+)	0.218	(2,2,0)	0.2-1.5 and 4-7
B499	15.2	0.349	(1,1,-)	0.349	(2,1,+)	0.100	(1,2,0) and (2,2,0)	0.3-1.5
Y123	15.8	0.056	(1,1,0)	0.212	(2,1,+)			0.4-0.7 and 3-6
B70	16.3	0.108	(1,1,0)	0.608	(2,1,0)	0.118	(2,2,0)	0.2-0.6
B37	20.1	0.562	(1,1,+)	0.264	(2,1,0)			1-2
Y4D	20.9	0.990	(1,1,-)	0.643	(2,1,+)			
B176	21.2	0.438	(1,1,+)	0.432	(2,1,+)	0.424	(3,1,+)	0.8-2
B194	22.2	0.454	(1,1,+)	0.468	(2,1,+)			0.5-2.5
B69	23.4	0.340	(1,1,+)					1-2.5
B328	23.7	0.332	(1,1,+)	0.132	(1,2,0)			0.6-1
B144	25.1	0.346	(1,1,+)	0.270	(1,2,0)	0.280	(2,1,+)	1-3
B146	25.2	0.346	(1,1,+)					1-3
Y22	26.9	0.666	(1,1,-)	0.261	(2,1,0)	0.183	(1,2,0) and (2,2,0)	0.4-0.7 and 0.8-4
B41	27.1	0.268	(1,1,+)	0.314	(2,1,+)			0.3-1
B471	30.7	0.227	(1,1,+)					1.6-3
B505	32.9	0.749	(1,1,+)	0.291	(1,2,0)	0.505	(2,1,+)	0.4-2
B512	33.8	0.458	(1,1,+)	0.293	(2,1,0)			0.8-2 1.2-2
Y168	33.6	0.346	(1,1,+)	0.207	(1,2,-)			0.9-2
Y169	33.6	0.376	(1,1,+)	0.357	(2,1,+)			1.2-2.6
B442	34.3	0.255	(1,1,+)					1-2
B443	34.3	0.333	(1,1,+)	0.333	(2,1,-)			1.2-2
B444	34.3	0.148	(1,2,0)					1.2-2
B250	34.9	0.364	(1,1,0)	0.442	(2,1,+)			1-2
B88	34.7	0.616	(1,1,0)	0.952	(2,1,+)			0.4-0.6 and 1.6-2.4
B2	36.2	0.272	(1,1,+)					0.3-0.7
Y76	35.3	0.244	(1,1,+)					1.4-3
B316	37.0	0.339	(1,1,+)	0.349	(1,2,-)	0.238	(2,1,0)	0.5-0.8 and 1-2
B533	38.7	0.705	(1,1,+)					0.5-1
B29	40.4	0.464	(1,1,+)					1-2
B80	40.9	0.516	(1,1,+)	0.178	(1,2,0)			0.4-0.8 and 1.5-4
Y173	40.9	0.627	(1,1,+)	0.216	(1,2,+)	0.366	(2,1,+)	0.5-3
B26	43.2	0.344	(1,1,+)	0.360	(2,1,+)			0.6-1.6
B522	43.4	0.318	(1,1,-)					1.6-2
B539	44.4	0.563	(1,1,-)	0.324	(1,2,0)			0.3-0.5 and 1.6-2
B441	45.8	0.333	(1,1,+)	0.239	(2,1,+)			1.4-3
B472	45.9	0.732	(1,1,+)	0.115	(1,2,+)	0.305	(2,1,+)	0.8-0.6 and 1.6-2
B91	47.8	0.572	(1,1,+)					0.2-1

$\ddot{A}'(f)/\ddot{A}(f)$ being the slope of logarithmic amplitude versus frequency curve, $T(\Delta)$ and $Q(\Delta)$ are travel time and mean value of Q at the distance Δ.

Considering only frequencies greater than about 0.4 Hz, we assume that $\ddot{A}'_0(f)/\ddot{A}_0(f)=0$, so, $T(\Delta)/Q(\Delta)$ can be deduced from slope of spectra towards high frequencies (Choudhury, 1972). Consequently, this method can be used as another approach of investigation of spectral behaviour independently of its maximum. This is done for all earthquakes and slopes calculated.

2.2. Data

The Table II gives experimental values of T/Q; we denote in parenthesis after each value:

1° a number which indicates the order of arrival of the phase (first or second),

2° a number indicating if it is the first or second observed slope for a given phase (see remark),

3° a sign which gives the quality of amplitude distribution to determine the slope: (+) slope obtained by numerous points in good alignement, (0) slope obtained with few points or numerous ones dispersed, (—) slope obtained with very dispersed points.

For each line, limits of considered frequencies are given.

Remark: The second observed slope is generally the same as that observed at short distances. It can be attached to local high frequency noise. Amplitudes of this second slope is a few per cent of the signal amplitude.

The general distribution of experimental values of T/Q as a function of epicentral distance appears in Table II above. Results are summarized as follows:

$$\Delta < 12^\circ : 0.05 < T/Q < 0.10$$
$$12^\circ < \Delta < 20^\circ : 0.10 < T/Q < 0.60$$
$$20^\circ < \Delta < 25^\circ : 0.60 > T/Q > 0.30$$
$$25^\circ < \Delta < 40^\circ : 0.30 < T/Q < 0.45$$

for a constant quality factor distribution—Q=1000—applied on P velocity model of Herrin, the limits of T/Q variations for the same four intervals are: 0.02-0.15, 0.15-0.28, 0.28-0.35, 0.35-0.45.

A comparison between these values and those observed demonstrate that a large discrepency exists in the distance range 12° to 25°. The figure (8), where experimental data are reported, shows more abundantly that the phenomenon of a rapide increase of T/Q is particularly evident at 14-15°, in agreement with results found from variations of global spectra. It is sufficiently clear from the above result that a low Q layer exists under the lithosphere of central Africa located more precisely under the North-Eastern part of Zaïre (fig. 5). It is also clear, in spite of a large scattering, that on the distance interval 20-25°, T/Q values are decreasing. This sense of variation would imply Q values increasing with depth. The resulting distribution of Q in the Upper Mantle would be roughly in agreement with three layers: M_1 (Moho-140 km), M_2 (140-235 km) and M_3 (235-540 km) as defined to describe the upper part of the Earth structure (Mohammadioun, 1967), where M_2 *only* is an absorbent medium.

From 25-30° to 45°, T/Q values are very much dispersed. Values from oceanic origin (shown by open symbols) being eliminated, the dispersion still remains very strong. Some earthquakes from Greece, Dodecanese, and Turkey are shown to be of a continental type without knowing their relationship to the east Mediterranean structure. Scarcity of data between 25° and 35° roughly, does not allow to draw a definite conclusion about the tendency increasing, constant or decreasing of T/Q variation up to 45°. More data are needed to precise this point.

2.3. Discussion

When variations of T/Q as a function of distance (fig. 8) are compared with global spectral anomalies (fig. 2), the existence of a low Q layer becomes evident. This low Q layer can be associated

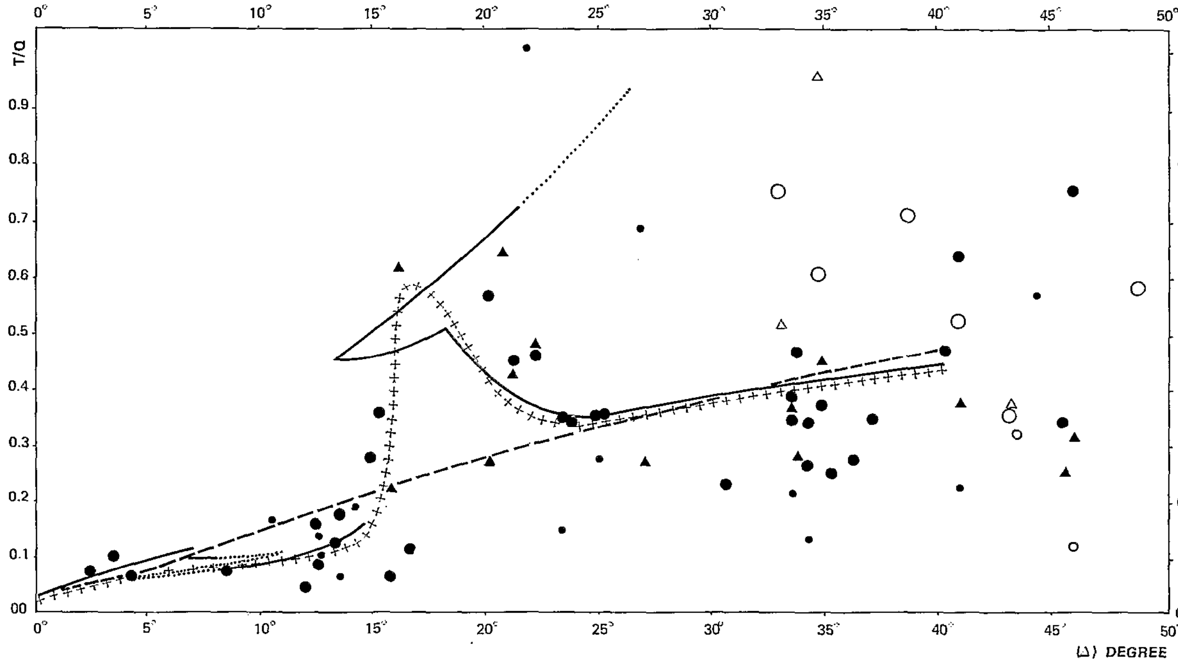


FIG. 8. — T/Q variations with epicentral distances. Experimental data are: great black circle for first arrival of P and first slope, little black circle for first arrival and second slope, black triangle for second arrival of P and first slope. Open symbols indicate oceanic origin. For curves, see legend in the text.

Variations de T/Q en fonction de la distance épicentrale.

with the low velocity layer of various models of Upper Mantle (NIAZI, ANDERSON, 1965, JOHNSON, 1967, JULIAN, ANDERSON, 1968, GREEN, HALES, 1968, HELMBERGER, WIGGINS, 1971). The depth of the top of this low Q layer (i.e. the thicknee of lithosphere) in the African shield should be estimated from local travel time and slowness determinations. At the present state of this work, the depth of the low Q layer as demonstrated by our results on T/Q and global spectra behaviour, has been estimated, from Herrin's velocity model (HERRIN, 1968), to be more than or equal to 160 km. Such a depth was proposed for the top of low velocity layer for Western United States (WIGGINS, HELMBERGER, 1973).

The relationship between Lithosphere and Asthenosphere being clear enough for short distances ($\Delta < 15°$) difficulty of interpretation remains, however, for larger distances. Effectively, the variation of T/Q and those of spectral maximum seem to be in contradiction from 20° to 25° epicentral distance: the first one involves an increase of mean value of Quality Factor while the second suggests a stronger attenuation. If we consider lateral variations in the Upper Mantle, we can roughly explain such a discrepency by the hypothesis that the variation of spectral slope resulting from the mean value of Q along the ray is not affected by the loss of high frequencies near the source—so far as the spectral maximum frequency is greater than 0.3, 0.4 Hz—but reflects the variation of mean attenuation properties along the ray through its progressive penetration in high Q media. Under this hypothesis, the drift of spectral maximum towards low frequencies (fig. 2 and 6) would reveal lateral inhomogeneities as discussed earlier; more precisely, the loss of high frequencies would result from propagation in a low Q rift structure in the source region. It must, however, be remembered, that the above interpretation is based on the assumption that $\ddot{A}_0$ is identical for all shocks.

From these various considerations, it is evident that variations of global spectra cannot be interpreted in details; consequently, we shall limit ourselves to T/Q result which is more representative of Upper mantle structure beneath the African shield. Quality Factor and Velocity models were computed in order to check experimental results of T/Q. Curves obtained for three models are represented on figure (8) as follows:

a) the dotted line represents a constant Q model (Q=1000) applied on Herrin's velocity model,

b) the crossed line, a variable Q model having a low Q layer starting at 185 km depth with Herrin's P model,

c) the full line, the same Q model as (b) with a modified Herrin's P model from Moho to 350 km. The main difference is the existence of a low velocity layer corresponding to low Q layer.

Conclusion

As can be seen in fig. 8, the numerical models do not satisfy entirely the observed data, we only can give approximate values of Q: 500 for the crust, 2000 for the Lithosphere, 200 to 800 for the Asthenosphere and 2500 for the Mantle. The main feature of our results is that constant Q or increasing Q are not compatible with observations. The structure of the Upper Mantle can be perfectible by use of local slowness data, propagation time and multiple arrivals; the hypothesis of possible interferences on P arrivals (McMECHAN, 1976) may be regarded.

The complexity in lateral extension of a high-attenuation medium (Rift structure) as revealed by various geological and geophysical works as well as results presented in this paper, clearly demonstrates the necessity of a large number of data and underlines the limits of acceptable assumptions.

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