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OBSERVED VARIATION IN THE SPECTRAL RATIO DISCRIMINANT FROM SHORT-PERIOD P WAVES

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Seismic Data Analysis Center

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Short-period P-wave first arrivals recorded at the Large Aperture Seismic Array in Montana from a large number of teleseismic events were filtered in the bands of 0.4-3.0, 0.4-0.8, 1.4-1.8, and 1.8-4.0 Hz. Estimates of signal amplitude in each band were used to form spectral ratios similar to those which have been proposed as discriminants between earthquakes and underground nuclear explosions. Values for the ratios scattered over nearly two orders of magnitude and apparently did not depend on earthquake size or epicentral distance. Average logarithms of spectral ratios for 166 separate geographic regions ranged

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over one magnitude unit and showed some correlation with known patterns of lateral Q inhomogeneity in the earth, suggesting that attenuation causes much of the spectral ratio scatter. However, attempts to correct LASA m_b on the basis of observed spectral ratios failed to reduce the scatter in an M_s-m_b plot. Variation in stress drops and other physical processes connected with the mechanism of earthquakes must cause as much variability in observed spectral discriminants as variations in attenuation. Spectral ratios for several explosions in the USSR did not separate well from those of the global population of earthquakes; and, more importantly, they did not separate well from earthquakes in the same regions as the explosions.

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OBSERVED VARIATION IN THE SPECTRAL RATIO
DISCRIMINANT FROM SHORT-PERIOD P WAVES

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ABSTRACT

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I. INTRODUCTION

In this report we describe a study of the variability of the shape of the amplitude spectrum in the main signal bandwidth of short-period P-wave teleseisms as seen on LASA recordings. The impetus for such a study comes from the numerous published suggestions, based on both theory and observation, that ratios of certain parts of this P-wave signal band may effectively serve as a means of distinguishing source type, i.e., underground explosion or natural earthquake. Before explaining the purpose and procedure of our study further, it seems appropriate to review past work on this spectral discriminant.

Early evidence of difference in P-wave spectra between earthquakes and underground explosions was presented by Willis (1963) and Frantti (1963), but these contributions represented only a few seismic events and suggested no quantitative discriminant parameter. In reports by Briscoe (1966) and Briscoe and Walsh (1967), a ratio of energy in two spectral bands was suggested as such a parameter and was evaluated for short-period P-waves recorded at LASA from several Eurasian events. They used the spectral bands of 0.3-0.7 Hz and 1.5-1.9 Hz. By analyzing a large population of Eurasian events which included many explosions, Kelly (1968) and Lacoss (1969) brought the short-period spectral ratio into prominence as a discriminant which correctly classified roughly 90% of their event sample. Based on calculations of S/N ratios for each event, Lacoss (using the spectral bands 0.45-0.95 Hz

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Willis, D.E. (1963), Comparison of seismic waves generated by different types of source, <u>Bull</u>. <u>Seism</u>. <u>Soc</u>. <u>Am</u>., v. 53, p. 965-978.

Frantti, G.E. (1963), Energy spetra for underground explosions and earthquakes, <u>Bull</u>. <u>Seism</u>. <u>Soc</u>. <u>Am</u>., v. 53, p. 997-1006.

Briscoe, H.W. (1966), Ratios of spectral densities, in Semiannual Technical Summary--Seismic Discrimination, 30 June 1967, Lincoln Laboratory, Lexington, Massachusetts.

Briscoe, H.W. and R. walsh (1967), Ratios of spectral densities, in Semiannual Technical Summary--Seismic Discrimination, 30 June 1967, Lincoln Laboratory, Lexington, Massachusetts.

Kelly, E.J. (1968), A study of two short-period discriminants, Technical Note 1969-24, Lincoln Laboratory, Lexington, Massachusetts.

and 1.45-1.95 Hz) determined that the LASA threshold for discrimination by spectral ratio was at approximatly m_b=4.8 for Eurasian events. This magnitude is often quoted as the conservative limit of positive identification by the M_s-m_b discriminant.

The P-wave spectral discriminant has continued to be studied in various forms. Perhaps the simplest is a measurement of the dominant or initial period of the P-wave signal in the time domain (Noponen, 1975; Lacoss, 1969). A form employed by Anglin (1971 and Weichert (1971) is the third moment of frequency (TMF) of the amplitude spectrum between 0.32 and 5.0 Hz, which rather successfully separated a large sample of Eurasian explosions and earthquakes. A recent innovation is the plotting of autoregressive parameters, which are fit to the time series of seismic P waves, by Tjostheim (1975). Again, this technique resulted in fair separation of a large sample of Eurasian events. Shumway and Blandford (1974) applied classical linear discrimination to the raw event spectra, using a learning and test population of Eurasian events. They found good results and suggested that those explosions which failed the discriminant were cratering events. These workers also suggested that the reason for the success of the discriminant, when applied to Eurasian explosions, was the cancellation of low-frequency P by pP.

Lacoss, R.T. (1969), A large-population LASA discrimination experiment, Technical Note 1969-24, Lincoln Laboratory, Lexington, Massachusetts.

Noponen, I. (1975), Compressional wave power spectrum from seismic sources--Final scientific report, Institute of Seismology, U. of Helsinki, Helsinki, Finland.

Lacoss, R.T. (1969), A large-population LASA discrimination experiment, Technical Note 1969-24, Lincoln Laboratory, Lexington, Massachusetts.

Anglin, F.M. (1971), Discrimination of earthquakes and explosions using short-period seismic array data, Nature, v. 233, p. 51-52.

Weichart, D.H. (1971), Short-period spectral discriminant for earthquake-explosion differentiation, Z. Geophys., v. 37, p. 147-152.

Tjostheim, D. (1975), Autoregressive representation of Seismic P-wave signals with an application to the problem of short-period discriminants Geophys. J, v. 43, p. 269-291.

Although all the above studies indicated that a P-wave spectral discriminant would be effective in identification of explosions, several shortcomings were evident. A very noticeable problem is the large scatter in the computed parameter, which can be as much as two orders of magnitude over earthquakes alone. When this is considered in conjunction with the fact explosions which were studied occurred in only a few, very tectonically-limited, test areas, there is cause for concern because the explosion population reflects only a few paths, and may possibly be biased in attenuation effects or otherwise. A more satisfactory study would present data for earthquakes and explosions in overlapping or common source regions, but the circumstances of known epicenters in Eurasia unfortunately do not allow such analysis to be applied on a large scale. However, for the Nevada Test Site, nearby seismic stations enabled Bakun and Johnson (1971) to study the spectral ratio of signals from earthquakes and explosions originating in a small source retion. While several independent earthquake sources could be discriminated from explosions on the basis of spectral ratio computed from the P_g phase, the aftershock earthquakes associated with high-yield detonations could not be distinguished from explosions of nearly equal m_b on this basis. These results perhaps imply that, when common paths are used, spectral ratio may have little or no value in discrimination. We feel that the paths from earthquakes and explosions studied in Eurasia may have different Q values, thus causing the separation of event types by spectral ratio, and also that Q variability may be a large contributing factor in the scatter observed for earthquake ratios.

Another serious problem with the spectral ratio discriminant is the lack of a unified and accepted theoretical basis. Several theoretical aspects of source behavior interplay: 1) the corner frequency relation with m_b, 2) the high-frequency asymptotic falloff of the spectral amplitude, and 3) the effect of pP cancellation on the spectrum of explosions. In spite of the fact that considerable progress towards resolution of the first two aspects have been made in recent years and that the third aspect can be predicted through well-known calculations, such knowledge may be only of limited use when applied

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Bakun, W. H., and L. R. Johnson (1972), Short-period spectral discriminants explosions, Geophys. J., v. 22, p. 139-152.

to the very narrow spectral range usually included in a spectral ratio definition. In practice, if only a small band is used, the true character of real seismic sources can overwhelm the characteristics predicted from ideal sources. Examples of such are: 1) many earthquakes are undoubtedly accompanied by sub-earthquakes (Blandford, 1975) which distort the spectrum, 2) the source-time function of earthquakes may be very irregular, 3) the explosion spectra will have varying shape due to different P-pP intervals, 4) pP may not be perfectly reflected due to spallation over explosions, and 5) changes in detonation medium and decoupling may greatly affect the exploion corner frequencies.

The brief mention of these theoretical source aspects indicates that some scatter in spectral ratios should be expected due to departures from ideal source conditions. But the chief factor responsible for scatter in spectral ratios may be the variation of Q throughout the earth. That this variations exists radially from the center of the earth presents little problems in treating spectral ratio of P-waves; but the well-documented lateral variations (Barazangi et al., 1975; Molnar and Oliver, 1969; Burton, 1972) must affect spectral ratios considerably in a more random fashion.

In this report we will document the variability of the P-wave spectral ratio discriminant by using a large set of P-waves recorded at the Montana LASA. Such documentation will enable us to grasp the real scatter in this discrimination parameter, to assess whether it is more stable over small source regions, and to make some inferences about Q variations in the mantle. Our purpose is to provide a backdrop of data against which the probable effectiveness of this discriminant used on a global scale could be judged.

Blandford, R.R. (1975), A source theory for complex earthquakes, Bull. Seism. Soc. Am., v. 65, p. 1385-1406.

Barazangi, M., W. Penington, and B. Isacks (1975), Global study of seismic wave attenuation in the upper mantle behind island arcs using pP waves, J. Geophys. Res., v. 80, p. 1079-1092.

Molnar, P., and J. Oliver (1969). Lateral variations of attenuation in the upper mantle and discontinuities in the lithosphere, J. Geophys. Res., v. 74, p. 2648-2682.

DATA

The P-wave signals in this study were those reported in the LASA daily bulletin prepared by the SDAC. A cutoff of roughly 95° distance is used for the LASA bulletin report on epicenters. The signals processed were recorded in the time period of January through August of 1972 and number about 4000 or on the order of 15 per day. The signals were recovered from LASA Event Processor (EP) tapes by association with the LASA bulletin. We recovered only the AO subarray beam on the signal to avoid signal loss which occurs at high frequencies in the full array beam since the signal correlation length decreases with increasing frequency. These signals were already screened by the LASA Detection Processor (DP) which allows only detections (on the full array beam) with a signal-to-noise ratio > 5(14 dB) to be recorded on the EP tapes. Thus our data base consisted of signals which already had fair S/N ratio, assuming that AO was not systematically low in amplitude relative to the full array beam. Further S/N tests, to be described later in this report, were made on the AO subarray signals in our analysis of their spectral content. The sampling rate of the data was 10 samples/second.

Hypocenters used in this report are those reported in the LASA bulletin. Considerable error in the epicenters exists due to the poor location capability of a single array; this error has been estimated theoretically to have a standard deviation of 2.5° by Shlien and Toksoz, 1973. Actual comparisons with NEIS epicenters (Shlien and Toksoz, 1973; Ahner, 1973; and Woolson, 1976) show that the standard deviation is roughly 2°-3°. Depths reported in the LASA bulletin are based on analysts' decisions on the appearance of pP in the LASA record; since these picks are uncorroborated by those from other stations

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Shlien, S. and M.N. Toksoz (1973), Automatic event detection and location capabilities of large aperture seismic arrays, Bull. Seism. Soc. Am., v. 63, p. 1275-1288.

Ahner, R.O. (1973), A comparison of the location refinement techniques in the SDAC LASA event processor, Report No. TR-73-5, Seismic Data Analysis Center, Teledyne Geotech, Alexandria, Virginia.

Woolson, J.R. (1976), unpublished memorandum, Teledyne Geotech, Alexandria, Virginia.

or by multistation hypocenter location, they are not highly reliable. No comparative studies were available with which to quantitatively judge the accuracy of the reported depths of focus in the LASA bulletin though. Furthermore, many pP are undoubtedly missed, and a normal (33 km) depth is arbitrarily assigned by the analyst.

Since both the epicenter and depth estimates from the LASA bulletin are used in presenting various results later in this report, we feel that the errors associated with them will have some adverse impact on those results. This is especially true for depth errors because deep earthquakes may have systematically different spectra than shallow earthquakes. An alternative would have been to collate our LASA signals with NEIS hypocenters, but the effort involved in that procedure did not seem justified in relation to what we felt would be only a small gain in the clarity and importance of the results.

METHODS OF CALCULATION

Signals taken from the EP tapes are shaped by the LASA instrument response which is highly peaked at 4 Hz, and we did not remove this response in out processing of the data since we are concerned with relative spectral values. Figure 1 illustrates the output of our processing for typical P-waves at LASA. For each of the events, the four traces represent outputs of filters with bandpasses, respectively, of 0.4-3.0, 0.4-0.8, 1.4-1.8, and 1.8-4.0 Hz. These filters are applied in the frequency domain by simulating the responses of analog filters with these corner frequencies and with rolloffs of 24 db/oct on either side. The amplitude of the traces is equalized for plotting, with the maximum amplitude in counts given at the left. Because of the peaked instrument response, the filter outputs do not necessarily have their predominant frequencies near the center of the filter bandpass.

Within a four-second window beginning at the maximum amplitude in the 0.4-3.0 Hz filtered P-wave as shown, the maximum amplitude (counts) in each of the four filter outputs for an event was determined. These time-domain amplitudes are the estimates for spectral amplitude within each passband. Previous work on spectral ratios has employed frequency-domain estimates exclusively. But since these estimates were usually based on sample lengths of 10 or even as much as 20 seconds of data following P onset, there was not assurance that the spectral content of the direct P-wave alone was accurately represented in the computed spectra.

The S/N ratio for each passband was computed by forming the ratio of maximum amplitude in the four-second signal window to the maximum amplitude within the noise shown preceding the P-wave. These S/N ratios and the "spectral" amplitudes were then stored along with the LASA hypocenter data. For the results presented in the remainder of this report, any bandpass output from the AO subarray with S/N < 2 was not used. This requirement more than halved the number of available signals.

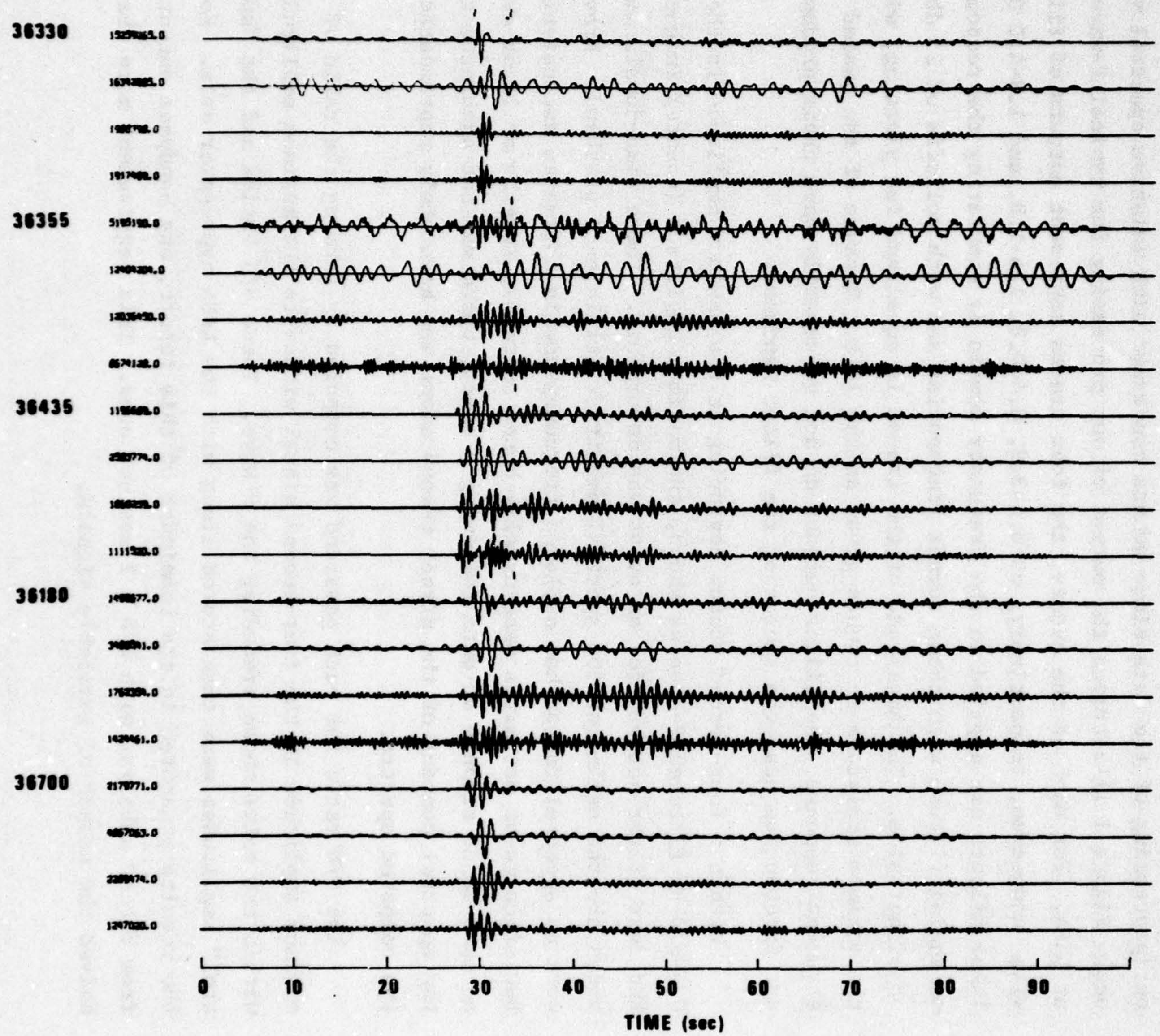

Figure 1 Bandpassed signals recorded at LASA: from top to bottom for each event, 0.4-3.0 Hz, 0.4-0.8 Hz, 1.4-1.8 Hz, and 1.8-4.0 Hz.

RESULTS

Results for the LL Ratio

In this report we define the "LL" spectral ratio as the ratio of maximum amplitude in the 0.4-0.8 Hz bandpass output to that in the 1.4-1.8 Hz output. This is a variant of the original Briscoe (1966) spectral ratio, and it is essentially the inverse of that presented in many reports by the Applied Seismology Group at Lincoln Laboratory.

Since our purpose is to examine the variation of this spectral ratio over different regions of the earth, it is desirable to investigate whether any systematic effects are present in the computed ratios. One effect that follows from source theory is that the spectra within our frequency band should be a function of source spatial extent, or magnitude, if the stress drop is fixed. Most source theories predict a teleseismic pulse with a flat low-frequency spectrum falling off as f^{-2} or f^{-3} above a corner frequency which decreases with increasing magnitude. For a f^{-2} model, the difference in the LL spectral ratio between very low and very high magnitude events should be the difference between amplitude ratios made at the center frequencies of 0.6 and 1.6 Hz for a flat (low m_b) and a f^{-2} (high m_b) spectral shape respectively. This amounts to a factor of about seven. Thus the LL ratio as defined above should be nearly an order of magnitude greater for large events compared to small events. In figure 2 are presented all LL ratios from our data base for which the S/N ratio exceeded 2.0 in both bandpass outputs. Also only shallow events (<60 km, as reported on LASA bulletin) were used. This figure shows no apparent trend for the LL ratio with increasing m_b, except beyond 5.5 perhaps; apparently in disagreement with the above argument. One reason for the real data not to follow theoretical trends may be that the small sample of events above m_b=5 is biased. Another reason may be that the instrument response, peaked at 3-4 Hz, distorts the time-domain estimate of spectral amplitude after application of narrow-band filters by leakage of higher freq-

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Briscoe, H. W. (1966), Ratios of spectral densities, in Semiannual Technical Summary--Seismic Discrimination, 31 December 1966, Lincoln Laboratory Lexington, Massachusets.

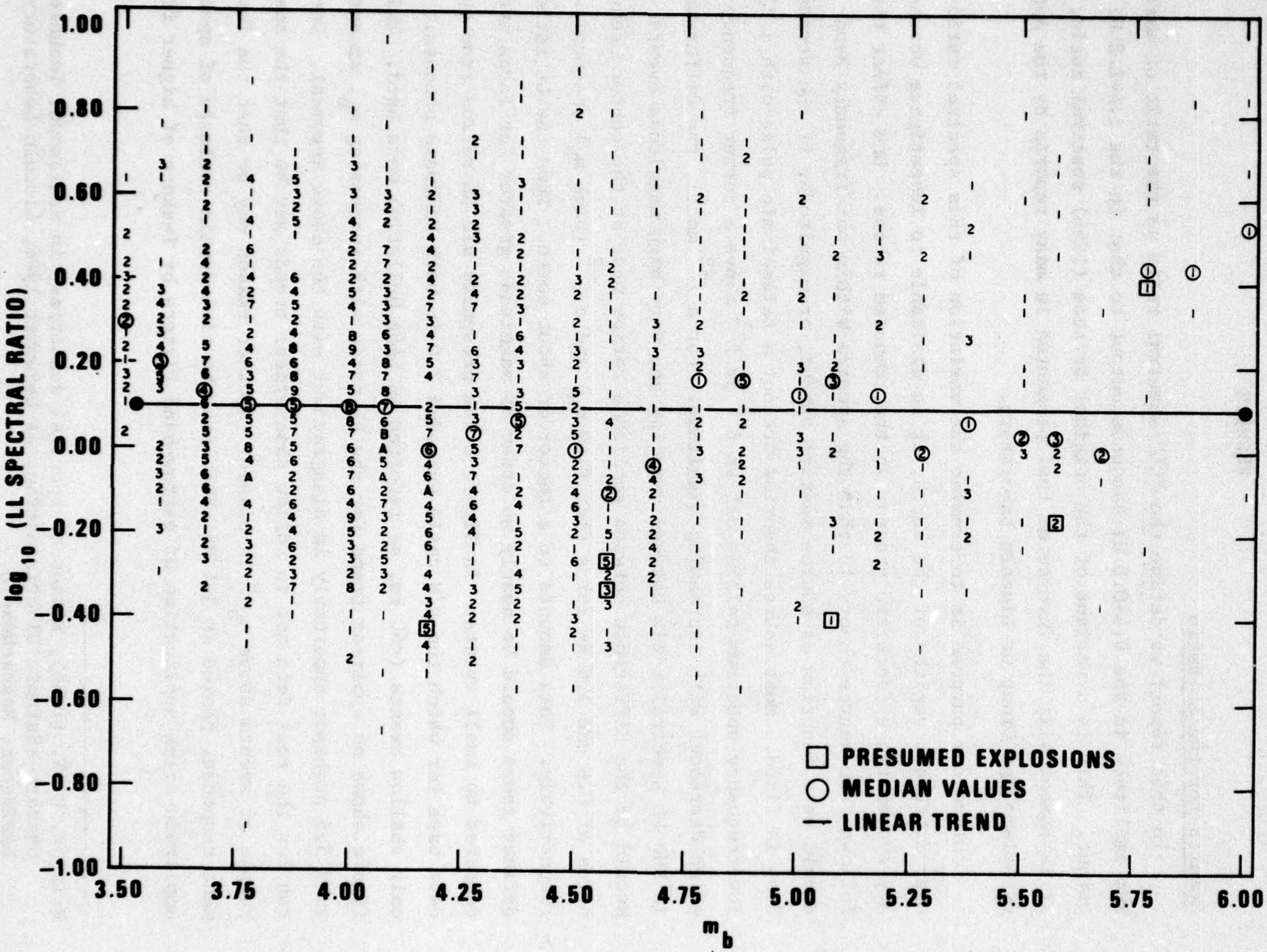

Figure 2   LL spectral ratio versus m_b for events with DOF < 60 km and $\Delta < 95^o$ from LASA.

uency components into lower bandpass outputs. A check of several events such as those in Figure 1 revealed that the dominant periods within the signal window for the second and third traces were always close to the center frequencies of 0.6 and 1.6 Hz though, and so this latter reason seems precluded. A remaining possibility is that the sample of ratios from lower magnitude events are heavily weighted with short epicentral distances and that a bias arises due to different attenuation effects between near and far events. This possibility can be checked by plotting the data of Figure 2 versus epicentral distance as in Figure 3. Here we have restricted the data base to $m_b > 4.6$ in order to insure essentially complete detection of events out to $\Delta = 95°$ for the AO subarray beam. This m_b cutoff is based on a full array 90% threshold of $m_b \simeq 3.9$ (Dean, 1971) up to $\Delta = 80°$ and a beaming loss of $\sqrt{M}$ in the S/N ratio where $M = 25$ in going from the full LASA array composed of 25 subarrays to the AO subarray. Although considerable scatter exists in the data of Figure 3, there is no indication of a trend with distance; and inference of a nearly constant attenuation factor is not unreasonable. This implies constant t* in the equation

$$A_o(f) = A_s(f)\ e^{-\pi f t^*} \cdot G(\Delta)$$

where subscripts o and s mean observed and source spectrum and $G(\Delta)$ is the geometrical spreading factor which is assumed to be independent of frequency. The parameter t* is given by

$$t^* = \int_s \frac{ds}{\alpha Q}$$

where α is the compressional-wave velocity, Q is the quality factor for compressional waves, and s is the path length.

Thus, we eliminate the possibility of significant distance bias in the results of Figure 2. The lack of empirical trend with m_b for the LL ratio is then unexplained, and we can only conclude that some unapparent bias in the data exists or that there is really a negligible average change with m_b. The negligible change could more easily be accommodated by the f^{-1} spectral slope at intermediate frequencies suggested by Brune (1970) in the case of partial

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Dean, W.C. (1971), Detection threshold of the LASA/SAAC system, Report No. 3, Seismic Array Analysis Center, Teledyne Geotech, Alexandria, Virginia.

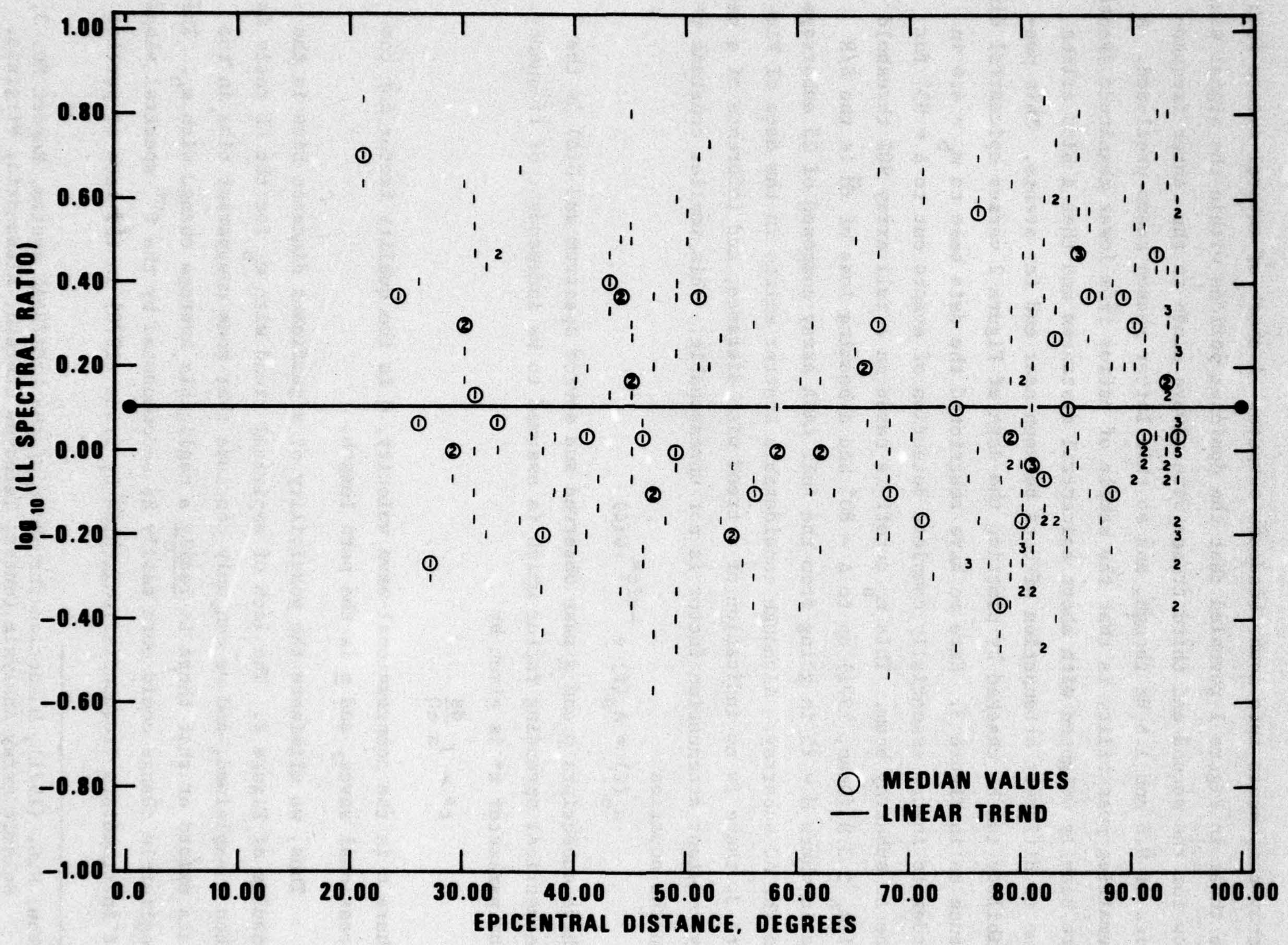


Figure 3 LL spectral ratio versus epicentral distance from LASA for events with $m_b \geq 4.6$ and DOF < 60 km.

stress drop, which is probably the rule rather than the exception for earthquakes. Another possibility is that numerous subearthquakes (Blandford, 1975) make the high frequency spectrum have a smaller slope. Still another possibility is that for each fixed m_b the partial stress drops for all earthquakes at that m_b are widely varying. Thus at each m_b there would be a wide range of corner frequencies, obscuring a trend toward more low frequencies at high magnitudes.

Together Figures 2 and 3 suggest that no systematic effects are contained in our computed spectral ratios. If these ratios are considered on a regional basis, then any variation from region to region should be reflective of real difference either in source spectral character or in the parameter t* along the particular paths to LASA. In Figure 4 are shown the values $10 \cdot \overline{\log_{10}(LL)}$ where $\overline{\log_{10} LL}$ is the average of the logarithm of the spectral ratios for events within given geographic regions, as configured by Flinn et al. (1974). Only regions within roughly 100° of LASA are represented due to the LASA bulletin cutoff, and one would not want to consider more distant regions anyway since effects of the core are present on P-waves from epicenters beyond roughly 95°. Many regions produced no data for this plot, and also regions with only one signal were ignored. A plot of the 166 geographic region numbers represented in Figure 4 is shown in Figure 5 for reference; the region numbers are plotted near the centroid of each region. We note that for region 329 in Central Asia the value plotted in Figure 4 is based on four presumed underground explosions and cannot be compared to earthquake results. Table I lists the data used in plotting Figure 4 and also the sample size NEQ and standard deviation SD of the individual data for each region. The approximate confidence limits for the mean $\overline{\log_{10}(LL)}$ in each region is given by $\pm(2*SD)/\sqrt{NEQ}$, and it is apparent that much of the variation seen in Figure 4 is well established by statistical measure. Examination of a map of SD values similar to Figure 4 but not shown here revealed no systematic variation clearly associated with tectonic character.

Blandford, R.R. (1975), A source theory for complex earthquakes, Bull. Siesm. Soc. Am., v. 65, p. 1385-1406.

Flinn, E.A., E.R. Engdahl, and A.R. Hill (1974), Seismic and geographical regionalization, Bull. Seism. Soc. Am., v. 64, p. 771-992.

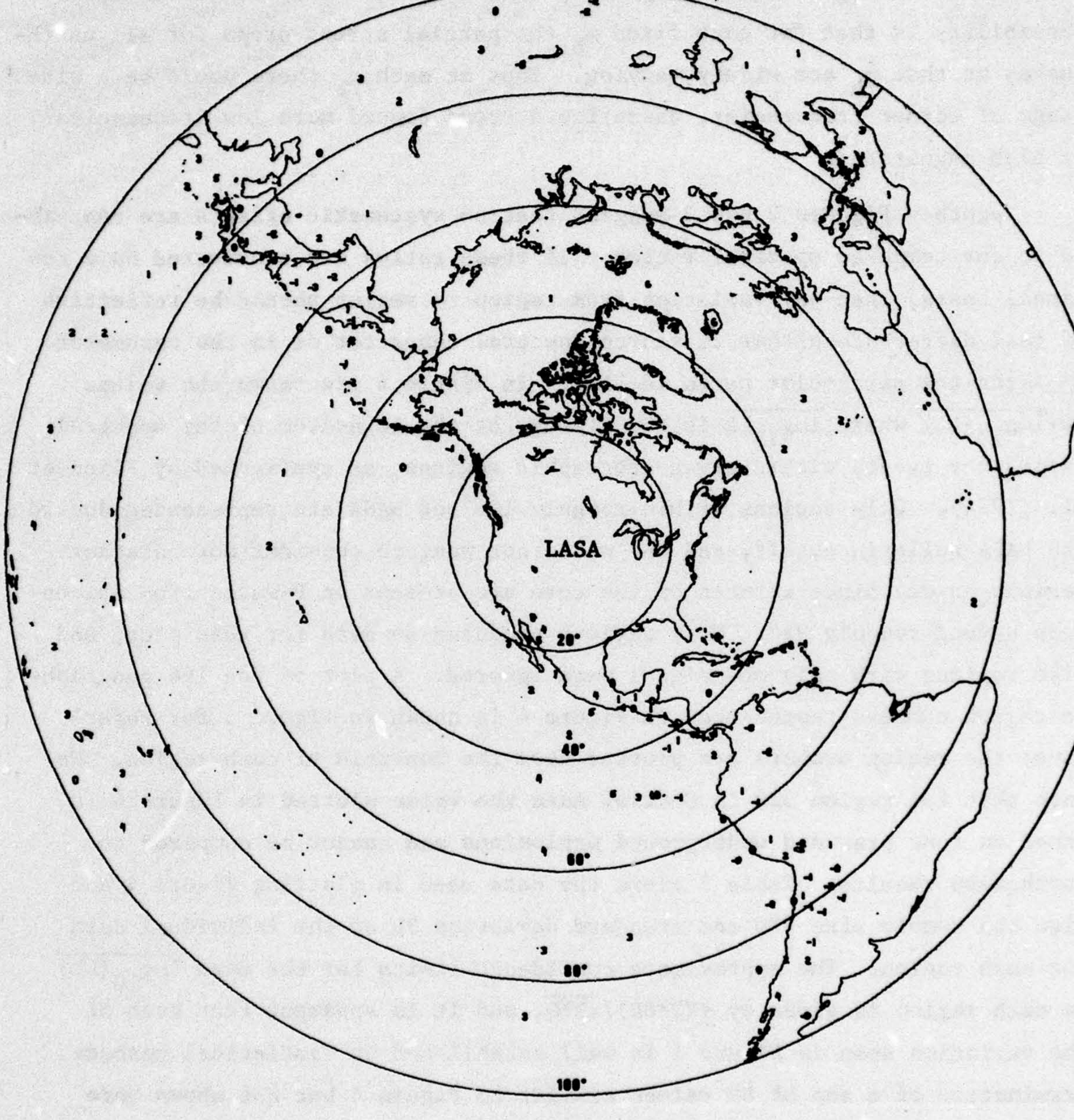
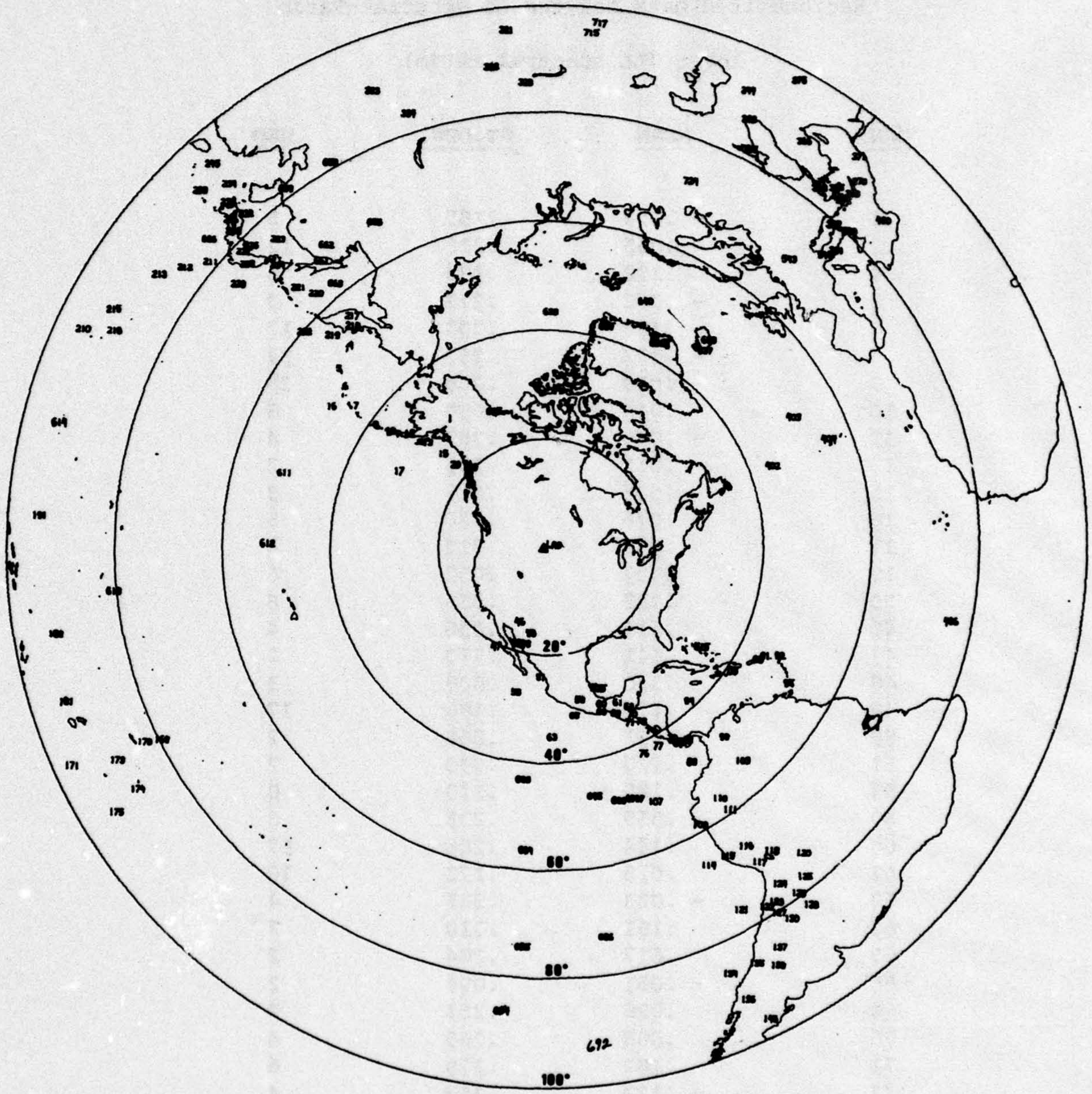


Figure 4 **Regional averages of $\log_{10}$(LL) for events with depth of focus < 60 km.**



Figrue 5   Geographic regions for which average $\log_{10}$(LL) was determined.

TABLE I

Regionalized Data for the LL Spectral Ratio

$\log_{10}$ (LL spectral ratio)

GRN	MEAN	ST.DEV.	NEQ
1	- .058	.185	5
2	- .055	.218	12
4	.192	.193	6
5	- .072	.277	3
6	- .150	.351	17
7	- .122	.252	38
9	- .054	.266	28
10	- .019	.080	8
12	- .049	.362	4
13	- .262	.105	2
15	- .096	.117	3
16	.076	.220	5
17	.042	.212	4
19	.307	.053	6
20	.413	.355	8
46	.514	.058	4
47	.341	.770	2
48	.370	.008	3
49	.473	.186	17
50	.397	.068	2
51	.120	.090	2
53	.185	.323	8
59	.355	.235	3
60	.144	.286	17
61	.075	.172	16
62	- .023	.335	4
63	.161	.310	7
65	.617	.284	2
66	- .061	.094	2
69	.026	.231	9
70	.068	.265	4
71	.182	.279	6
73	- .123	.367	4
74	- .064	.247	7
76	- .066	.260	19
77	.083	.324	4
78	- .099	.314	4
80	- .097	.042	2
81	- .096	.057	2
83	.166	.383	11
88	.087	.033	3

TABLE I (Continued)

Regionalized Data for the LL Spectral Ratio

$\log_{10}$ (LL spectral ratio)

GRN	MEAN	ST.DEV.	NEQ
90	- .243	.319	2
91	- .210	.097	2
92	- .060	.334	6
94	.031	.185	4
95	- .222	.075	2
99	- .342	.181	18
103	- .171	.201	13
107	- .077	.133	4
109	.015	.202	2
110	.096	.228	5
111	- .096	.141	7
114	- .046	.172	3
115	.027	.271	14
116	.067	.263	8
117	.179	.250	3
118	- .155	.094	6
120	.051	.480	2
121	- .047	.176	2
122	.038	.203	9
123	.077	.182	6
124	- .059	.269	7
125	- .040	.371	8
127	- .090	.329	8
128	- .202	.070	3
129	- .100	.153	12
130	- .091	.351	2
134	.309	.218	7
135	.197	.216	14
136	.396	.518	5
137	- .091	.111	5
139	.174	.351	2
146	- .002	.166	3
169	.387	.237	14
170	.497	.105	4
171	.044	.213	14
173	.291	.234	38
174	.429	.243	13
175	.432	.444	4
181	.260	.257	30
183	.214	.245	15
191	.352	.177	3

TABLE I (Continued)

Regionalized Data for the LL Spectral Ratio

$\log_{10}$ (LL spectral ratio)

GRN	MEAN	ST.DEV.	NEQ
210	.167	.198	7
211	- .013	.254	73
212	.091	.223	10
213	.097	.136	11
215	.113	.194	9
216	.181	.305	7
217	.109	.232	9
218	- .107	.213	20
219	.022	.201	20
220	- .277	.053	3
221	- .107	.197	42
222	.011	.209	11
223	- .168	.262	4
224	- .013	.208	10
226	- .144	.274	2
227	.018	.147	5
228	- .139	.328	12
229	.051	.205	41
232	- .042	.024	3
233	- .127	.197	7
234	.473	.300	3
235	.297	.365	3
236	.126	.454	2
238	.207	.037	2
245	.331	.181	12
321	.019	.209	17
323	.349	.308	3
329	- .288	.146	4
332	.339	.223	7
334	.188	.414	2
344	.099	.158	4
357	- .288	.156	2
362	.309	.056	3
363	.006	.098	6
364	.299	.261	8
365	.136	.230	3
366	.195	.226	21
368	.438	.166	3
370	.120	.181	6
371	.262	.062	5
375	.078	.099	4
381	.246	.342	4

TABLE I (Continued)

Regionalized Data for the LL Spectral Ratio

$\log_{10}$ (LL spectral ratio)

GRN	MEAN	ST.DEV.	NEQ
382	.013	.146	2
390	.208	.139	3
400	.144	.297	4
402	.243	.247	47
403	.344	.206	24
404	.219	.161	6
406	.234	.246	32
515	.036	.030	2
525	.290	.275	8
543	.286	.140	2
545	.016	.041	3
611	.133	.144	3
612	.228	.069	4
614	.324	.144	4
618	.358	.008	2
633	.074	.231	7
635	.097	.220	4
636	.446	.325	5
637	.390	.199	12
638	.636	.129	2
640	.575	.257	6
656	.145	.261	5
658	.004	.075	2
659	- .145	.357	4
661	- .117	.016	2
662	.150	.436	2
663	- .176	.198	7
666	.496	.230	2
670	.106	.169	2
677	.174	.150	4
684	.293	.218	9
685	.262	.264	16
686	.252	.023	2
692	.320	.319	6
693	.354	.163	18
694	.372	.278	26
695	.572	.255	11
696	.260	.311	2
697	.348	.210	2
715	- .201	.206	7
717	- .175	.000	2
724	- .262	.090	3

Results for the HF Ratio

Before interpreting the results contained in Figure 4, we will examine another spectral ratio defined using our highest bandpass filter from 1.8-4.0 Hz. The ratio of the maximum amplitude in the 0.4-0.8 Hz output to that in this output is termed here the "HF" spectral ratio. Presumably, if Q effects are highly variable, this ratio should have even greater scatter than the LL ratio. Figure 6 is a plot of this data, with the restriction of S/N ratio > 2.0 for both traces as for the LL ratio. Comparison of Figure 6 with Figure 2 indicates no significant increase in scatter, and so effects other than Q on the spectra must be large. The fitted line in Figure 6 does indicate a change in spectral ratio with m_b, as predicted from source theory, but not nearly as large as expected.

Figure 7 shows the data of Figure 6, with a cutoff below $m_b = 4.6$, plotted versus epicentral distance; and, as for the LL ratio in Figure 3, no systematic trend with distance is apparent for the HF ratio.

Figure 8 presents the regional averages for the logarithms of the HF ratio, and Table II lists these data. Only 134 regions are represented here, 32 less than for the LL ratios in Figure 4, because many more 1.8-4.0 Hz bandpass outputs were screened out by the S/N ratio requirement than 1.4-1.8 hz outputs. The regional pattern shown in Figure 4 is essentially repeated by Figure 8.

Q Variations

Neglecting all other contributions to the scatter in the data, let us focus on the possible range in Q, or rather t*, if this parameter alone caused the observed variation in the spectral ratios. (Later we will show why t* alone cannot explain the scatter.) By using equation (1) above, the difference between two observed values of LL would be

$$(\log LL_1 - \log LL_2) = \pi(f_H - f_L)\ (t_1^* - t_2^*)/\log_e 10 \qquad (2)$$

where f_H and f_L can be taken as the center frequencies of the two passbands used in obtaining LL, namely, 0.6 and 1.6 Hz. Using these frequency values and the maximum spread in $\log_{10}(LL)$ of roughly 1.6 from Figure 2 for the left-hand side of (2), we can estimate the maximum t* spread as roughly $(t_1^* - t_2^*)$

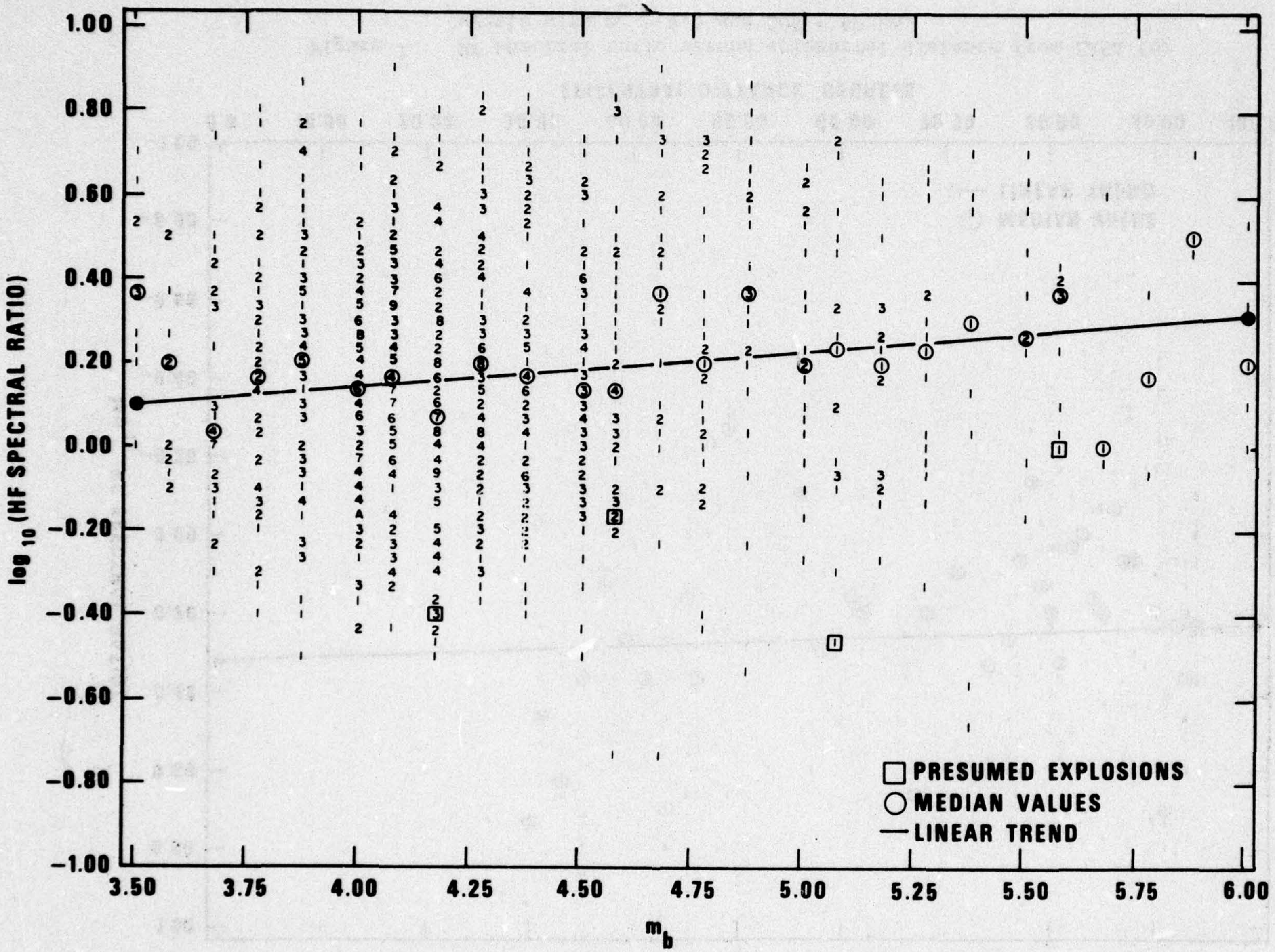


Figure 6 HF spectral ratio versus m_b for events with DOF < 60 km and Δ < 95° from LASA.

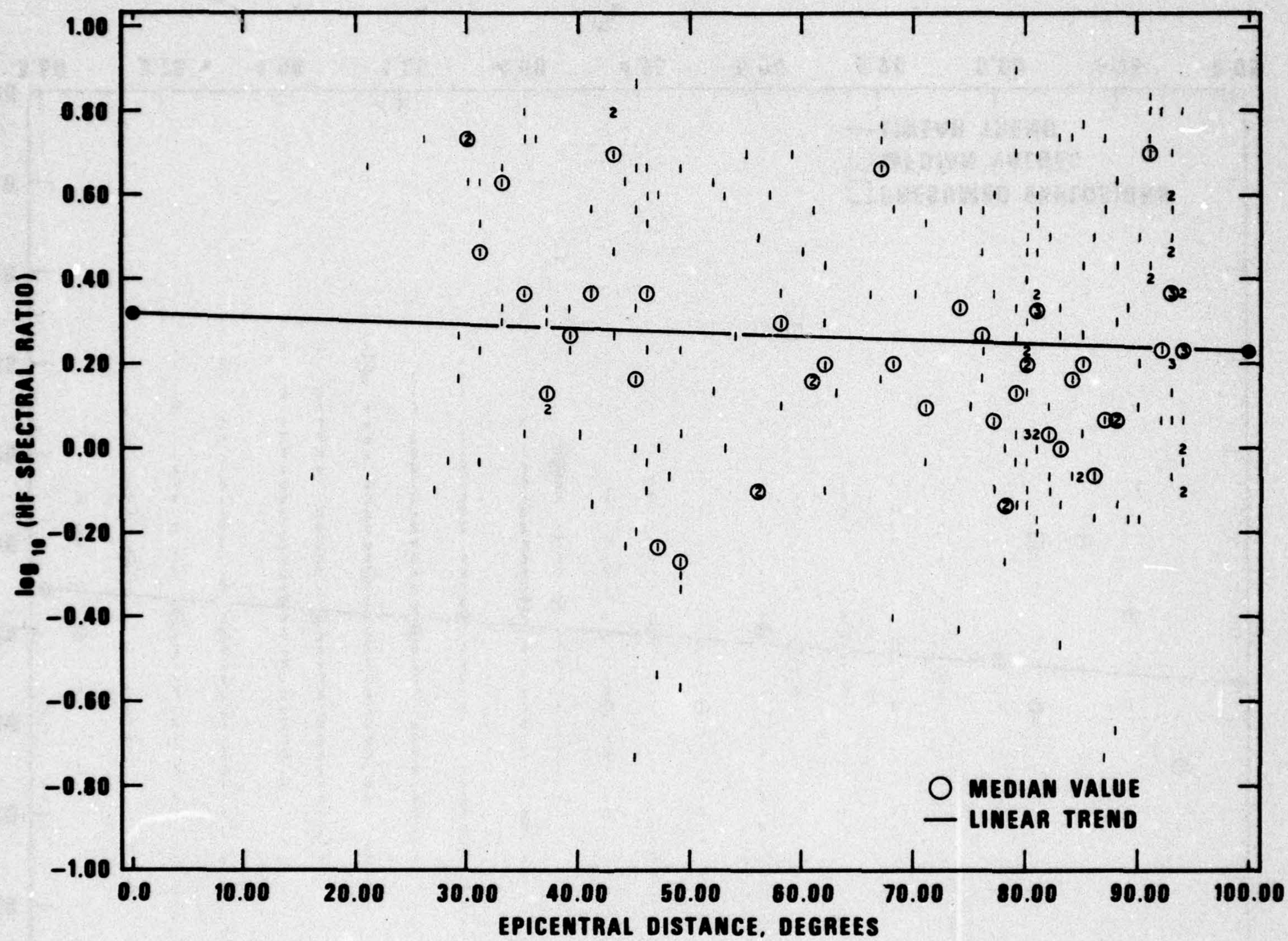


Figure 7 HF spectral ratio versus epicentral distance from LASA for events with $m_b \geq 4.6$ and DOF < 60 km.

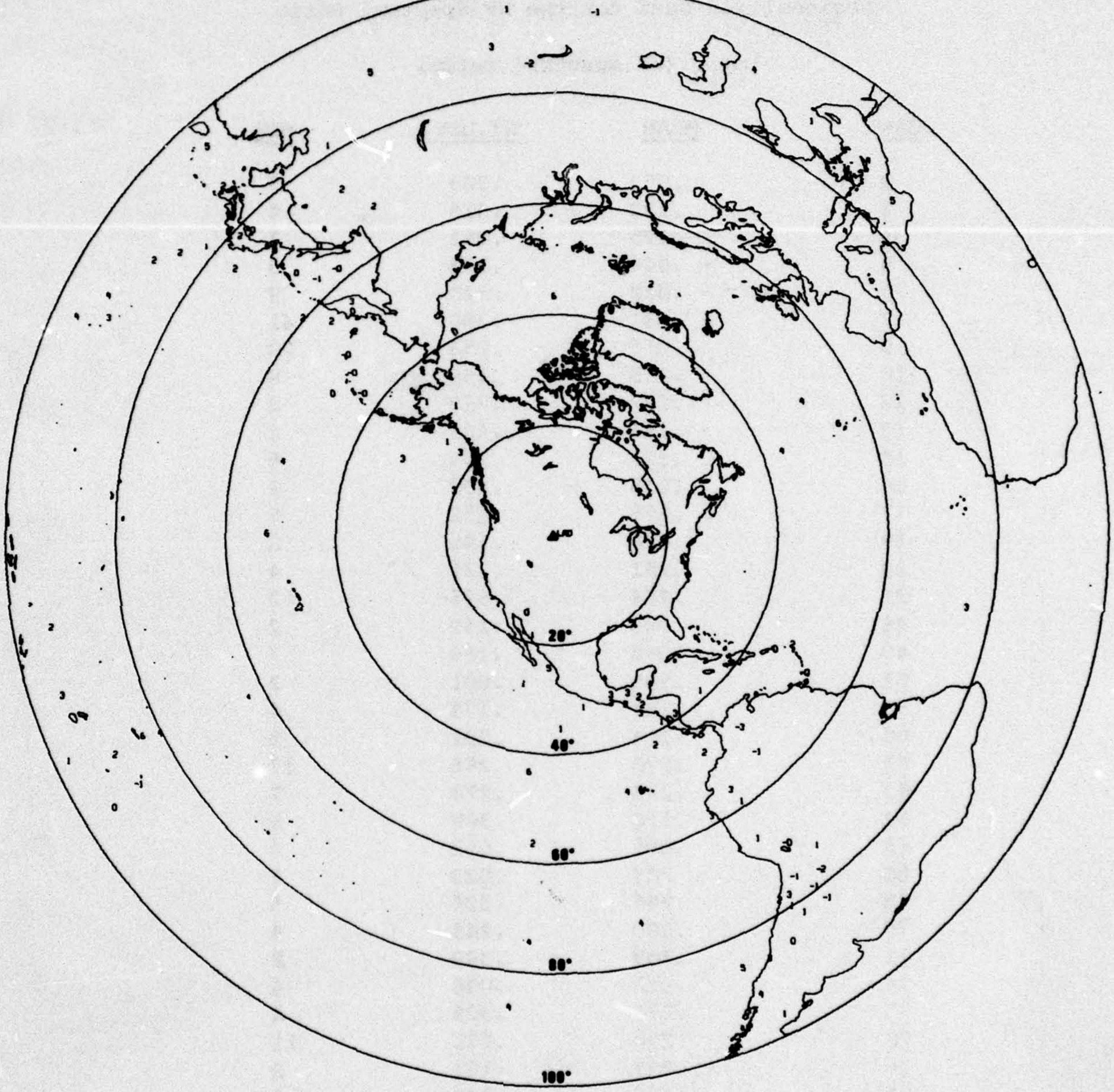


Figure 8   Regional averages of $\log_{10}$(HF) for events with DOF < 60 km.

TABLE II

Regionalized Data for the HF Spectral Ratio

$\log_{10}$ (HF spectral ratio)

GRN	MEAN	ST.DEV.	NEQ
1	.052	.209	7
2	.132	.315	6
4	.175	.061	7
5	- .044	.133	3
6	- .022	.340	9
7	- .031	.280	41
9	.015	.258	20
10	.155	.158	8
12	.047	.379	2
13	.184	.401	2
15	.127	.192	6
16	.015	.497	2
17	.275	.252	5
19	.626	.195	3
20	.251	.337	4
21	.254	.635	2
46	.005	.152	2
49	.090	.180	7
51	.503	.001	2
53	.111	.179	4
60	.264	.292	9
61	.276	.266	17
62	.234	.228	7
63	.118	.368	5
65	.085	.052	2
66	.263	.023	2
69	.266	.304	5
70	.200	.245	4
71	.349	.340	8
74	.082	.096	9
75	.276	.329	4
76	.206	.271	15
78	- .237	.101	2
81	.176	.149	2
83	.207	.356	9
90	- .018	.389	2
92	.164	.422	7
94	.149	.340	5
95	- .028	.075	3
99	- .303	.224	18
103	- .096	.267	13
109	.251	.423	4

TABLE II (Continued)

Regionalized Data for the HF Spectral Ratio

$\log_{10}$ (HF spectral ratio)

GRN	MEAN	ST.DEV.	NEQ
110	.292	.377	4
111	.054	.167	9
115	.136	.240	15
116	.063	.338	4
118	-.030	.156	8
120	.120	.474	3
122	.163	.292	6
123	.298	.213	5
124	-.121	.127	2
125	-.190	.235	8
127	.124	.509	6
128	-.072	.048	3
129	-.055	.170	17
130	.127	.484	2
134	.475	.212	6
135	.252	.230	9
136	.376	.336	6
137	.014	.269	5
146	.108	.102	2
169	.442	.031	4
170	.597	.002	3
171	.129	.136	11
173	.232	.282	16
174	-.063	.496	8
175	.046	.008	2
181	.282	.263	18
183	.172	.318	4
210	.385	.166	3
211	.143	.228	67
212	.243	.270	5
213	.232	.184	9
215	.380	.304	7
216	.340	.235	7
217	.286	.268	6
218	.044	.226	23
219	.226	.265	18
220	-.080	.112	3
221	.037	.218	33
222	.184	.265	7
224	.028	.223	13
226	-.031	.397	2
227	.206	.155	7

TABLE II (Continued)

Regionalized Data for the HF Spectral Ratio

$\log_{10}$ (HF spectral ratio)

GRN	MEAN	ST.DEV.	NEQ
228	.120	.263	10
229	.204	.215	30
230	.000	.180	2
232	.191	.128	4
233	.253	.313	8
235	.158	.187	2
245	.475	.268	9
321	.107	.145	10
323	.516	.240	2
329	-.241	.221	4
332	.347	.343	4
344	.389	.091	2
357	-.181	.017	2
363	.378	.136	5
364	.358	.389	5
365	.149	.067	2
366	.144	.209	10
368	.502	.111	3
370	.384	.208	5
381	.386	.337	6
400	.516	.333	2
402	.422	.273	32
403	.379	.241	13
404	.563	.155	5
406	.345	.250	14
525	.354	.136	5
611	.366	.085	3
612	.557	.137	2
633	.571	.255	4
635	.506	.332	3
636	-.221	.467	3
637	.320	.502	2
640	.185	.245	2
656	.232	.423	5
657	.190	.558	2
658	.095	.051	2
659	.354	.121	2
661	.024	.000	2
662	.386	.337	3
663	-.045	.381	5
677	.441	.261	3

TABLE II (Continued)

Regionalized Data for the HF Spectral Ratio

$\log_{10}$ (HF spectral ratio)

GRN	MEAN	ST.DEV.	NEQ
682	.214	.007	2
683	.356	.099	2
684	.353	.162	5
685	.393	.176	9
692	.497	.185	3
693	.636	.157	5
694	.186	.324	12
696	.403	.034	2
715	- .106	.000	2

≤ 1.2. A similar calculation for the HF ratio using a spread of 1.6 from Figure 6 and f_H = 2.9 hz gives roughly $(t_1*-t_2*) \leq .5$. However, the data in Figure 6 do not include many events which appeared in Figure 2 (1022 versus 1438) because the higher bandpass output did not pass the S/N test. The ratios in these cases would generally scatter towards the top of the measurable HF ratios shown in Figure 5 and probably increase the total scatter significantly if there had been S/N > 2.0 on the 1.8-4.0 Hz output. Also the high frequency band probably cannot be characterized by its center frequency of 2.9 hz since the spectrum changes so rapidly in the band 1.8-4.0 Hz. Thus, we regard $\Delta t* \leq 1.2$ estimated from the LL ratios as more reliable. If we assume a minimum possible t* of 0.2 for a very high Q path to LASA (Der, 1976), then the maximum t* is 1.4, giving a range which is not unacceptable. However, we note that much of the scatter in Figure 2 must be source related (pP interference, corner-frequency variation, time-function variation, etc.); and so the inferred range of t* is surely less than 1.2.

Scatter and Regional Variation in Spectral Ratios

In relation to discrimination, the outstanding feature of the data presented here is the large scatter in the spectral ratios as best seen in Figures 2 and 6. Roughly the same scatter can be found in the LL ratio for 156 Eurasian earthquakes analyzed by Lacoss (1969). This scatter ranges up to 1.6 logarithmic units and is more than half as large as that in M_s-m_b observed at LASA at a given magnitude for earthquakes from several regions (Capon et al., 1969). This occurs for the LL ratio even though we have included only a very small spectral bandwidth in the estimate of this ratio, 0.4 to 1.8 Hz. The scatter in LL and HF ratios can be explained by acceptable variations in t*, or Q, over the various paths to LASA although many source-related effects must contribute also.

It is reasonable to presume that these spectral variations are coherent

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Lacoss, R.T. (1969), A large-population LASA discrimination experiment, Technical Note 1969-24, Lincoln Laboratory, Lexington, Massachusetts.

Capon, J., R.J. Greenfield, and R.T. Lacoss (1960), Long-period signal processing results for the Large Aperature Seismic Array, Geophysics, v. 34, p. 305-329.

over some range of azimuth and distance from LASA, i.e., are regional in character. Figures 4 and 8 can be interpreted as a verification of this presumption, especially since examination of these maps reveals a similarity between the location of high and low attenuating source regions defined by the spectral ratios and the location of such regions from other types of studies (P and pP waves by Barazangi et al., 1975; S waves by Molnar and Oliver, 1969). Oceanic-ridge source regions have consistently high ratios in our figures. This suggests, but does not demand, high attenuation, as also inferred by Molnar and Oliver. Due to the large errors in the LASA epicenters, it is not possible to clearly distinguish the adjacent high and low attenuation zones which were well defined by Barazangi et al. in regions of plate subduction. Also, most of the events east of Chile, and in other regions characterized by deep events, must be deeper than 60 km; but pP was not detected by the LASA analyst, which in itself suggests higher attenuation in the zone above the earthquakes. Since our method of analysis was not designed to study attenuation directly, only crude inferences can be made from our spectral ratios.

Because much of the overall scatter in the LL or HF spectral ratio is due to regional variations, this scatter could be reduced by regional corrections as has often been suggested for M_s-m_b discrimination. Yet, Tables I and II reveal that considerable scatter still exists for values within a given region. A typical standard deviation is perhaps 0.20 for the logarithms of a region's ratios while that for all the ratios in both Figures 2 and 6 is 0.30. While we must recall that the errors in location by LASA will cause some intermixing of values from adjacent regions, we fully expect that most of this scatter would remain with highly accurate locations since it is due to variations in source character and depth of focus and to lateral inhomogeneities within a given region. Moreover, if the parameters M_s

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Barazangi, M., W. Penington, and B. Isacks (1975), Global study of seismic wave attenuation in the upper mantle behind island arcs using pP waves, J. Geophysics. Res., v. 80, p. 1079-1092.

Molnar, P. and J. Oliver (1969), Lateral variation of attenuation in the upper mantle and discontinuities in the lithosphere, J. Geophys. Res., v. 74, p. 2648-2682.

and m_b are any guide, subdividing into very small source regions may not greatly reduce the scatter within each of these small regions; e.g., the study of LASA M_s-m_b in three Kamchatka subregions by Blandford and Clark (1975) showed that at least 1.5 orders of magnitude variation was present in M_s at a given m_b for shallow earthquakes all within an area defined by a circle with roughly 1° diameter.

Explosion Ratios

In Figures 2 and 6 we have identified those spectral ratios for seven presumed underground nuclear explosions in the USSR. The epicenters were identified as possible explosions by one or more of these criteria: 1) assigned zero depth on the NEIS list, 2) their appearance in the Uppsala, Sweden, reports on Soviet explosions, and 3) their low M_s values in von Seggern (1976). In general these explosion points, either the LL or HF ratios, lie among the very lowest of computed values for all the earthquakes. We note again that our data is probably contaminated with a large number of deep events, but not reported as such on the LASA bulletin; these events may contribute heavily to the lower spectral ratio values in Figure 2 and 6. We must not established the degree of attenuation along the paths from the Soviet test sites to LASA relative to other paths. In our data base, the number of natural events in those geographical regions which include the Soviet explosion sites in Kazakh, the Urals, and above the Black Sea are insufficient to determine t* with any confidence. From spectral analysis of explosions though, von Seggern and Sobel (1976) have determined t* for Kazakh to LASA to be roughly 0.4, a value which alone cannot account for the appearance of the explosion points so low within the global sample of earthquakes.

Blandford R.R. and D.M. Clark (1975), Variability of seismic waveforms at LASA from small subregions of Kamchatka, Report No. TR-75-12, Seismic Data Analysis Center, Teledyne Geotech, Alexandria, Virginia.

von Seggern, D.H., (1976). Final report on the analysis of the VLPE network Report No. TR-76-1, Seismic Data Analysis Center, Teledyne Geotech, Alexandria, Virginia.

von Seggern, D.H., and P.A. Sobel (1976). Study of selected Kamchatka earthquakes in a seismic discrimination context, Report No. TR-76-__, Seismic Data Analysis Center, Teledyne Geotech, Alexandria, Virginia.

In Figure 9 we present the explosion LL ratios with only those ratios from earthquakes in a limited area composed of 10 geographic regions around the USSR explosion sites. These regions together comprise an area having no known rift or subduction zones, except for the Hindu Kush seismic zone, and which should therefore be rather homogeneous in regard to Q properties. The seven explosion points do not separate well from the earthquake population. Evidence that the remaining points are indeed earthquakes and not explosions comes from their M_s-m_b values (von Seggern, 1976) which are not anomalous and lie well above those of the seven explosions. Also, we attempted to collate all earthquakes with logLL < 0 in this figure with the NEIS list. The collated ones are circled, and in all cases the NEIS depth was < 70 km. Thus low earthquake spectral ratios are not due to deep Hindu Kush events.

Corrections to m_b Based on Observed Spectral Ratios

In this section, an attempt will be made to utilize the average spectral ratios for various geographic regions, as determined in this report, for estimating an m_b correction. The premise that the spectral ratios are an indicator of relative attenuation has already been shown to be tenable by its correlation with known geophysical data. There are several factors contributing to the spectral ratio parameter and to m_b such as stress drop, corner frequency, and incoherence in the source time function; nonetheless, we expect that our spectral ratio results can be helpful in reducing the scatter of m_b, as estimated at LASA, for a given true magnitude. Since the true magnitudes are unavailable, we take M_s as the reference magnitude. We assume without evidence that M_s variations do not correlate with the m_b variations. This should be especially correct since we will use network-determined M_s values (von Seggern, 1976) for which LASA was not a reporting station.

Mapping the spectral ratio into an m_b correction is done by first using equation (2) to obtain the t* variation. For those regions which had enough events to calculate an average logLL as given in Table I, we see that the spread is roughly 1.0 log unit, which when substituted in (2), gives a Δt^*

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von Seggern, D.H. (1976), Final report on the analysis of the VLPE network, Report No. TR-76-1, Seismic Data Analysis Center, Teledyne Geotech, Alexandria, Virginia.

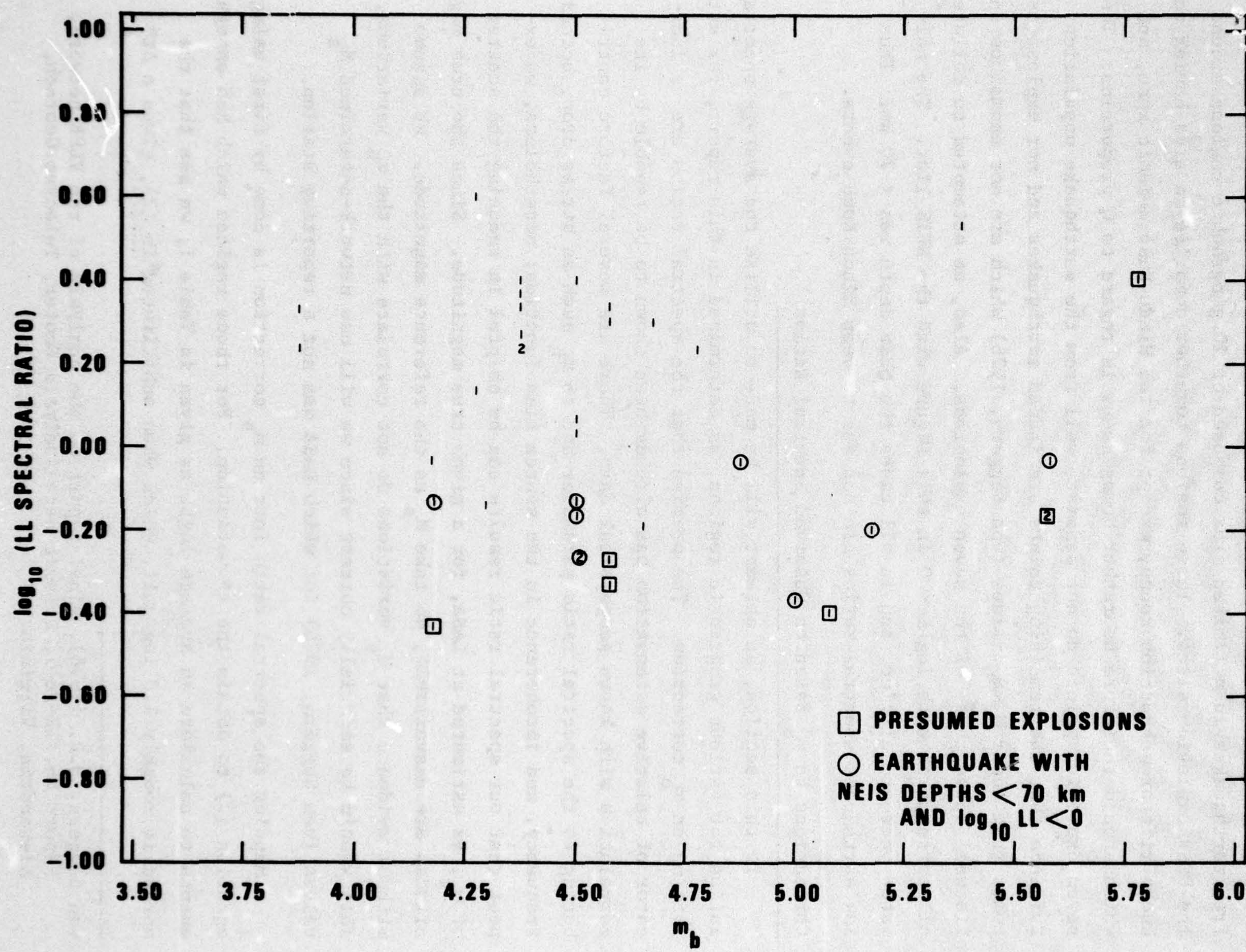

Figure 9 LL spectral ratios for events in geographic regions 321, 329, 332, 336, 344, 357, 362, 715, 717 and 724 (see Figure 5).

of 0.75. Substituting this Δt^* value in equation (1) and taking logarithms results in an Δm_b (assuming f = 1 Hz) of approximately 1.0 between signals along the lowest and highest attenuating paths to LASA. Thus, we propose this simple correction to LASA m_b's:

$$m_b' = m_b + \overline{\log_{10} LL} \tag{3}$$

The proposed corrections then are identical to the numbers listed in Table I. We will use only those regions (108) which have NEQ $\geq$ 4 in this table however.

The LASA m_b's were collated with the VLPE M_s values; this was possible for only 349 events since the epicenter lists were different and since an M_s was not reported by the VLPE network for many events within the 95° ring surrounding LASA. In Figure 10a are plotted separately the originally reported LASA m_b's and the m_b"s corrected according to relation (3) versus M_s. As clearly represented in the plots, the calculated standard deviation of the data has not been reduced. The standard deviations are the results of maximum-liklihood line fitting under the assumption of equal variance in M_s and m_b. Even though some of the scatter of the data in Figure 10a is due to M_s, that part due to m_b, and therefore the overall scatter, was expected to be reduced. This result suggests that the spectral ratios have little actual correlation with attenuation differences inferred from our analysis of spectral ratios versus geographic region. To affirm this result, we extracted only those events from geographic regions with an m_b correction given by relation (3) of $\overline{\log LL} > 0.3$ or $\overline{\log LL} < -0.1$. This left only 110 events from 21 geographic regions, and their M_s-m_b plot are shown in Figure 10b. Again, even though we have taken only that data which is subject to large corrections in m_b, the scatter of the data is not significantly reduced. It may be that have over-corrected the m_b values with relation (3); to test this idea we replotted the data of Figures 10a and 10b with corrections to m_b exactly one-half those already used, but no reduction in scatter occurred.

If spectral ratio were an indicator of attenuation differences and thus variation in observed m_b for a given source spectrum, then we would fully expect reduction in m_b scatter. In fact, we have not demonstrated any improvement; and this is especially unexpected because the spectral ratios of p-waves from various regions seemed to correlate well with direct or indirect geophys-

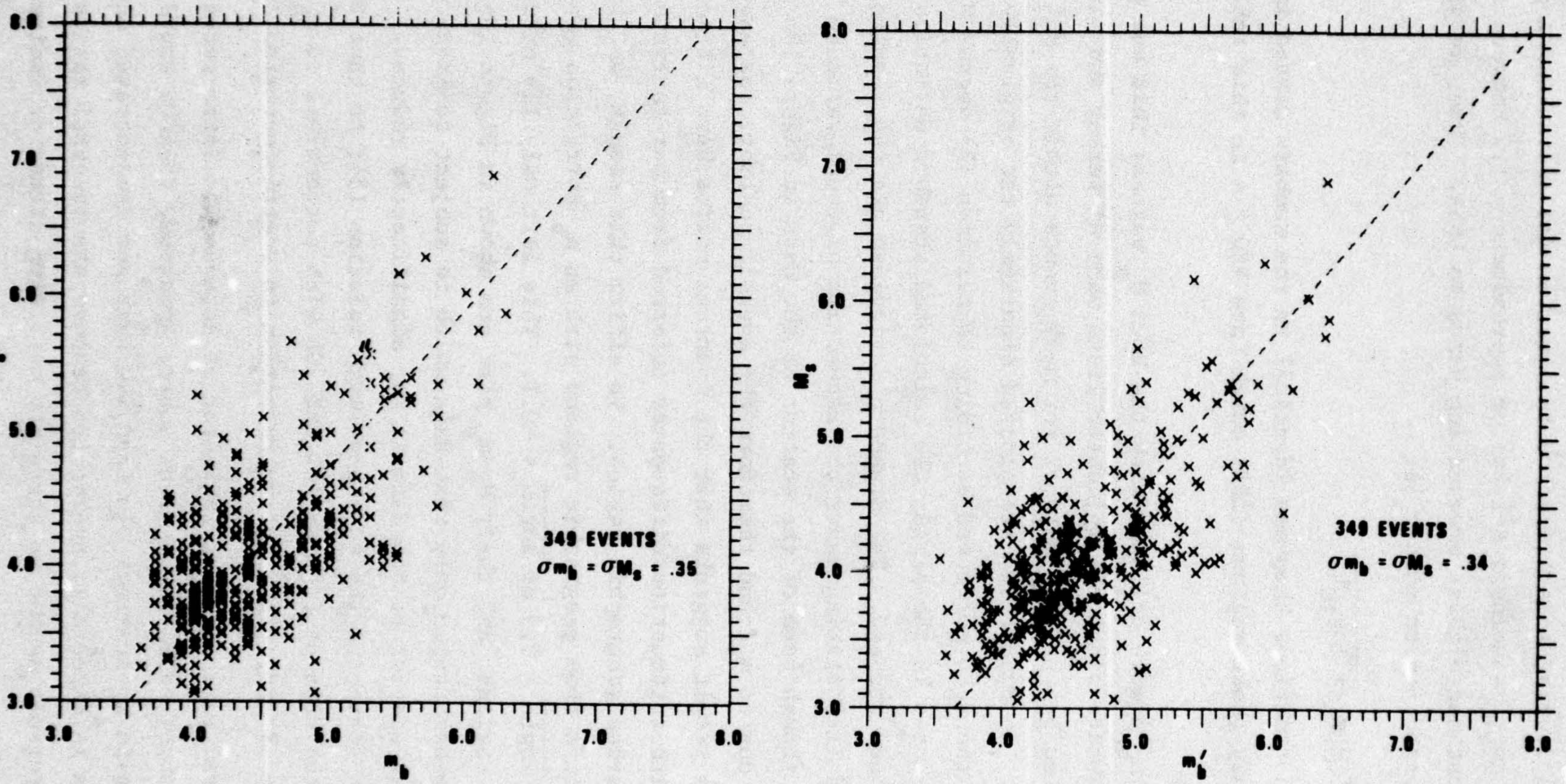

Figure 10a LASA m_b versus VLPE M_s before and after correcting according to average LL ratio for the geographic region of the event.

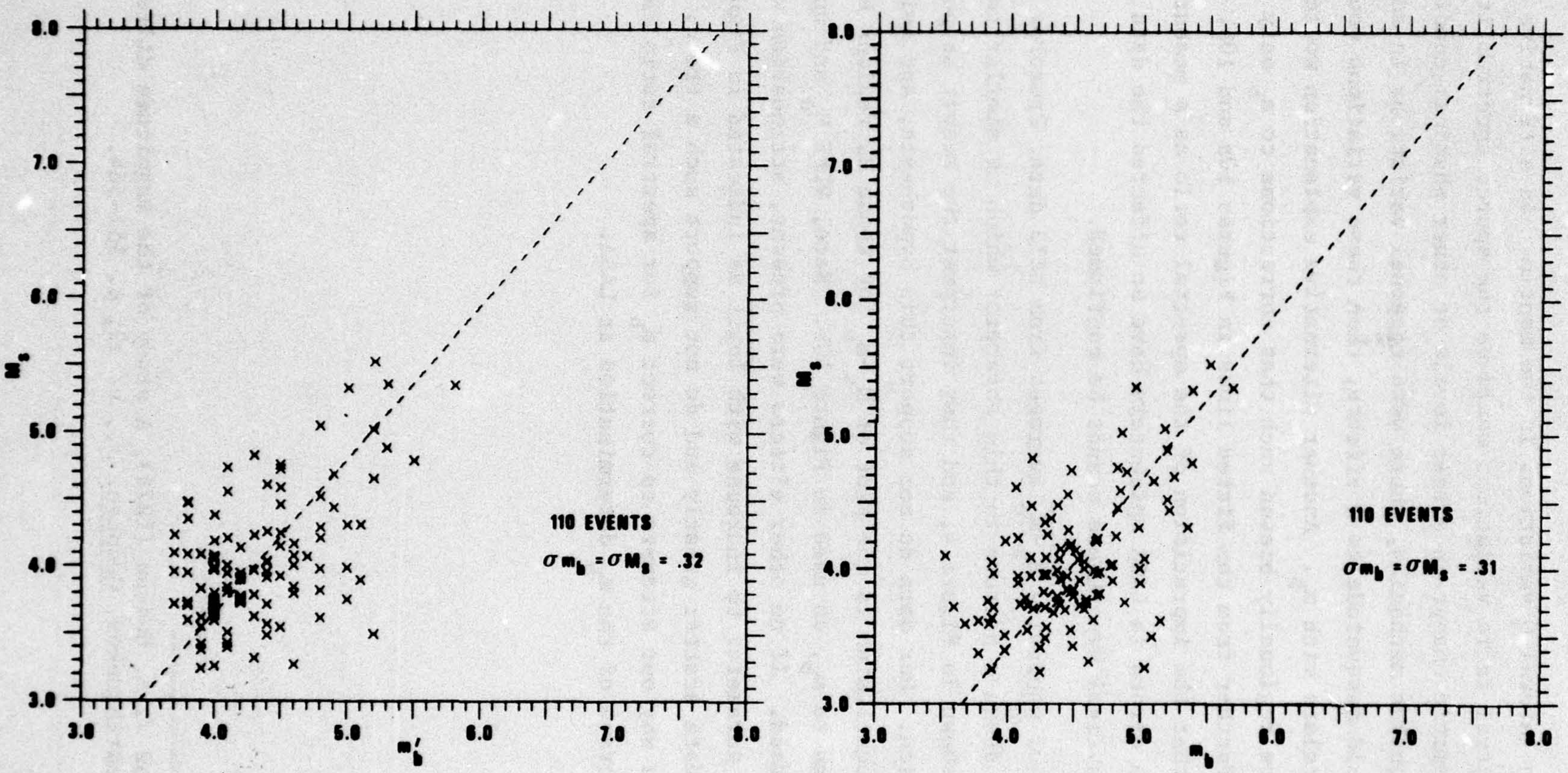

Figure 10b Subset of data from Figure 10a which includes only events in those geographic regions where average LL ratio was > 0.3 or < -0.1.

ical evidence on lateral Q variations in the mantle. An alternative explanation of the spectral ratio variations would be the source spectrum itself. If, due to systematic change in stress levels or other physical conditions affecting earthquakes mechanics, there were regional variations in source spectra that could masquerade as Q effects, then these variations would not necessarily correlate with m_b. Another alternative explanation would be that the M_s values are regionally biased such that corrections to m_b only serve to move the point farther from the fitted lines in Figures 10a and 10b. A third alternative is that the imprecision of the spectral ratio as a measure of attenuation or the errors in LASA epicenters have so affected the data that reasonable geophysical hypotheses cannot be confirmed.

In a study of wouldwide $\overline{M_s-m_b}$ averages from NEIS data, Prozorov and Hudson (1975) have shown a pattern in this parameter which is similar to our $\overline{\text{logLL}}$ averages shown in Figure 4, and they interpret the result as possibly due to attenuation. Our data do not support this hypothesis, and evidence against such a conclusion is the plot of M_s-m_b for those 21 regions having large corrections to m_b, as used in Figure 10b. Here, VLPE M_s and uncorrected LASA m_b are used. If no other effects were present, attenuation would cause the M_s-m_b parameter to increase with $\overline{\text{logLL}}$ as indicated in Figure 11; but the actual data scatter greatly and do not support such a trend. Figure 11 clearly shows why our attempts to correct m_b for spectral ratio failed to improve the accuracy of the m_b determination at LASA.

Prozorov, A., and J. A. Hudson (1974), A study of the magnitude difference M_s-m_b for earthquakes, Geophys. J., v. 39, p. 551-564.

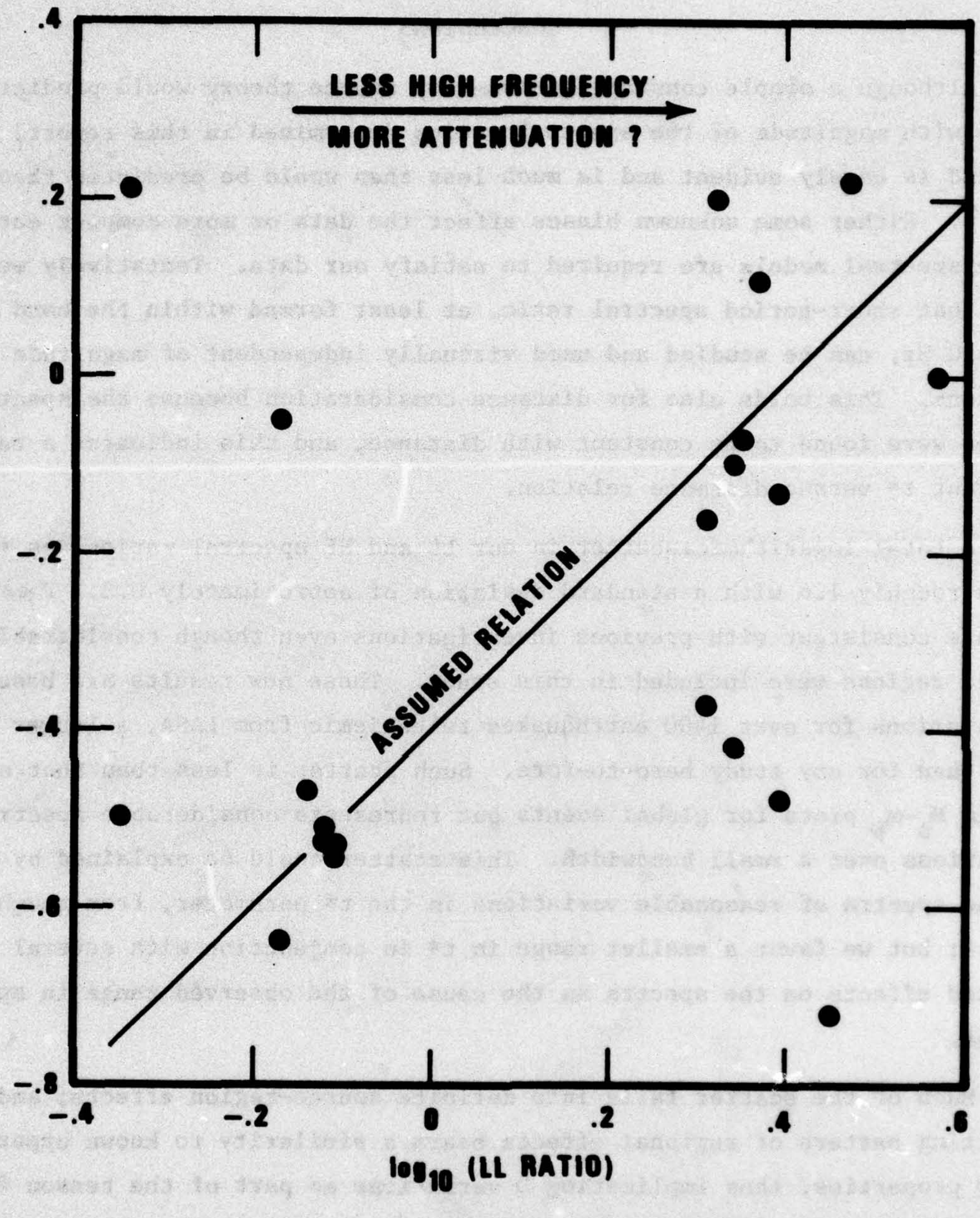

Figure 11 M_s-m_b for 21 geographic regions versus the average $\log_{10}$(LL ratio) for that region.

CONCLUSIONS

Although a simple constant stress-drop source theory would predict a trend with magnitude of the spectral ratios determined in this report, such a trend is barely evident and is much less than would be predicted theoretically. Either some unknown biases affect the data or more complex earthquake spectral models are required to satisfy our data. Tentatively we suggest that short-period spectral ratio, at least formed within the band of 0.4-4.0 Hz, can be studied and used virtually independent of magnitude considerations. This holds also for distance consideration because the spectral ratios were found to be constant with distance, and this indicates a rather constant t* versus distance relation.

A total logarithmic scatter in our LL and HF spectral ratios was found to be roughly 1.6 with a standard deviation of approximately 0.3. These values are consistent with previous investigations even though considerably more source regions were included in this study. These new results are based on computations for over 1400 earthquakes teleseismic from LASA, a larger data base than for any study here-to-fore. Such scatter is less than that observed in M_s-m_b plots for global events but represents considerable spectral variations over a small bandwidth. This scatter could be explained by effects on the spectra of reasonable variations in the t* parameter, from roughly 0.2 to 1.4; but we favor a smaller range in t* in conjunction with several source-related effects on the spectra as the cause of the observed range in spectral ratios.

Much of the scatter falls into definite source-region effects; and the resulting pattern of regional effects bears a similarity to known upper mantle Q properties, thus implicating Q variations as part of the reason for the observed scatter. But intraregional scatter of spectral ratios remains high, with a typical value of 0.2 for the standard deviation, and may be caused by variations in source character. This implies that even for low Q regions, the probability of observing an "explosion-like" spectral ratio is not negligible. For high Q regions, the probability may become distressingly high in a test-ban monitoring scheme which used spectral ratio as a discriminant. This extreme variability has not been so clearly pointed out in earlier work by other authors.

In this report we attempted for the first time to reduce scatter in M_s-m_b plots by correcting the LASA m_b according to average observed spectral ratio from the region of the epicenter. The attempt was unsuccessful. Our conclusion from this failure is that the LASA m_b's are not well correlated with spectral shape at LASA. We did not, however, design our experiment to study these problems directly and do not yet discount completely the gains to be made in discrimination by correcting m_b for attenuation differences, along with perhaps regional M_s corrections. These corrections will require detailed source-path analysis though.

The best check on spectral ratio as a discriminant involves comparing explosions with earthquakes in close enough proximity that the effect due to Q along the path is essentially unchanged. We have examined a small body of data which probably fits this criterion, namely epicenters in the southwest USSR, and found that the separation of source type on the basis of the LL spectral ratio is unsatisfactory. This result is consonant with that for earthquakes and explosions in the NTS area and further degrades the standing of P-wave spectral ratio as a reliable discriminant by itself. However, we have seen that the explosion ratios are among the lowest of the earthquake ratios; and on this basis the use of spectral ratio in a multidiscriminant procedure would be worthwhile if we could be certain that the discriminantion were not due to the explosions being still in a source region of higher Q.

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