T-2819

Q Determined from Local Earthquakes in the Charleston, South Carolina, Coastal Plain

by

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A thesis submitted to the Faculty and Board of Trustees of the Colorado School of Mines in partial fulfillment of the requirements for the degree of Master of Science Geophysics)

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ABSTRACT

Coda Q, or Qc, is determined from the rate of coda decay in the Coastal Plain near Charleston, South Carolina. The mean of Qc calculated from individual estimates is near 190 at 1 Hz, and increases to over 1650 at 10 Hz. The variation with frequency is similiar to that reported by several other researchers, for eastern United States attenuation parameters. A possible temporal change in Qc is identified, high values being associated with increased seismic activity during late 1977. Temporal changes have been observed elsewhere and laboratory measurements on Q under different stress states correlate high Q with high stress. A spatial variation in Qc is identified, increasing from west (Qc at 10 Hz equal to 850) to east (Qc at 10 Hz greater than 2100) across the coastal plain. The demarcation of high to highest zones of Qc is coincident with the western boundary of the innermost isoseismal of the Charleston 1886 earthquake. It is hypothesized that the 1886 earthquake was mislocated originally because of this spatial variation in energy attenuation that was unrecognized at that time.

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ACKNOWLEDGMENTS

This project was supported by the Nuclear Regulatory Commission, U. S. Geologic Survey, and Colorado School of Mines. Special thanks is given to the individuals in the Computer Center at the Golden office of the U. S. Geologic Survey who helped prepare the data and programs used in this research. Helpful discussions with Maurice Major, G. A. Bollinger, and Roger Stewart are gratefully acknowledged.

INTRODUCTION

Seismic hazard in an area is dependent on two factors, the nature of the earthquake source and the nature of the medium transmitting the seismic waves. There are two properties of the medium that are important, the velocity of wave propagation and attenuation of energy transmission. It is well known that velocity is dependent on wave type, location, and frequency. Attenuation is also dependent on these factors.

Earthquakes occur near Charleston, South Carolina, and are associated with an earthquake, of estimated body-wave magnitude 6.7, that occurred near Charleston in 1886. A permanent seismic network was installed in the 1970's by the U.S. Geologic Survey, in order to provide data for research aimed at understanding the causes of these earthquakes. In this study, we add to the knowledge on attenuation of earthquake waves, and methods of determining that factor, by determining Q for earthquake generated shear waves in the South Carolina Coastal Plain in the 2-10 Hertz frequency band. Data from nine earthquakes were suitable for use in this study. A map of the study area is shown in figure 1, and a table listing earthquake location parameters follows in table 1. Dates are indicated using

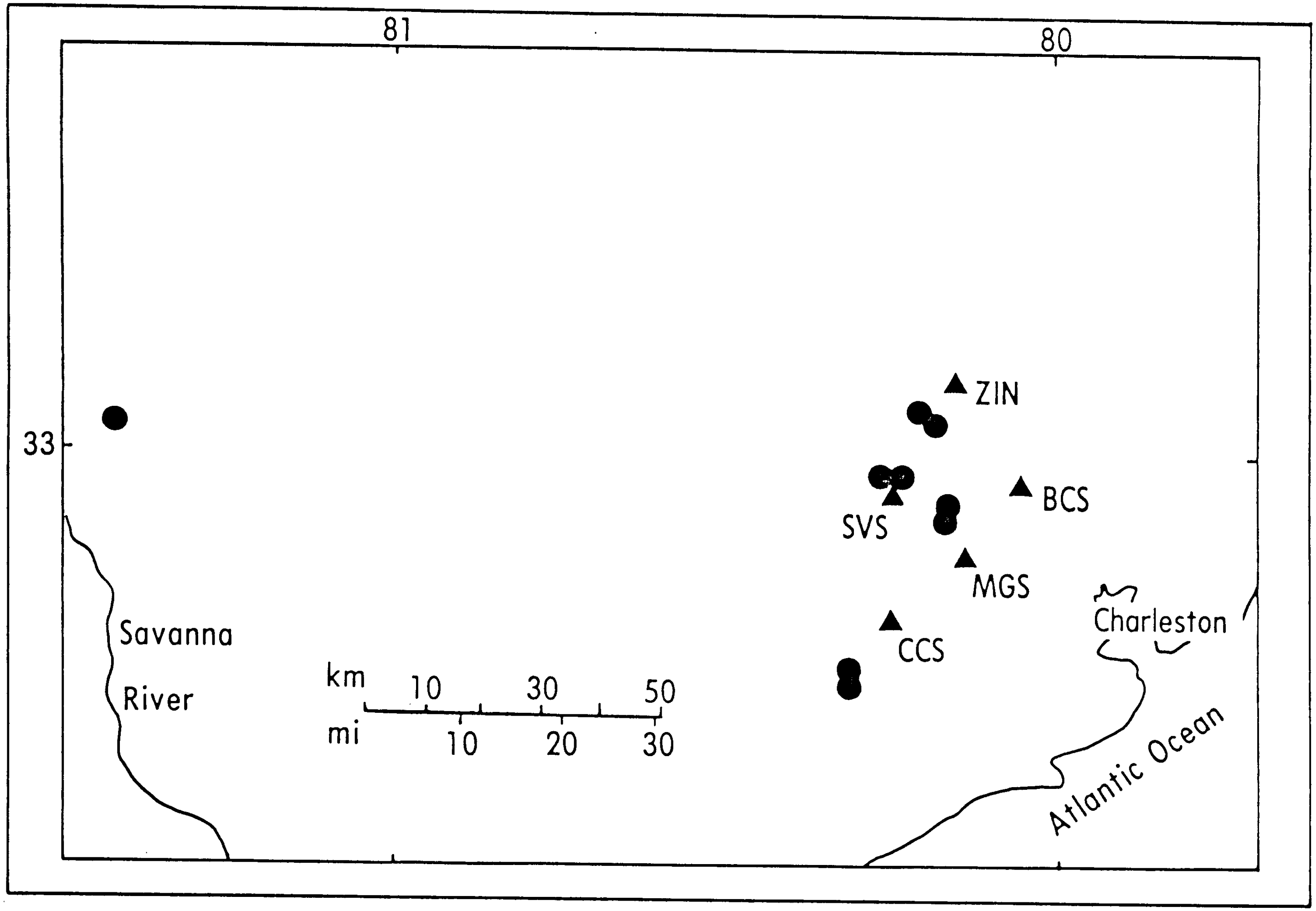

Figure 1. Area of study in southeastern South Carolina. Triangles represent seismograph stations, and circles earthquakes, used in this study.

Table 1

a. Earthquake Location Parameters

DATE	TIME(UTC)	LAT(N)	LON(W)	DEPTH	MAG
yrmndy	hrmn:sec	deg	deg	km	local
770120	0405:45.7	32.93	80.16	7.9	1.9
770605	0042:29.7	33.05	80.41	3.5	2.7
771215(a)	0715:55.1	32.98	80.27	14.6	2.1
771215(b)	1916:43.6	32.94	80.17	8.7	2.6
771216	1114:34.4	32.74	80.32	8.1	2.1
771216	1125:31.8	32.73	80.32	7.1	2.3
780907	2253:23.0	33.06	80.21	9.9	2.6
790127	2355:15.7	33.06	80.18	6.6	2.8
790811	0211:56.5	32.99	80.23	12.6	2.5

b. Epicentral distance (km)

DATE	BCS	CCS	MGS	SVS	ZIN
770120	10	15	4		
770605	125	111	120	109	
771215(a)	18	19	15	2	17
771215(b)	10		6	8	18
771216		13	26		
771216	38		27		
780907	16	28	19	11	7
790127	13	28	18	12	6
790811		19	13	3	

two digits each for year, month, and day, as 770605 for June 5, 1977.

A number of investigators have reported on similiar studies of Q. Sipkin and Jordon (1979) and Der et al (1981) determined Q for mantle shear waves, finding it strongly dependent on frequency. They showed almost constant Q at frequencies below 0.1 Hz, with exponentially increasing Q above 0.1 Hz. Dwyer (1981) showed Q increasing from 1300 at 1 Hz to 3000 at 10 Hz using local earthquakes in the New Madrid, Missouri, seismic zone. His determinations of Q were from distance-amplitude relationships. Pulli and Aki (1981) applied the coda Q method (Aki, 1969) to data from New England earthquakes, and found Qc increasing from 850 to 1500 in the 1.5 to 6 Hz range. Singh and Hermann (1983) used the coda Q method and found Qc increasing from 900 to 1800 in the 1 to 10 Hz range in the central United States, from 400 to 1000 in the northeastern United States, and from 300 to 1200 in the western United States. These results are similar to those obtained by Aki and Chouet (1975) for central California, by Mitchell (1981) for the northeastern United States, by Nuttli (pers. comm. in Singh and Hermann, 1983) for the central and western United States, and by Gupta (pers.

comm. in Singh and Hermann, 1983) for the eastern United States.

The purpose of this study is to use the previously developed technique for determining Q from the earthquake coda for finding Qc in the coastal plain near Charleston, South Carolina and to examine the reasonableness of the results in light of geologic constraints. The implications of this report will affect seismic hazard evaluation and seismic engineering in Charleston, and on coastal regions generally.

This paper is organized in four main parts. In the first, background geologic information and the tectonic setting is presented. Second, three possible models for the generation of the coda are discussed, along with the basis for selecting one of the models. Third is a section on data collection and processing, and the fourth part is the discussion of results. A brief summary of major conclusions ends the paper.

GEOLOGY AND TECTONIC ENVIRONMENT

The Coastal Plain of the Eastern United States is roughly 200 kilometers wide in South Carolina, and terminates to the west at the Fall Line, so named because of the topographic discontinuity that produces waterfalls on its whole length. Paleozoic and older rocks on the west are exposed in the Appalachian Range and Piedmont provinces. On the east side of the Fall Line, Mesozoic and younger sedimentary rocks thicken seaward over a crystalline basement. Borehole cores have defined 750 - 800 meters of Upper Cretaceous and Cenozoic unconsolidated sediments near Charleston (Gohn, 1983). Seismic refraction profiles have shown this layer to be 600 - 1100 meters thick across the eastern part of the Coastal Plain near Charleston (Ackermann, 1983). These cover 300 meters of lower Mesozoic basalts and an unknown thickness of lower Mesozoic (Jurassic) red beds (Gohn, et al, 1983). COCORP reflection profiles show a strong reflection at the top of the basalt, with a transparent zone down to 18 kilometers (Schilt, et al, 1983). Buried Mesozoic basins or grabens were identified from gravity and magnetic profiles in the Charleston area (Phillips, 1977; Long and Champion, Jr., 1977). The basalts and redbeds have been correlated with rifting of the Atlantic Ocean 200 million years ago with subsequent passive continental margin deposits (Gohn, et al., 1983). The crust-mantle boundary, identified both in refraction (Ackermann, 1977; Amick, 1979) and reflection profiles (Schilt, et al., 1983), occurs 30 - 34 kilometers deep.

In 1886, an earthquake of estimated body wave magnitude 6.7 (Bollinger, 1983) occurred 15 kilometers northwest of Charleston, causing extensive damamge in Charleston and the surrounding areas. It was felt over half of the United States. There was no surface fault evident, although bent railroad tracks, sandblows, and severe liquifaction damage were observed in the meizoseismal area.

Studies directed to seek a seismogenic source were initiated in the early 1970's. A permanent earthquake detection and location network was installed by 1974, with additional instrumentation installed in the next few years. Earthquakes have been recorded in the meizoseismal area, and are thought to be related to the 1886 shock. First motion data from current earthquakes indicates reverse motion on northwest striking faults, but northeast striking faults are also possible (Tarr and Rhea, 1983).

A northeast oriented maximum compressive stress regime is accepted by most workers in the area (Amick, 1983).

SELECTION OF THEORETICAL MODEL FOR GENERATION OF CODA

In this thesis, the object of analysis is that part of the earthquake signal following the direct P and S arrivals, labelled "coda" on figure 2. It is a quasisinusoidal arrival, and is generally considered to be caused by one or more of the following effects: one, an undesirable resonance in the instrumentation or in the coupling of the seismometer to its immediate environment; two, dispersion associated with a wave guide between the source and receiver; and three, scattering effects. It is believed that the final option is the most probable cause in this case. The following three sections expand on these three effects, explaining why the scattering model is preferred.

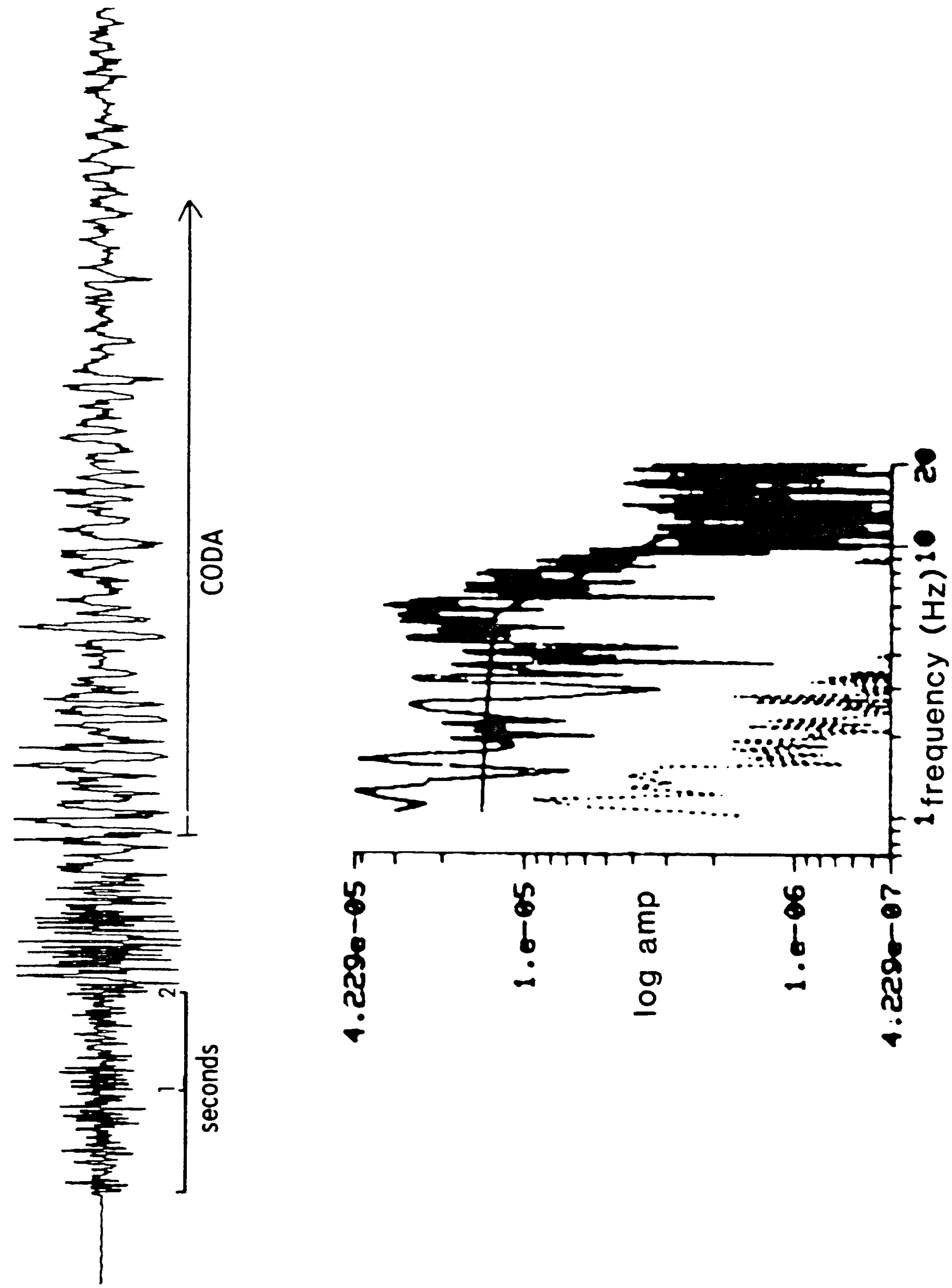

Figure 2. Seismogram from station MGS for earthquake on 771215a, showing time scale, partial coda, and spectral content of coda.

Ground-Seismometer Coupling Model

Resonant coupling of the seismometer with the ground could result in increased signal amplification at some frequency misrepresenting actual ground motion. Hoover and O'Brien (1980) addressed this issue, showing through several numerical and experimental examples that spurious signal amplifications do occur. The amount of amplification and the frequency of the peak response are dependent on geophone shape and mass, and soil type. They consider the "soil parameter" and the "plant parameter" separately.

The local soil type in the Charleston area Coastal Plain is saturated sandy clay. Hoover and O'Brien use tight sand for one example, a good approximation to the coastal plain sediments. The ratio of compressional wave velocity to shear wave velocity is 3.5 for the tight sand, corresponding to a Poisson's ratio of .45. The "soil parameter" is the ratio of shear to compressional wave velocity, or 0.286.

The "plant parameter", E, is dependent on mass of the geophone, its radius, and the surrounding soil's density. At the near surface, the wet clayey sand has a density approximated by 1.7 g/cc. The geophone is imbedded in a concrete slab, and the problem will be examined using the dimensions and mass of the concrete slab. We do so because the coupling factor being examined is between the ground and the geophone, and the concrete slab acts as the geophone in this context. The slab measures 60x90 cm, or approximately a 40 cm radius pad. Its mass is about 50 kilograms. The plant parameter, E, equals $(\text{mass}/\pi p r^3)$, or in this case $50000\text{g} / \pi(1.7\text{g/cc})(40\text{cm})^3$, or .15.

Hoover and O'Brien (1980) examine the frequency response of amplification in terms of dimensionless frequency p, equal to rw/a, where "r" is geophone radius, "w" is angular frequency, and "a" is compressional wave velocity in the adjacent soil. Using 500 mps for a, 40 cm for r, and 20π for angular frequency (corresponding to 10 Hz), frequency p is .05.

Comparing these values for soil parameter, plant parameter, and frequency p with Hoover and O'Brien's (1980) results (figure 3), there appears to be no need for concern about spurious signal amplification in this study. As can be seen in figure 3, the peak in signal amplification decreases as the plant parameter decreases. Their lower limit for plant parameter (E=1) is six times the plant parameter (E=.15) in this study. We can expect the shape of the amplitude response for our case to be approximately

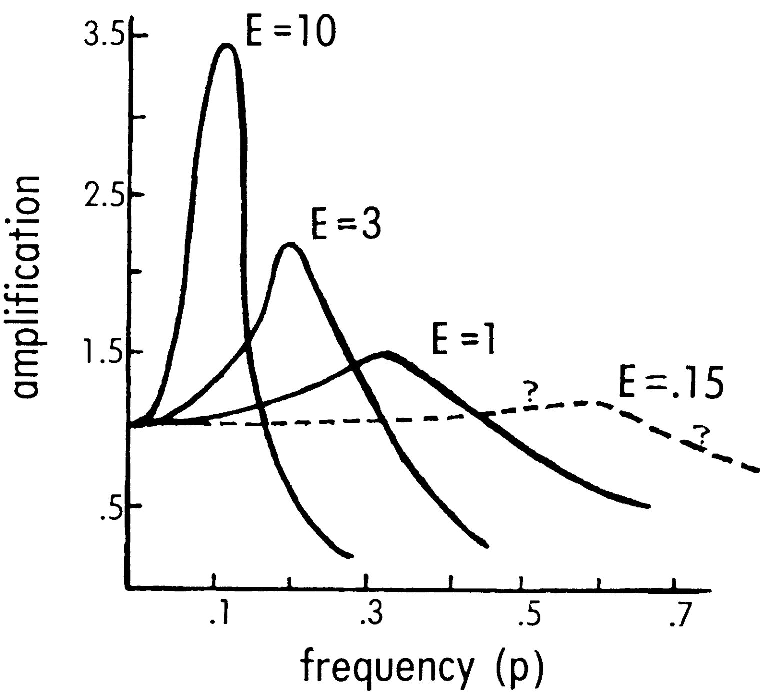

Figure 3. Calculated resonance for geophone planted in a tight sand, from Hoover and O'Brien (1980). In this study, frequency p is .06 and plant parameter E is .15. See text for explanation of terms.

as shown by the the dotted line in figure 3. Also, signal amplification does not begin to occur in their example (for E = 1) until frequency p is .1 or greater. Amplification reaches 1.5 times at frequency p of .3, six times the frequency of interest in this study.

Unrealistic physical parameters for geophone radius and mass, and soil characteristics, would have to be assumed in order to predict serious signal amplification caused by coupling between the geophone and ground. This cause for the coda or for the observed increase in Qc at high frequencies does not appear to be valid.

As will be discussed later, different values of Qc were determined for earthquakes close to and distant to the stations. If ground-seismometer coupling dominated the signal, observing spatial variation in Qc would be unexpected. Spatial variation is suggested by the data, and we are further supported that coupling resonances are not influencing our results.

Wave Guide-Surface Layer Model

Normal local earthquake wave analysis assumes a simple flat earth model, velocity increasing gradually and smoothly with depth. Energy dissipation models are then spherical spreading ones. If a low velocity surface layer is included, these assumed energy dissipation models could be incorrect. Simple geometrical spreading could be altered by wave entrapment. In the Coastal Plain, there is such a layer, 600-1100 meters thick, consisting of Cenozoic unconsolidated sediments (see Geologic and Tectonic Environment section) that could be acting as a wave guide. If the wave guide is significant, the amplitude of some modes of the waves in the coda would be maintained by constructive interference and entrapment, and produce apparent high Qc values.

The evidence that such a surficial wave guide exists comes from corehole data and examination of three component earthquake records taken from the study area. Corehole data (Gohn, 1983), with corroboration from refraction data (Amick, 1979), indicate this suspect layer is 750 meters thick, and has a compressional wave velocity of 2200mps. The material beneath it has a compressional velocity of 6000 mps.

Examination of a three component earthquake record shows some interesting features. As shown in figure 4, the P-wave is easily identified on the vertical motion record at top, and correlates well with first motions on the

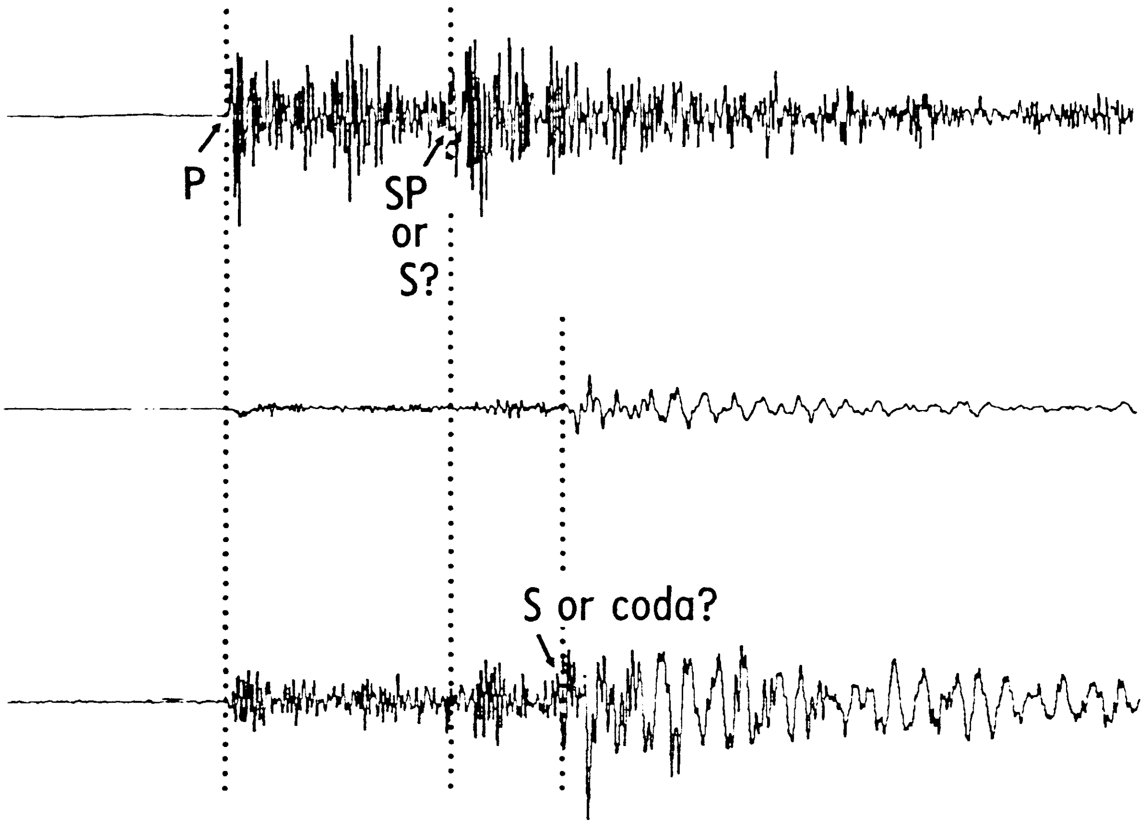

Figure 4. Seismograms from earthquake recorded on three component, Geotech L-7, system, from station MGS for earthquake on 790811. Vertical motion record on top, and two horizontal motion records below. Phase arrivals indicated.

horizontal motion records below. The next largest arrival on all records appears to be an S arrival, but there is a third arrival on the horizontal motion records not yet explained. The time difference between the S-wave arrival and the next arrival on the horizontal motion record is .8 seconds. There is consistently this amount of delay (between .7 and 1.0 seconds observed for nine earthquakes in the study area) between these two arrivals.

It is possible to hypothesize a travel path in which an S-wave hits the bottom of the 750 meter thick surface layer and a large part of its energy is converted to compressional waves. This compressional wave, and the original shear wave, travel to the surface. The difference in their travel times is dependent on the difference in velocities of shear and compressional waves in the surface layer. Using the measured travel time difference (figure 3) of .8 seconds, compressional velocity of 2200 mps and layer thickness of 750 meters, the shear wave velocity is 660 mps. This corresponds to a Vp/Vs ratio of 3.35 and a Poisson ratio of .45 (similiar to the tight sand from Hoover and O'Brien's (1980) study). Such material is very different from the material below this layer which has an assumed normal Vp/Vs ratio of 1.73 and Poisson ratio of

.25. (No evidence contradicting this hypothesis is available.) We can conclude that the surface layer has very different properties than the material beneath it, and there is a strong impedance contrast at the boundary between the two.

Low velocity surface layers with high Poisson's and Vp/Vs ratios are common throughout the world. In borehole measurements in coastal regions of the United States and Japan (Nicholson, 1983), Vp/Vs ratios as high as .45 were determined for near surface material. Also, evidence supporting a high Poission ratio in the Charleston area is that a large amount of damage from the 1886 earthquake was caused by liquifaction.

Having established that a low speed surface layer exists, it is appropriate to inquire into whether or not waves guided in it contribute largely to the coda that is observed. For example, are the velocity - frequency relations predicted for the wave guide about those that are characteristic of the real seismograms? For this discussion, we examine only the frequency of Rayleigh modes that would be associated with the wave guide. (Love modes are not useful because all the data used in this study is from vertical component seismometers, which do not detect motion from Love waves.) We are concerned mainly with low order Rayleigh modes, because experience suggests higher modes (>>4th) are rarely seen in earthquake seismology.

We have computed the phase and group velocity curves characteristic of certain Rayleigh modes of the model determined from the earlier discussion on S to P conversion. This model and its group velocity curves are shown in figure 5. The fundamental Rayleigh mode was not located in the computer search. A shear mode, going to cutoff (phase velocity equal to 3.47 k/s) at a period beyond four seconds is the lowest mode mapped. This may be the first shear mode (or the first higher Rayleigh mode in some nomenclatures) but we are not certain. Its identity is not crucial to the argument, however. The important point is that, in the frequency range of interest for this study (3 to 10 Hz), we have to appeal to the n + fifth and higher shear modes of the Rayleigh wave to appear and dominate the signal. Only rarely do such high modes dominate the seismic signal, and we consider it remarkable if they do so in this case. The higher modes are also sharply trapped and should be poorly excited by sources outside the wave guide. Here, earthquakes occur at depths of 6 to 15 km, 6 to 20 times deeper than the surface layer. Thus, one has

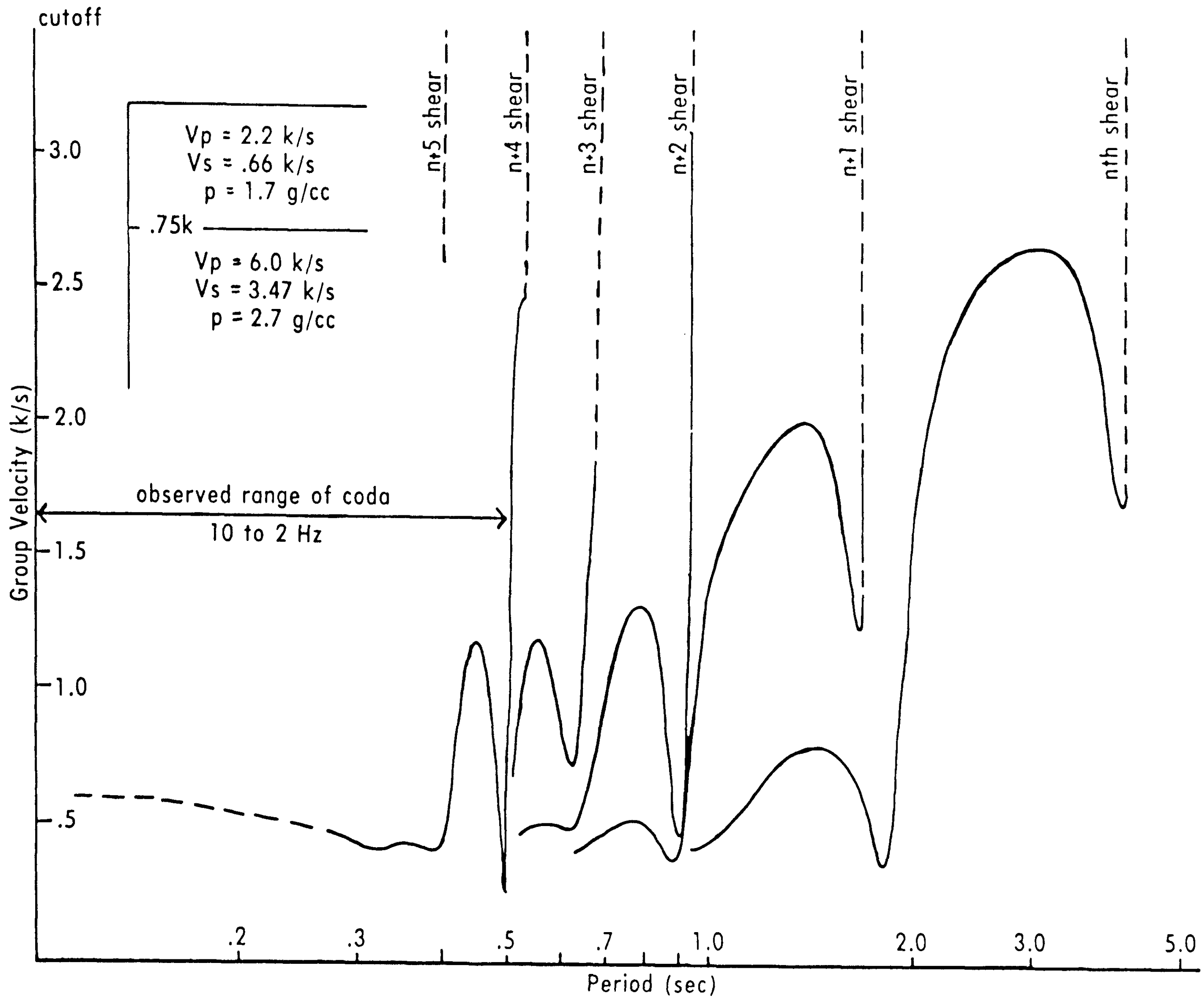

Figure 5. Group velocity curves of Rayleigh waves in low velocity surface layer over half space model, shown in upper left inset. Vp is compressional wave velocity, Vs is shear wave velocity, and p is density.

to imagine that the higher shear modes are vigorously excited by sources far below the wave guide, while they also are vigorously trapped within the wave guide.

We conclude that accepting the surface layer-wave guide model as the dominant source of waves in the earthquake coda involves assuming remarkable conditions. We think these conditions make this model, at the least, no more attractive than the next model to be presented.

Backscattering Model

In this model, the medium is assumed to be a homogeneous matrix with randomly distributed heterogeneous inclusions acting as sources for backscattering. Backscattered waves are modelled as simple, single scattered waves. The coda consists of the superposition of independently scattered waves, with later arrivals coming from more distant scatterers. Distances between sources and receivers are small compared to the distance to the farthest scatterers, such that the duration of primary waves is small compared to the duration of the coda. The earthquake source is spatially small compared to the volume covered by the scattered waves, which can be defined by shear velocity times coda duration. Geometric spreading of the backscattered wave is described by the inverse square law, corresponding to spreading for body waves.

The waveforms and phase arrivals shown in figure 4 can be explained via this backscattering model. The P and S wave arrivals identified on the vertical and horizontal record would be correctly identified. The later arrival on the horizontal records would correspond to the later arriving backscattered coda waves. The backscattered waves have to travel to a scatterer and then to the receiver before being recorded. There would be a travel time delay between the direct and backscattered shear waves. This time delay may be what is observed in figure 4 between the questioned S arrivals.

We consider the backscattered model best suited for explaining the coda's generation and use it to determine Q from the coda's amplitude delay. Aki and Chouet assert (Aki and Chouet, 1975, equation 27, pg 3329) that the average amplitude in a segment of the coda Δt in length, centered at time t, viewed through a bandpass filter of center frequency w, is of the form:

$$A(w|t) = c\ t^{-a} \exp(-wt/2Q),$$

where

$A(w|t)$ is the average amplitude of the filtered signal over a time window centered at t;

t is the center of the time window, referenced to time after the earthquake origin;

c is the coda source factor at frequency w, is independent of time, and radiation from the source has no azimuthal dependence;

a is a constant depending on geometric spreading here taken to be 1;

Q is the "quality factor" of interest in this paper, herein denoted Qc for "coda Q".

Equating natural logarithms of both sides gives

$$\ln A(w|t) = \ln c - \ln t - wt/2Qc.$$

The term ln c is a function of the source, here taken to be independent of time and radiation patterns, and is therefore not a function of factors influencing energy loss in the medium. It can, therefore, be treated as a constant, K, in this problem. Further simplification of this equation gives

$$\ln A(w|t)\ t = K - (w/2Qc)\ t.$$

Thus ln At is a linear function of t with slope equal to (-w/2Qc). The source term K determines the vertical position of the line, but does not effect the slope of the line from which Qc is determined.

The value for Qc determined in this manner is a combination of scattering losses and anelastic losses. These effects cannot be separated because of the similarity in their mathematical form. Aki (1980a,b) and Dainty (1981) have addressed this issue, and in their opinion, at low frequencies scattering losses dominate, while at higher frequencies anelastic losses dominate.

DATA COLLECTION AND PROCESSING

Raw data used in this study was from small earthquakes which occurred in the Coastal Plain of South Carolina. Approximately thirty earthquakes, ranging in magnitude from 1.5 to 2.8, occurred in the years from 1977 to 1980. Mark Products L-4C seismometers, with free period equal to 1 Hz, responding to vertical velocity, were used in the network recording these earthquakes. Seismometers were critically damped (.8), and responded to a maximum of 2.8 volts/cm/sec (7 volts/in/sec). Maximum mass displacement corresponded to four volts. Magnification varied by more than 18 dB, from 13,000 to 110,000 at 3 Hz. There was constant magnification above 3Hz, 3 db down at 1 Hz.

Data was transmitted via unity gain amplifiers and telephone carrier to a recording site at a local college. Data was recorded on several analog paper printers (helicorder, heat sensitive paper and pen) and also on analog magnetic tape. Frequency range of the analog tape was restricted because of tape speed. The gain was constant to 30 Hz, 3 db down at 33.6 Hz, and rolled off quickly above that. Analog tapes with earthquakes of interest were sent to U.S. Geological Survey offices in Golden for processing. Tapes were digitized at 130 Hz, to allow data examination out to 30 Hz without worry of signal aliasing. These digitized traces were then demultiplexed and displayed on a Tektronix 4014 computer terminal screen.

Characteristic phase arrivals were picked (P and S waves), and signals examined for clipping. This occurred when translated ground motion exceeded four volts up or down. Many of the data channels were clipped, and could not be used for further analysis. The earthquake source zones and stations used in this study are shown in figure 1. A sample seismogram, with the coda indicated, and coda spectral response displayed, is shown in figure 2.

Of the remaining signals, spectral analysis was done to examine frequency content and noise levels. Signals contaminated by high noise were removed at this stage. Amplitude-spectra of the remaining data were then examined, out to 15 Hz. The normal spectral shape appeared as quasi-constant amplitude at lower frequencies with a monotonic amplitude decrease above some corner frequency. Corner frequencies were in the 5 to 10 Hz range. Beyond 15 Hz, spectral levels did not continue to decay as expected, but remained high or increased in some cases. For this study, we decided not to examine the higher frequency (greater than 15 Hz) data because it was not well behaved.

Station signals were input to a computer program written by Robert Hermann (Hermann, 1977), at St. Louis University. The unfiltered signals were examined for dominant frequency content and amplitude in succesive 3.85 second time windows referenced to earthquake origin time. This permitted creating amplitude (A) - time (t) pairs from which plots could be generated according to ln(At) versus t. This result for the unfiltered signal is shown in the top diagram on figure 6a. A logarithmic least squares determined line was fit to the data points. The slope of

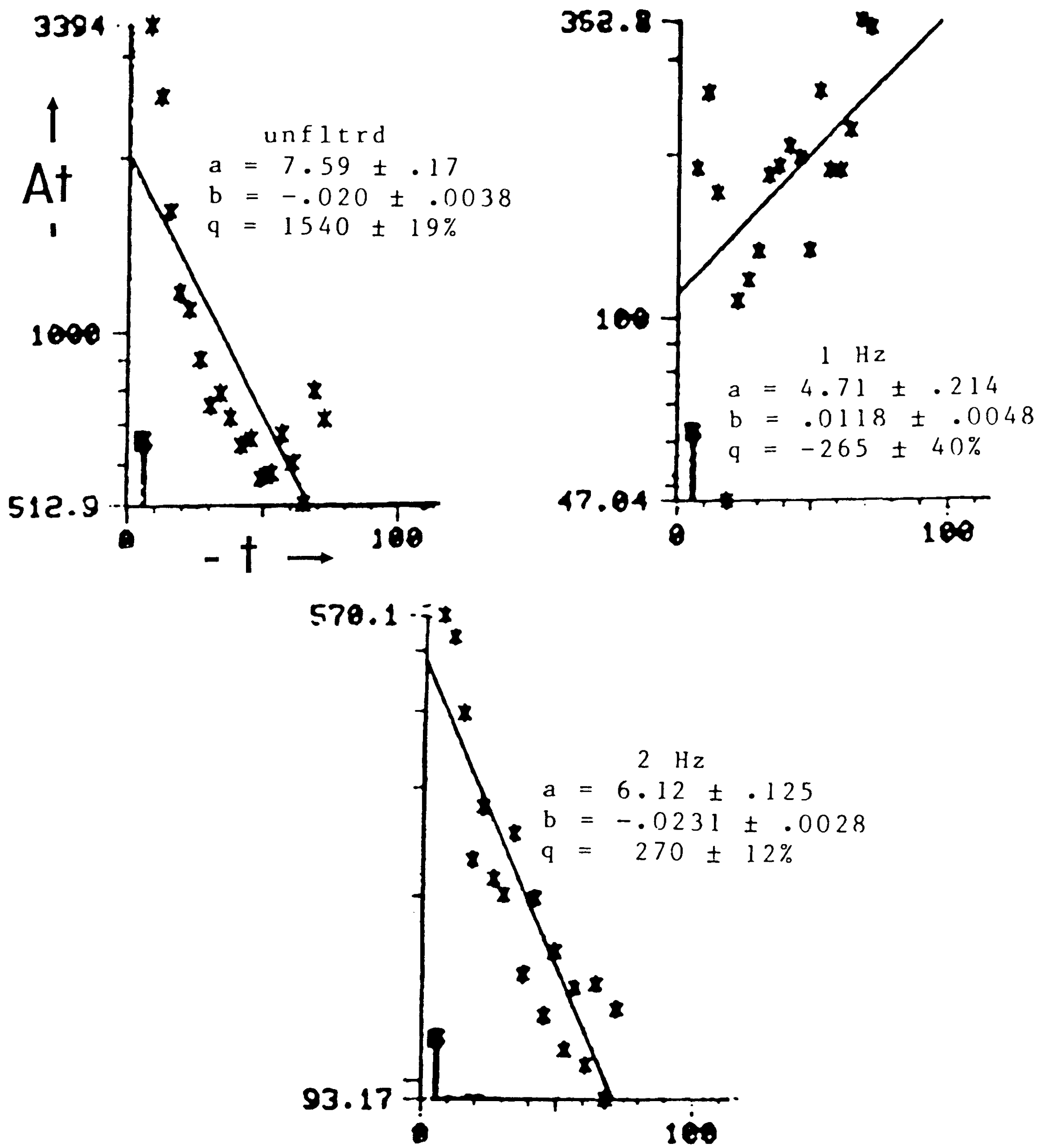

Figure 6. Computer output for Qc determination from record in figure 2. On the vertical axis, amplitude times time is plotted on a logarithmic scale. On the horizontal axis time in seconds is plotted on a linear scale. a) Unfiltered, 1, and 2 Hz filtered trace analyses.

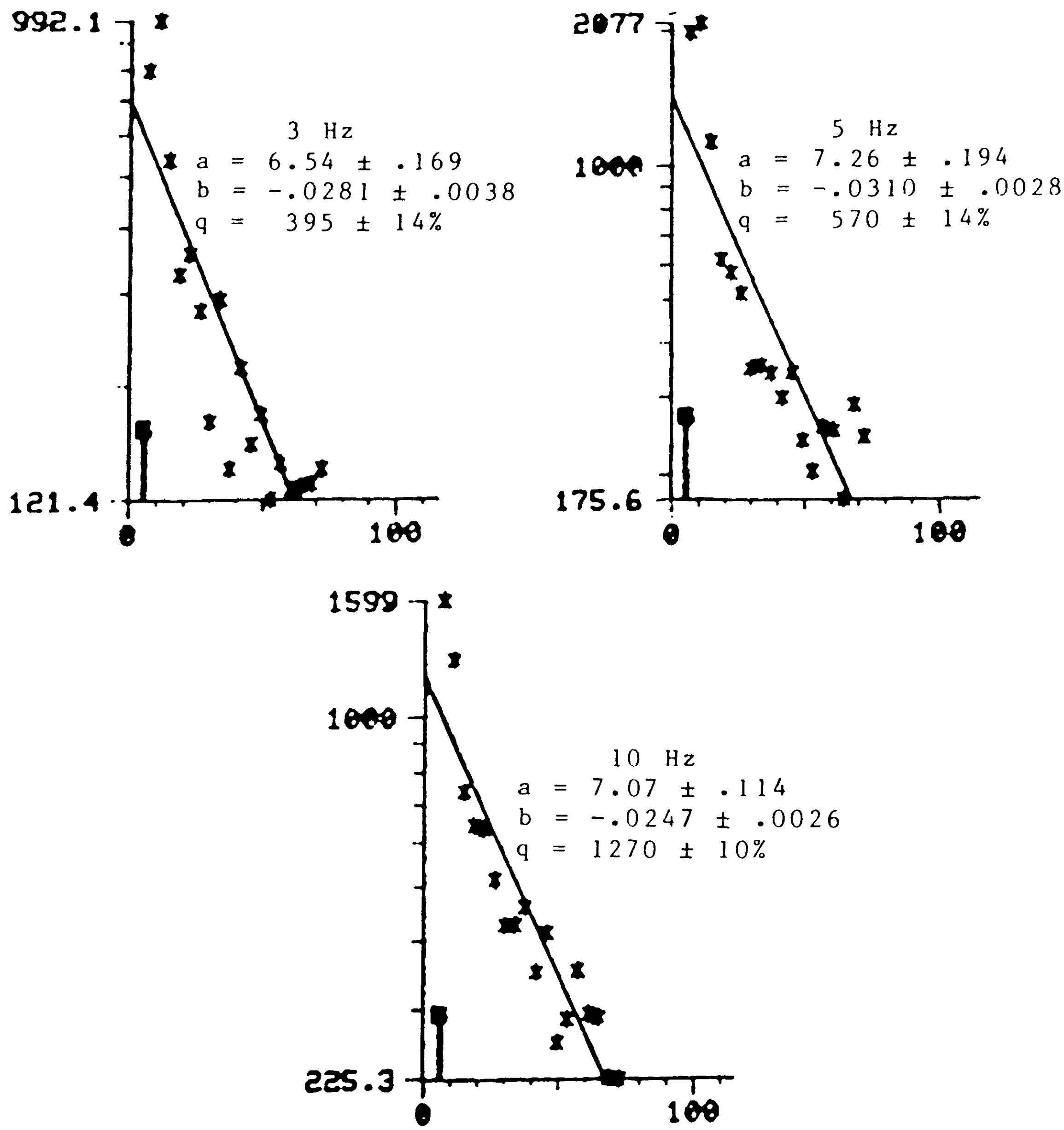

Figure 6. Computer output for Qc determination from record in figure 2. On the vertical axis, amplitude times time is plotted on a logarithmic scale. On the horizontal axis time in seconds is plotted on a linear scale. b) 3, 5, and 10 Hz filtered trace analyses.

the line and dominant frequency were used to algebraically determine Qc from slope = $-\pi f/Qc$ (method described previously in this paper). A table, listing coefficients of the equation $\ln(At) = a - bt$, where $a = \ln(Ao)$ and $b = \pi f/Qc$, error estimates on those coefficients, the calculated value of Qc, and its error estimate, was printed adjacent to the plot of data.

Error estimates were computed in the following way. The standard deviation of a and b was computed from the square root of the variance of each of those values. The relative error in q was determined from the ratio of the standard deviation of b to b, and is expressed as a percentage.

The process described above on the unfiltered signal was applied to filtered signals after narrow bandpass filters (six pole Butterworth filter, low pass at $2^{-1/2} f_c$, and high pass at $2^{1/2} f_c$) at 1, 2, 3, 5, and 10 Hz had been applied. The center frequency (f_c) of each filter was used in the slope-Qc equation mentioned earlier. The computed amplitude-time plots, with respective numerical result tables, were printed on a single page. These could then be examined for each station's record of each earthquake (figure 6a and b). If the relative error was less than

30% (00% corresponding to a perfect fit), the value determined for Qc was considered acceptable. If the value of Qc was negative, that result was not used. Such a situation could arise when the signal was severely noise contaminated at a particular frequency so that A remained essentially constant while time increased. This occurred most often at 1 Hz, but also occurred at other frequencies.

At this stage, there was data from nine earthquakes remaining. Four earthquakes had data from only two stations, although two of these earthquakes occurred within ten minutes of each other, within 1.5 kilometers of each other, and were at approximately the same depth and magnitude. Data from these two earthquakes was combined, making only two earthquakes with data from two stations. This makes four earthquakes with data from three stations, one earthquake with data from four stations, and one earthquake with data from five stations. The computer estimates for Qc each show five significant figures, but since we surmised the data was not that well controlled, we reported values of Qc to only three significant figures. Table 2 lists the individual values determined for Qc and the error coefficient for each value.

Table 2

Station-Frequency Coda Q determinations

DATE	STA	FILTER FREQUENCY and %ERROR				
yrmndy		2 ±%	3 ±%	5 ±%	10 ±%	unfltr ±%
770120	BCS				2000 ±22%	2060 ±23%
	CCS		400 ±15%	780 ±29%	2360 ±10%	1300 ±09%
	MGS		280 ±22%	390 ±14%	1940 ±20%	1370 ±14%
770605	BCS	190 ±18%	260 ±22%	350 ±17%	770 ±12%	850 ±08%
	CCS	170 ±19%		510 ±20%	800 ±06%	840 ±08%
	MGS	120 ±18%	170 ±17%	280 ±22%	610 ±12%	740 ±16%
	SVS	170 ±29%	300 ±24%	350 ±18%	820 ±16%	870 ±16%
771215a	BCS	260 ±20%	490 ±22%	1040 ±18%	2340 ±20%	
	CCS	540 ±29%	540 ±26%	630 ±16%	1440 ±12%	1940 ±11%
	MGS	270 ±12%	330 ±14%	570 ±14%	1270 ±10%	1540 ±19%
	SVS			930 ±18%	1870 ±11%	2860 ±21%
	ZIN	250 ±13%	360 ±16%	660 ±15%	2560 ±27%	1880 ±20%
771215b	BCS				2940 ±14%	3160 ±13%
	MGS		850 ±29%	1160 ±12%	2410 ±15%	2160 ±14%
	SVS	600 ±15%	800 ±17%	1160 ±14%	2210 ±13%	1910 ±08%
	ZIN	790 ±24%	1210 ±23%	1630 ±21%	2920 ±11%	2680 ±11%

Table 2

Station-Frequency Coda Q determinations

DATE	STA	FILTER FREQUENCY and %ERROR				
yrmndy		2 ±%	3 ±%	5 ±%	10 ±%	unfltr ±%
771216_1	CCS			970 ±27%	2080 ±14%	1940 ±12%
	MGS		860 ±23%	790 ±13%	1820 ±15%	1680 ±11%
771216_2	BCS				2790 ±17%	3290 ±21%
	MGS	700 ±22%	640 ±17%	780 ±11%	1840 ±18%	1860 ±15%
780907	MGS	920 ±17%	1140 ±14%	1220 ±06%		2290 ±14%
	ZIN	530 ±15%	840 ±12%	1350 ±09%	2290 ±15%	2550 ±07%
		(unfltrd BCS:3070±10%, CCS:1570±05%, SVS:2940±14%)				
790127	BCS				2140 ±14%	2110 ±18%
	CCS		450 ±19%	960 ±21%	1620 ±09%	1390 ±07%
	MGS			1220 ±24%	2700 ±24%	2690 ±17%
	ZIN	310 ±18%	350 ±16%	570 ±11%	1660 ±10%	1390 ±09%
		(unfltrd SVS:1720±10%)				
790811	CCS	120 ±20%	120 ±22%	200 ±12%	590 ±18%	610 ±05%
	MGS			390 ±20%	770 ±08%	930 ±14%
	SVS		510 ±26%	580 ±23%	1240 ±28%	1230 ±17%

RESULTS

Summary of Numerical Values

Figures 7 to 14 display the values determined for Qc from this study. There is one figure per earthquake. Solid circles are values determined from narrow band frequencies centered on 2, 3, 5, and 10 Hz. Solid squares are values from the unfiltered signal. The unfiltered signal was dominated by 10 to 20 Hz frequencies, and plotted at 10 Hz. A straight line, corresponding to $Qc=Qo\ (f/fo)^{n}$ where fo is 1 Hz, has been fit, using logarithmic least squares, to the values on each plot. The precedent for frequency dependent Q has been established, as discussed earlier, and appears to occur in this data set. At higher frequencies, greater than .1 Hz, there should be stronger frequency dependence (Sipkin, 1983, pers. comm.). Perhaps a curved line on the log-log plot, flattening toward lower frequencies, would be a better description of Qc. There are two good reasons not to do this. One, the equation for the curved line would be difficult to find. Two, results have already been published for the rest of the United States using the straight line equation, and can easily be compared to these results.

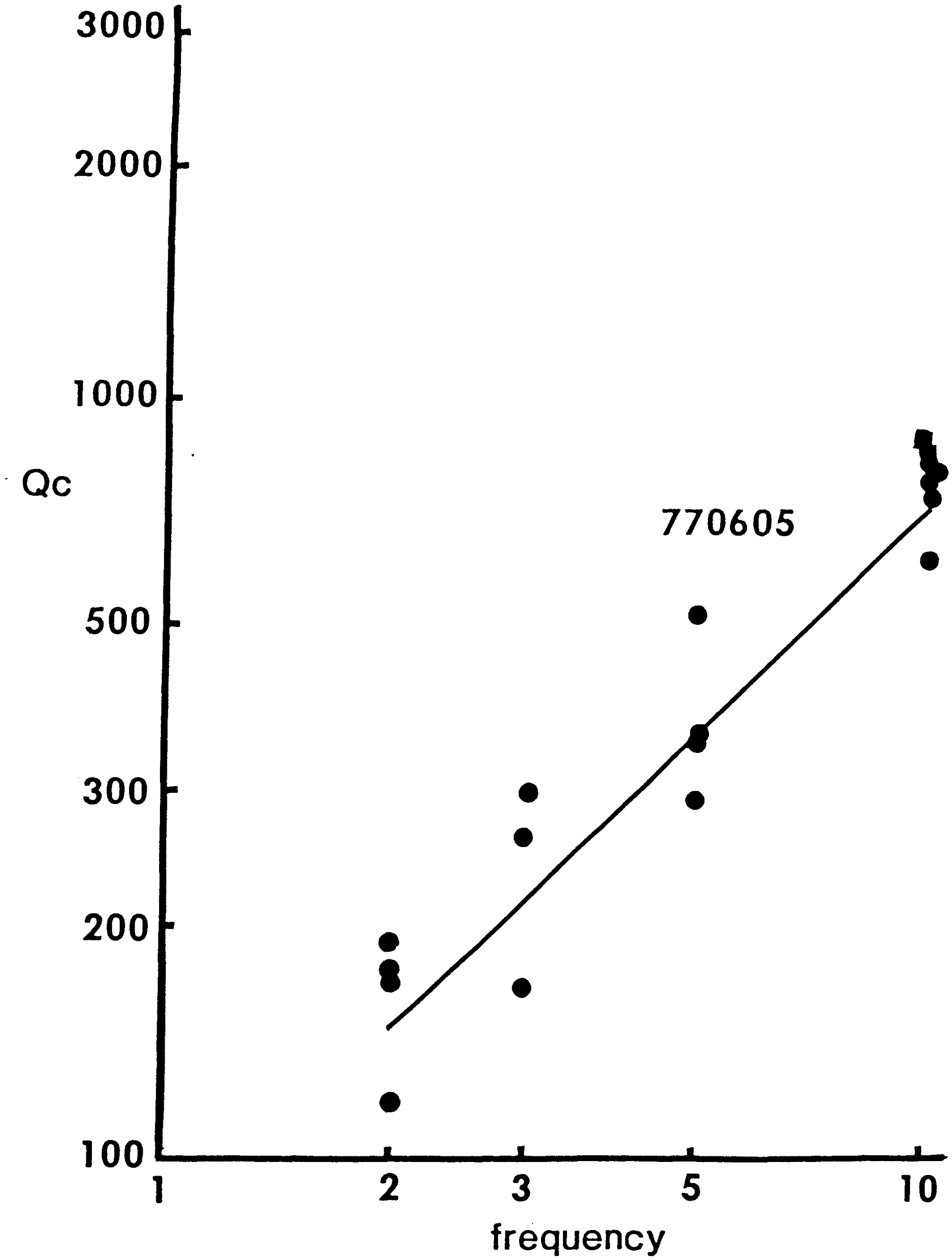

Figure 7. Plot of Qc versus frequency for earthquake on 770605. Solid circles are from narrow bandpass filtered data, solid squares are from unfiltered data. Line is the logarithmic least squares fit.

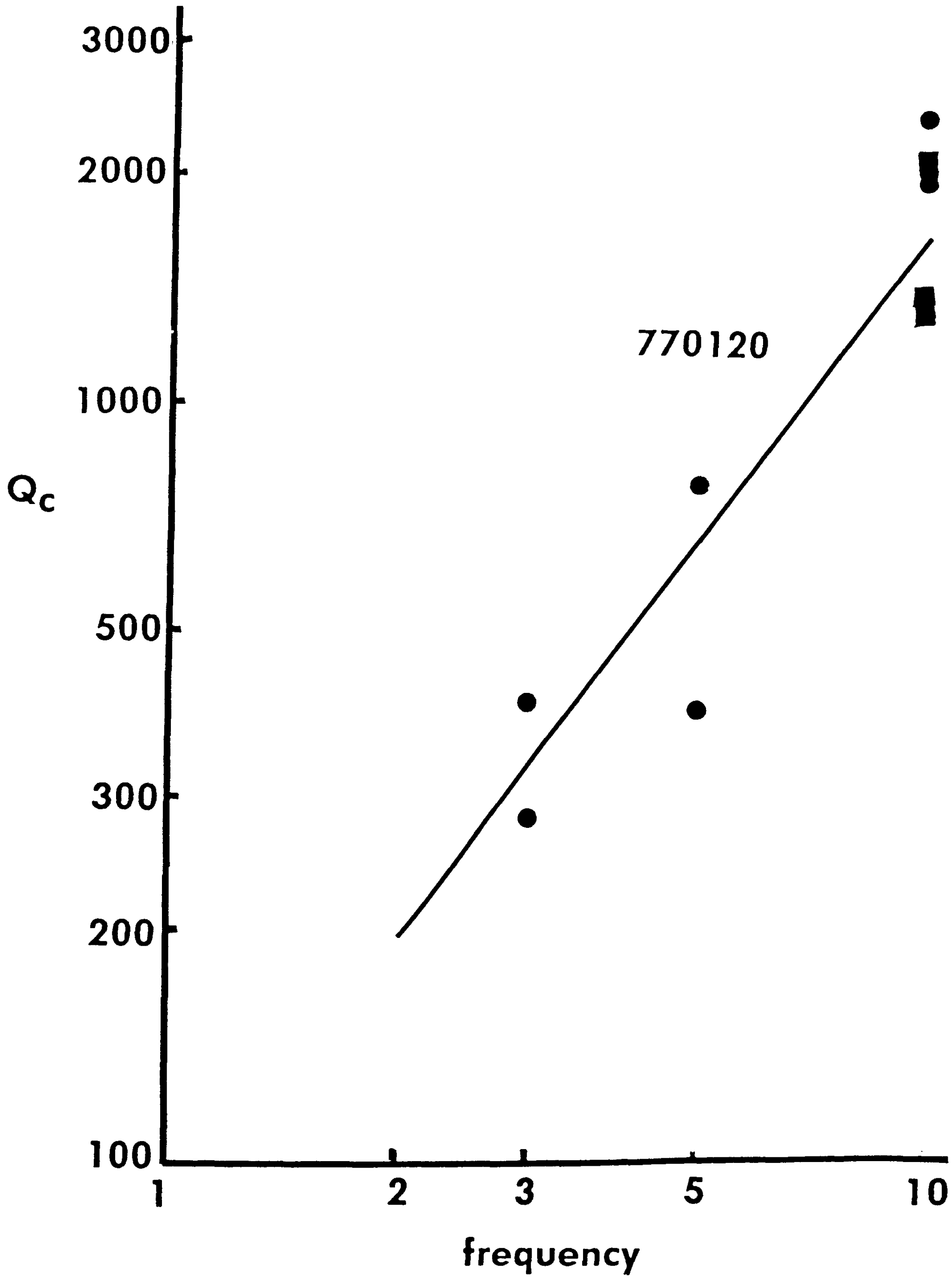

Figure 8. Result of Qc determinations for earthquake on 770120. Same symbols as figure 7.

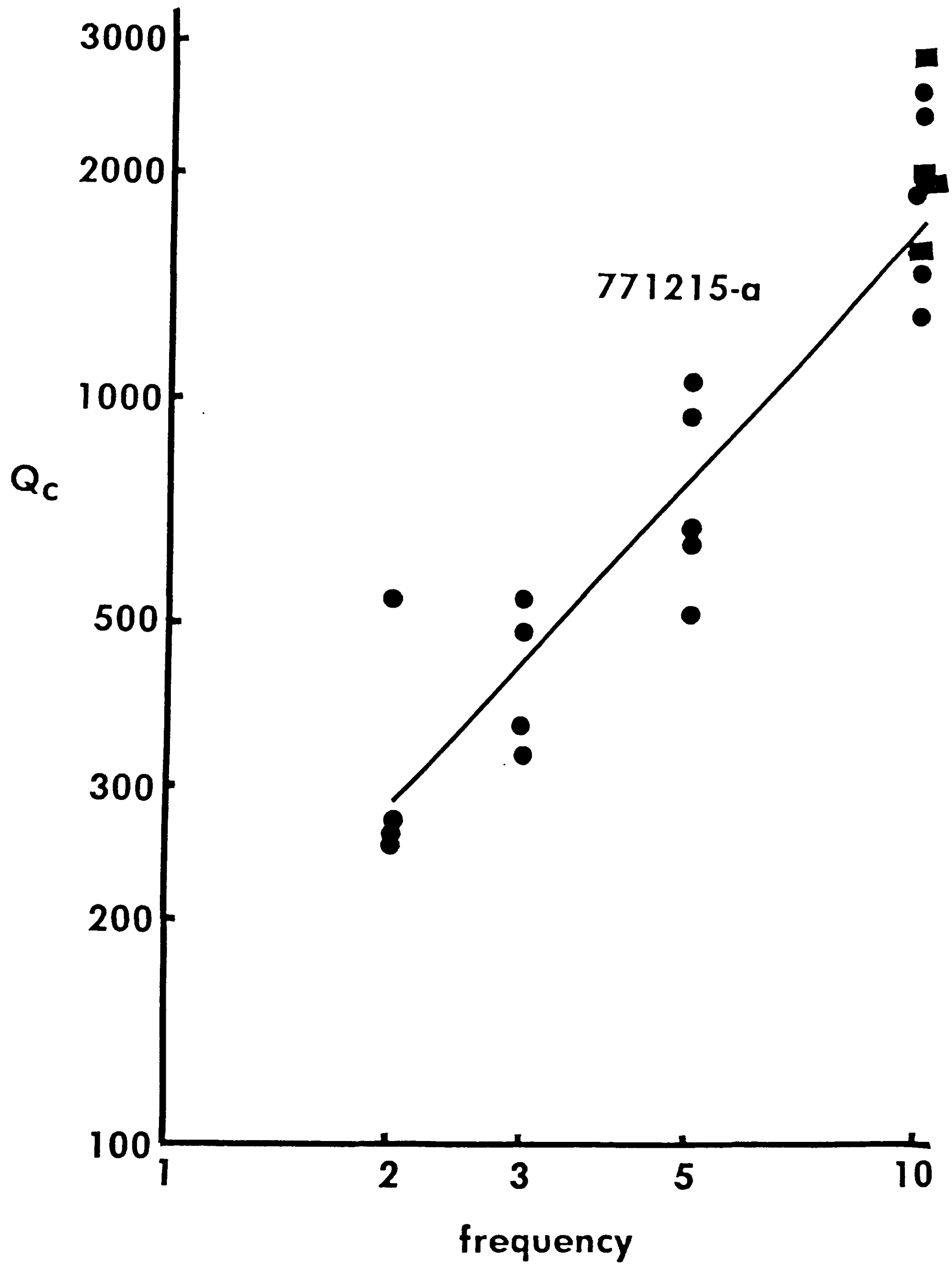

Figure 9. Result of Qc determinations for earthquake on 771215 at 0715 Greenwich Time. Same symbols as figure 7.

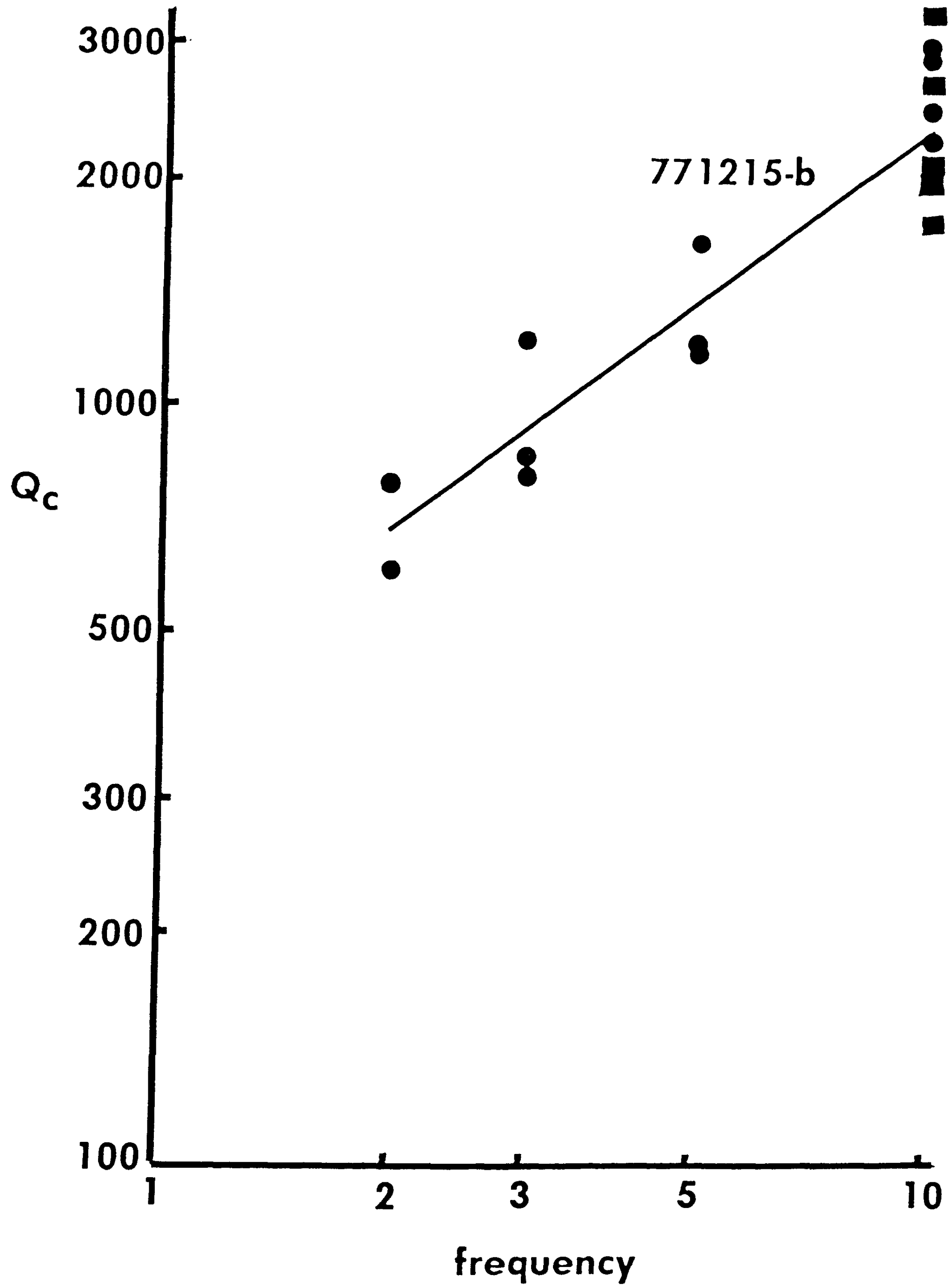

Figure 10. Result of Qc determinations for earthquake on 771215 at 1916 Greenwich Time. Same symbols as figure 7.

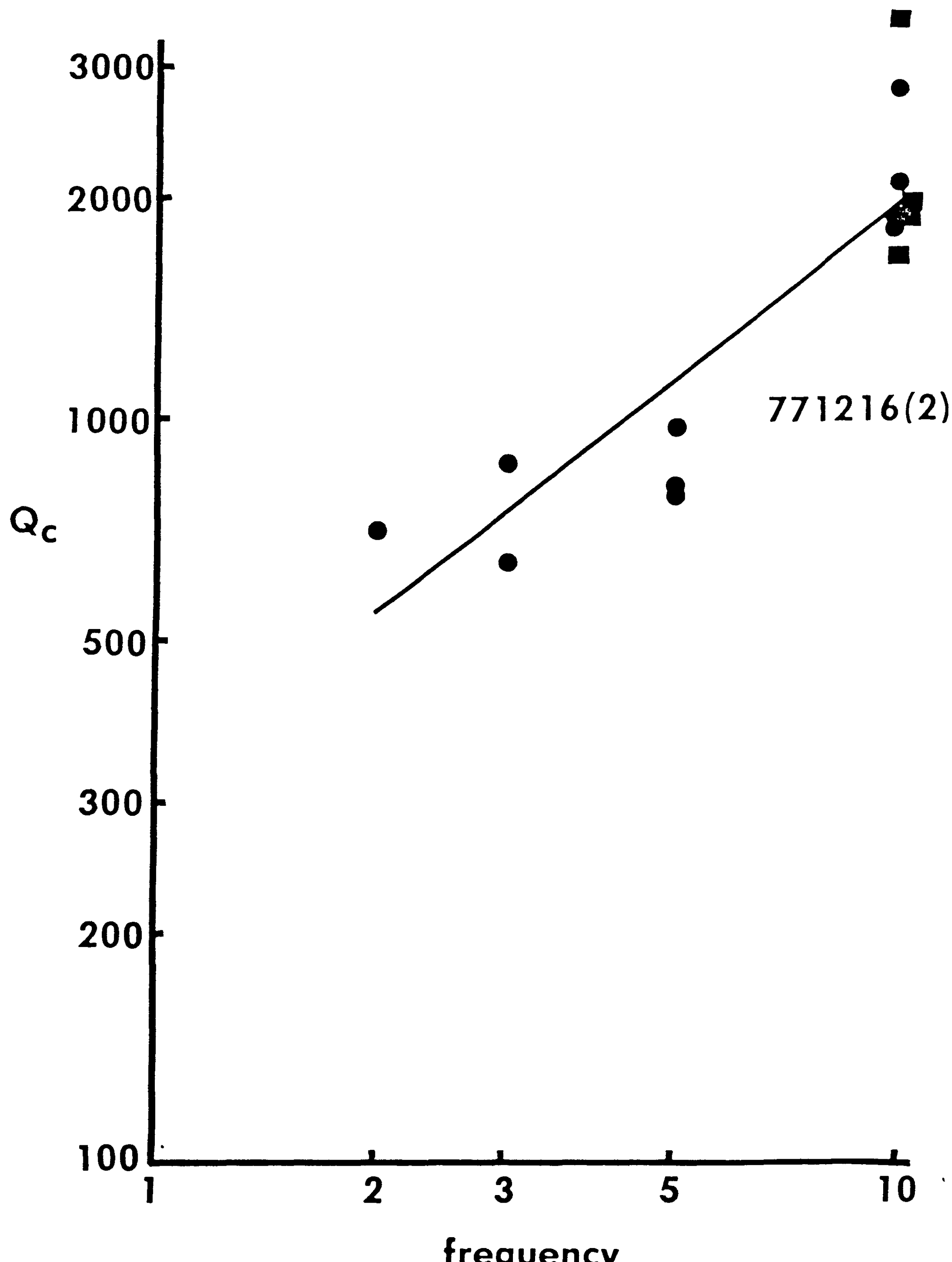

Figure 11. Result of Qc determinations for two earthquakes at 1114 and 1125 Greenwich Time on 771216. Same symbols as figure 7.

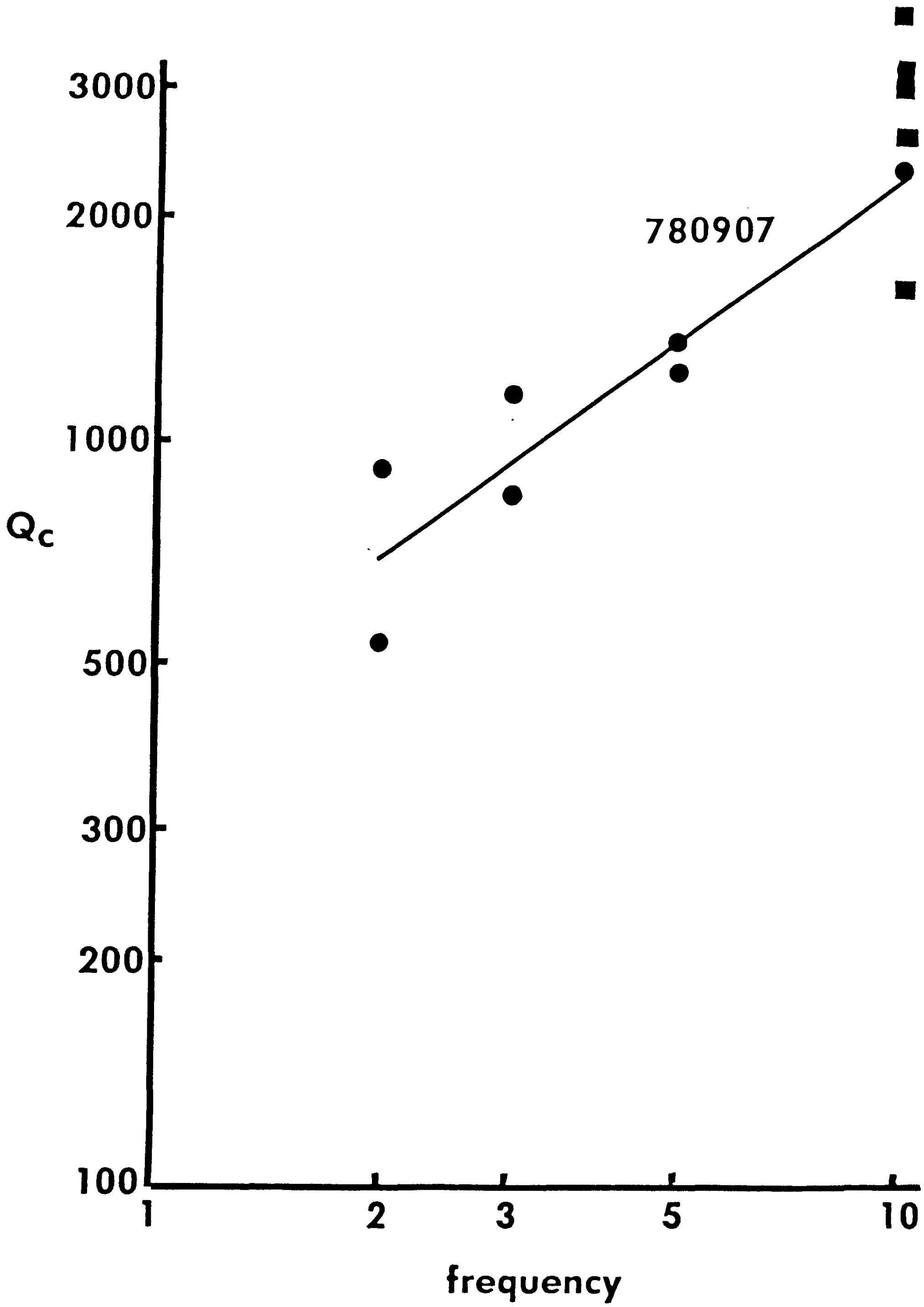

Figure 12. Result of Qc determinations for earthquake on 780907. Same symbols as figure 7.

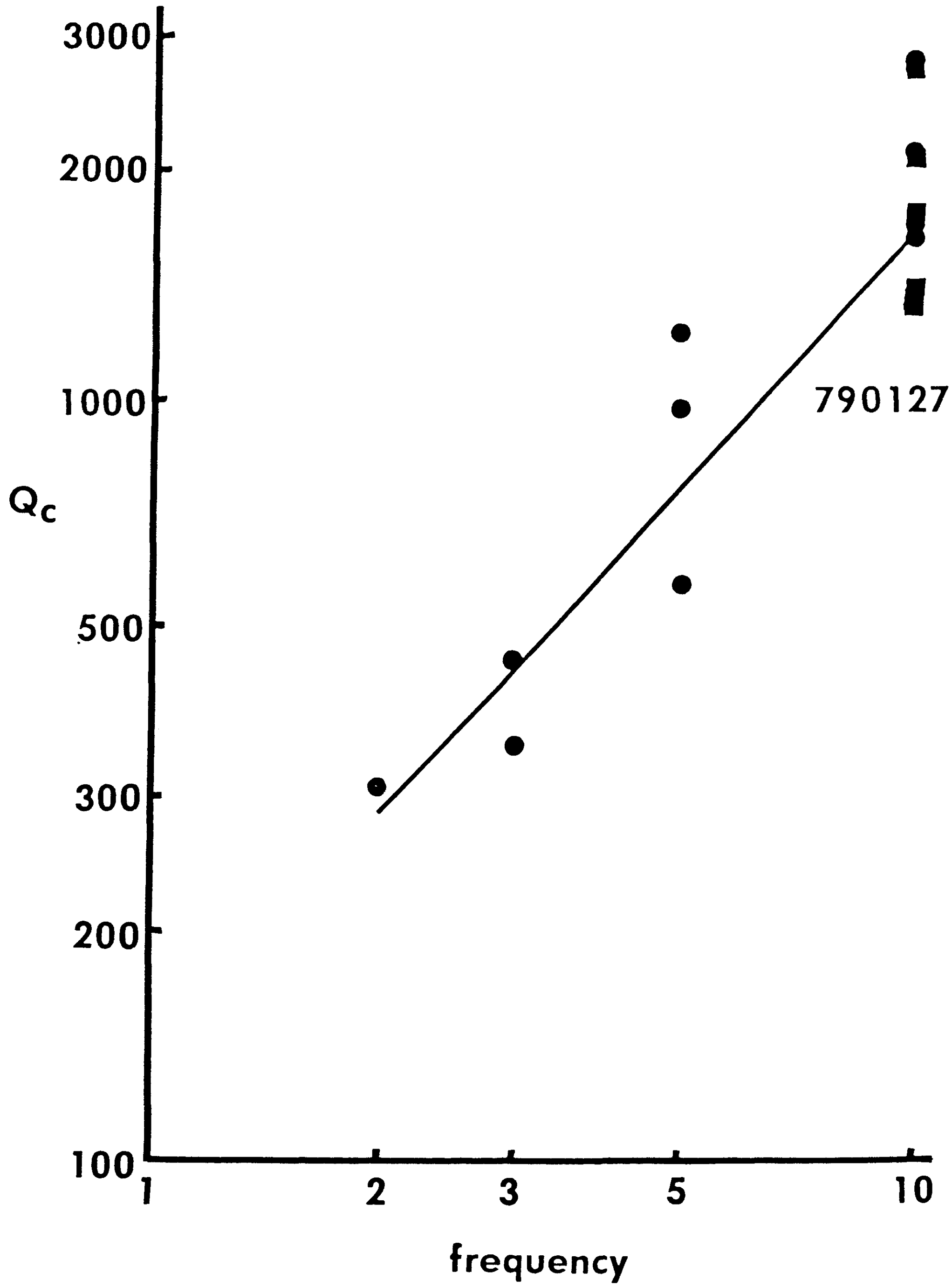

Figure 13. Result of Qc determinations for earthquake on 790127. Same symbols as figure 7.

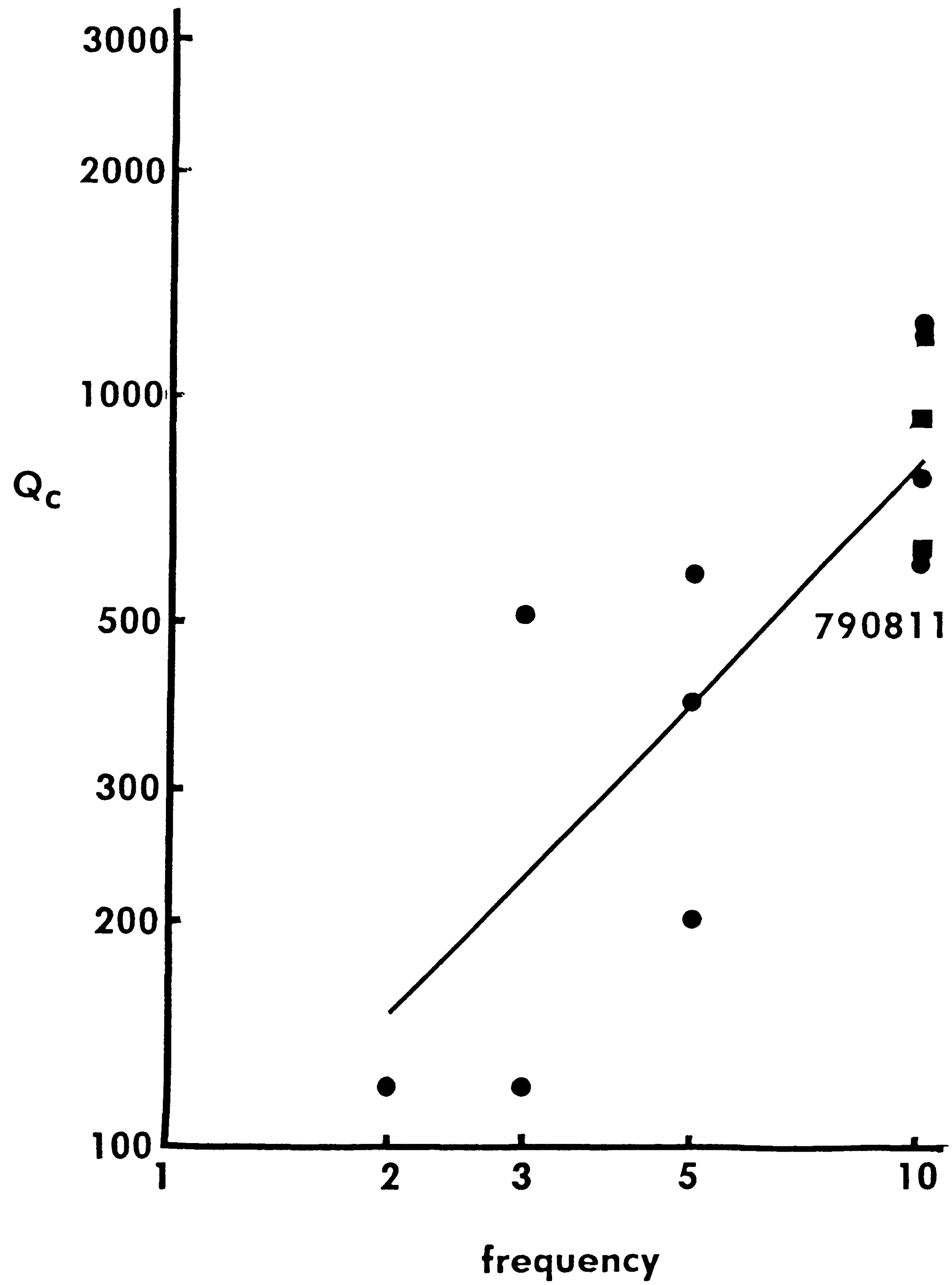

Figure 14. Result of Qc determinations for earthquake on 790811. Same symbols as figure 7.

It is not expected that the Qc determinations be significantly different for each earthquake studied, but we plotted the results so that we could examine this tenet. In the development of this theory, there was the assumption that the source function does not affect the attenuation function, and there should be no differences one earthquake to another.

Results for an earthquake on 770605 are shown in figure 7. This earthquake was in the southwestern part of the state (figure 1), over 100 kilometers (60 miles) from the receiving stations. Its path intercepts more and deeper material than paths for the other earthquakes studies. The Qc values fit, very well, the line described by $Qc=80f^{93}$. The largest deviation from this line is at 5 Hz, where data points are plus/minus 100 from this line.

For the 770120 earthquake (figure 8), $Qc=80f^{1.29}$. There is limited data available for this earthquake, none below 3 Hz. The remaining data points fit the line quite well, however.

For 771215a (figure 9), there are many data points, and only a couple more than 200 away from the line described by $Qc=135f^{1.10}$. The position of these data points is different from those for 771215b (figure 10).

This earthquake occurred twelve hours after the first one, was just over a kilometer away, was six kilometers shallower, and was .6 magnitude larger. There is good fit of the data to the line $Qc=400f^{.75}$.

The data for 771216 (figure 11) are not positioned distinctly different from those for 771215b, but are higher at all frequencies than 771215a. There is fairly large scatter of data about the line $Qc=310f^{.79}$, up to 200 away from the line for most data points.

Data for 780907 (figure 12) are indistinguishable from data for the last two earthquakes. The line fit to this data corresponds to $Qc=420f^{.73}$. There is reasonably good fit to this line.

The position of data for 790127 (figure 13) is much lower than the previous examples. The scatter is not large, and the line corresponds to $Qc=140f^{1.07}$. The position of data for 790811 (figure 14) is lower still. Although there is a high degree of scatter, the fitted line is described by $Qc=70f^{1.06}$, and is the lowest value for Qc.

The preceeding results are summarized in figure 15 and table 3. The heavy line through the center is the fit to all the data, assuming no systematic variation within it, and corresponds to $Qc=190f^{.94}$.

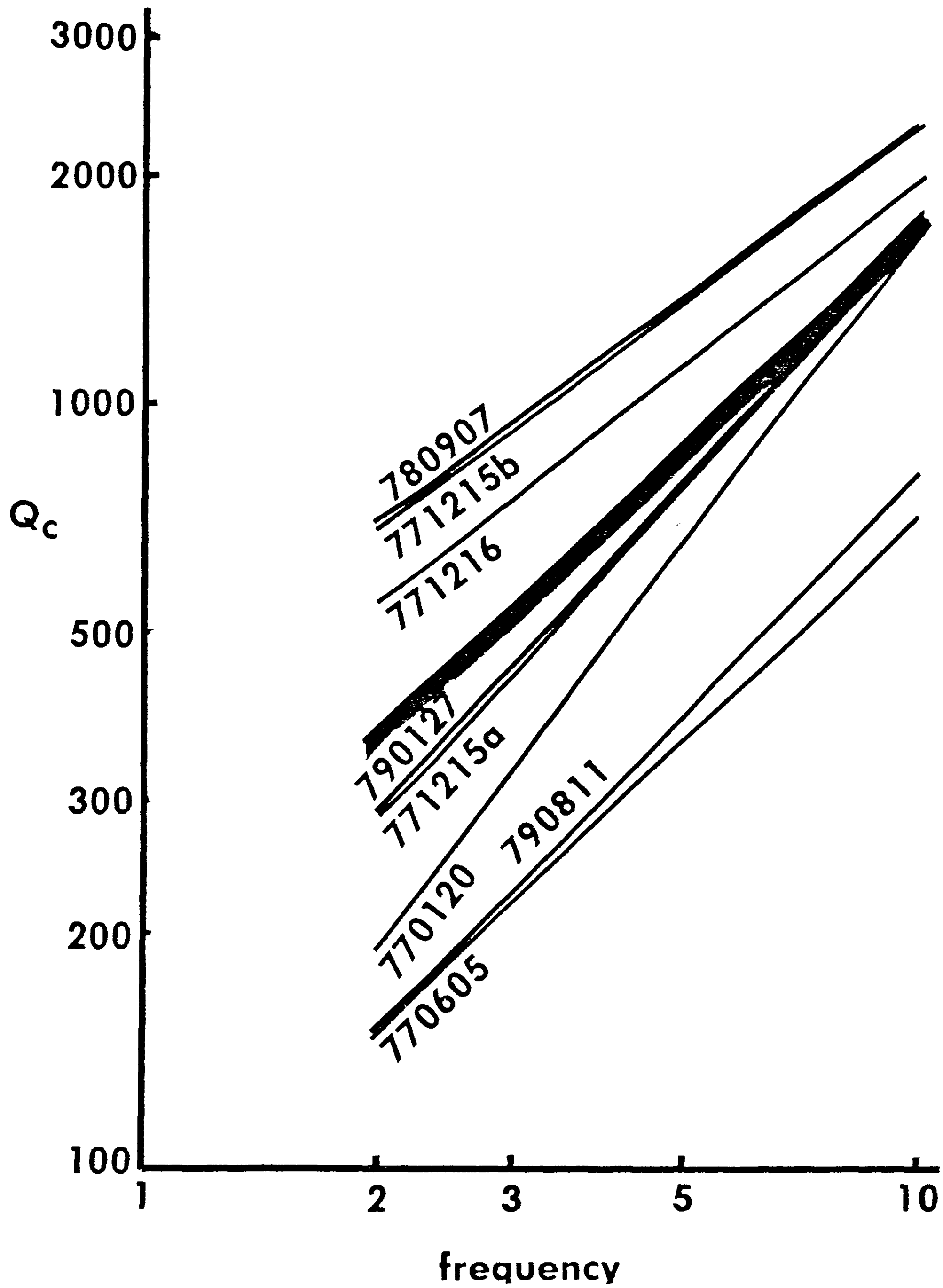

Figure 15. Summary of preceding results with numerical mean indicated by heavy line. Mean value follows the relationship $Qc = 190\ f^{.94}$.

Table 2

Coefficients for $Qc = Qo\ (f/fo)^n$

$fo = 1$ Hz

DATE	Q_2	Q_{10}	Q_0	n
770605	150	680	80	.93
770120	195	1560	80	1.29
771215a	290	1700	135	1.10
771215b	670	2250	400	.75
771216	540	1910	310	.79
780907	700	2250	420	.73
790127	290	1640	140	1.07
790811	150	800	70	1.06
ALL	360	1650	190	.94

Comparison with Other Results

Singh and Hermann (1983)

The values for Qo, presented in table 2, are not strikingly different from previously published values for Qc in the lithosphere. The values for the exponential increase, n, are higher than previously published values. Using this same method and programming, Singh and Hermann (1983) determined Qo and n over the United States. For the western United States, Qo was 200 to 800 and n was .4 to .6. In the central United States, Qo was 800 to 1300 and n .1 to .3. For the New England area, Qo was 200 to 800 (200 nearest the coast and 800 nearest the Appalachians) with n ranging from .3 to .4. There are several differences between Singh and Hermann's study and this one, however.

In the Singh and Hermann (1983) paper, frequencies studied were 0.5 to 5 Hz. In this study, we examine frequencies from 2 to 10 Hz. Stronger frequency dependence at higher frequencies is expected, as mentioned earlier, but the degree of increased dependence is uncertain. A comparison of this study's result with a couple of Singh and Hermann's (1983) results for New England is shown in figure 16. The two sets of results do not appear grossly incompatible especially when considering the amount of scatter

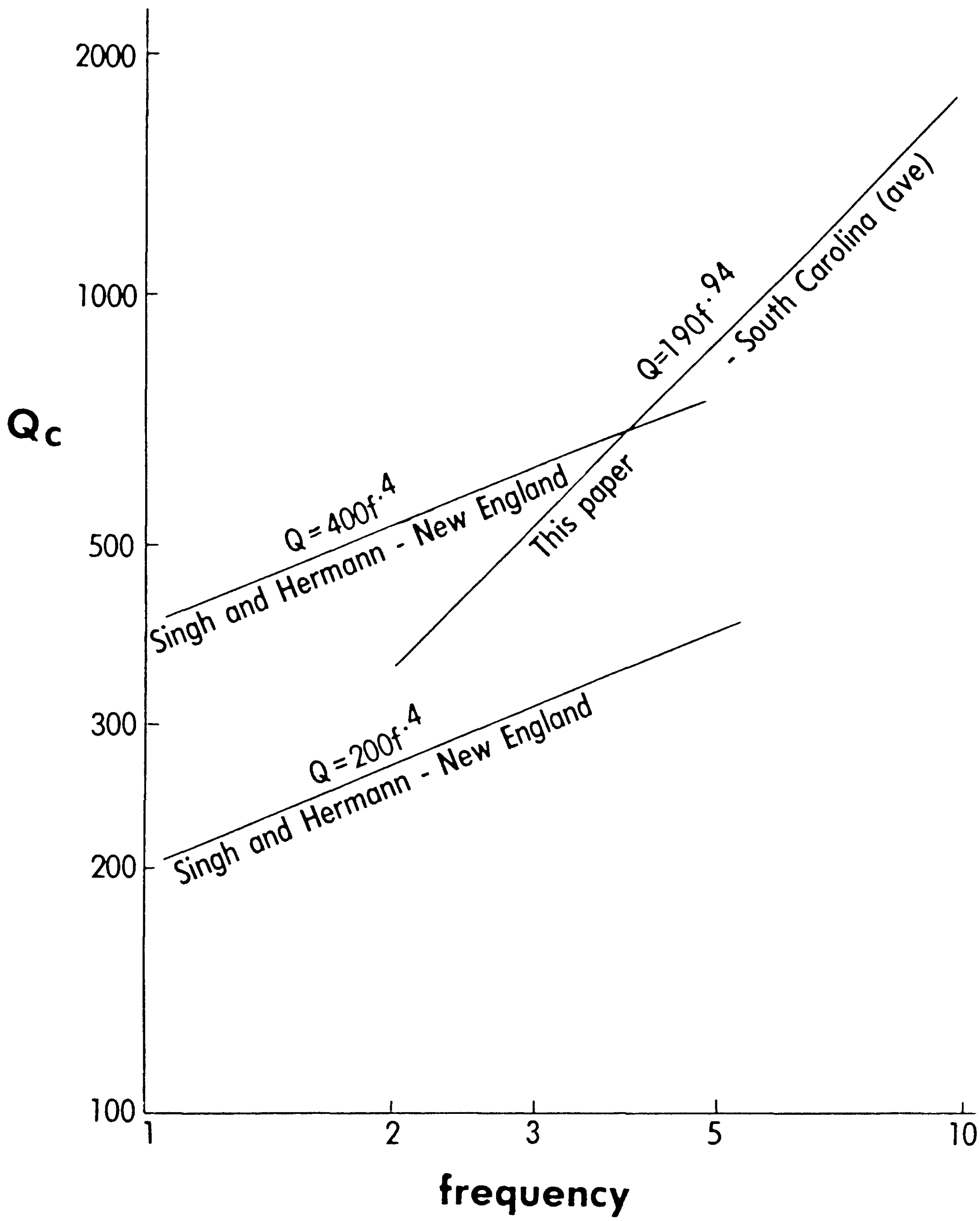

Figure 16. Qc versus frequency for this study and Singh and Hermann's (1983) northeast United States results.

in this study, and also probably in their study. One can visually smooth the set of lines into one line with variable slope.

Other differences between this study and Singh and Hermann (1983) are that they used larger magnitude earthquakes and longer source - receiver distances. The different magnitudes of earthquakes studied should not have any effect, again assuming the validity of Aki and Chouet's (1975) model in which the source does not influence attenuation. The longer path lengths between source and receiver would have an effect, since the longer paths could intercept parts of the lithosphere with different properties than the restricted, relatively shallow, crust in the Coastal Plain.

Dainty (1981)

Dainty (1981) showed that Qc measured in this way is influenced by anelastic and scattering losses. By comparing results of Aki (1980a) and Rautian and Khalturin (1978), he showed that although measured Qc's were different below 5 Hz, they approached a similiar value above 10 Hz. He thought the high frequency Qc was dominated by intrinsic Q, dependent on anelastic losses, and the lower frequency Qc's were dominated by scattering losses. Different values meant there were different mean free paths between scatterers in different regions. Scattering Q ranged from 200 to 500 at 2 Hz, while intrinsic, anelastic Q was constant at 2000. This favorably compares with results from this study, in that at low frequencies Qc is less than 500, and increases to more than 1500 at higher frequencies.

As discussed earlier, Qc should be considered the combination of intrinsic and scattered Q's. Assuming intrinsic Q dominates at high frequencies and does not radically change between 1 and 10 Hz, and assuming scattered Q dominates at low frequencies, theoretically we can make some estimate on the size of the mean free path in the crust. Dainty (1981) derived the equation $1/Qc(w) = 1/Qi + v/wL$ for this relationship. Here Qc stands for coda Q, Qi for intrinsic Q, w for angular frequency, v for shear wave velocity, and L for mean free path. Pulli (1983) applied this equation to his work, and determined the mean free path at 1 Hz, where scattering is thought to dominate. For the long duration Qc, he found that the mean free path was 400 kilometers, but that it was 80 kilometers for the short duration Qc. Pulli assumed the long duration result reflected properties in the mantle and the short duration result represented crustal properties. His interpretation makes physical sense in that the shorter free paths and greater frequency dependence determined from short duration codas imply the crust is more heterogeneous than the mantle, a likely situation. The absolute values for the lengths of the mean free paths is still open to question, however.

For the South Carolina study, assuming intrinsic Q is 1650, and using the 1 Hz Qc value of 190, gives a mean free path of 100 kilometers, not significantly different from Pulli's (1983) result. We do not believe this is an absolute number for this value, but it is a number to compare with other applications of Dainty's equation.

Pulli (1983)

Pulli (1983) examined the effect of determining coda Q from codas having durations longer than 100 seconds and shorter than 100 seconds. He found that Qc determined from short length records were more frequency dependent (higher n) and had lower values for Qo than Qc determined from long records. Figure 17 shows a result of his work. The lower curve shows the relationship of Qc to frequency determined

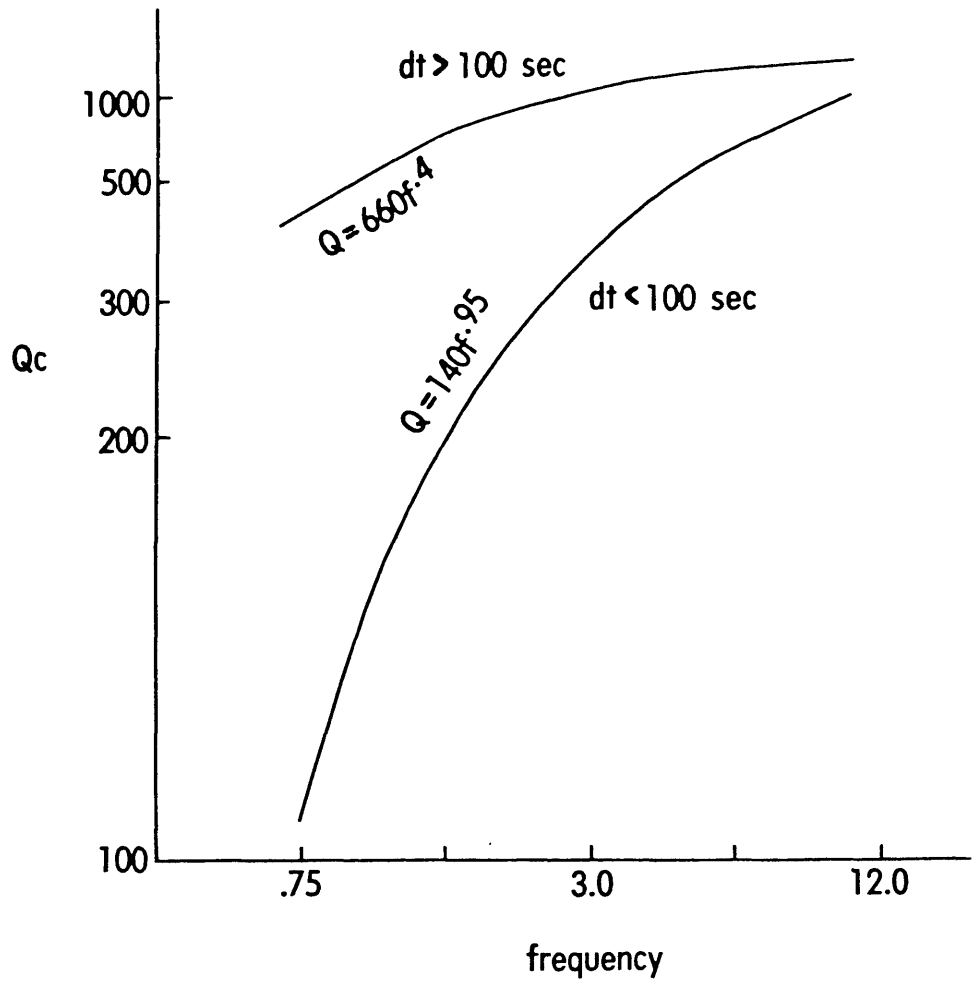

Figure 17. Qc versus frequency from Pulli (1983). Upper curve shows result from records having longer than 100 second codas, lower curve shows same for less than 100 second coda records.

to 1300 at 10 Hz. For the long duration codas, the upper curve shows Qc ranging from 660 at 1 Hz to 1600 at 10 Hz. In our study, the coda durations were all less than 100 seconds, and most were less than 45 seconds. The average Qc, (table 3 and figure 14) determined from all data, was 190 at 1 Hz increasing to 1650 at 10 Hz. The frequency dependence of Qc in our study was .94, compared to .95 for Pulli's (1983) short duration coda Qc. We can reasonably assume that if we had longer codas in the Coastal Plain, the Qc we determine would show less frequency dependence and have higher values at lower frequencies, just as Pulli's (1983) results show.

A Novel Temporal Variation

Assuming the curves plotted for each earthquake (figure 16) are exact values for Qc, there is a temporal variation of Qc. In figure 18, the value of Qc at 3 Hz was picked off the line for each earthquake except 770605. This earthquake's result was not used because it occurred in a different source zone and other factors could be influencing its value. The 3 Hz values from all other earthquakes were plotted versus time, in years. A low level existed prior to December, 1977, but then abruptly

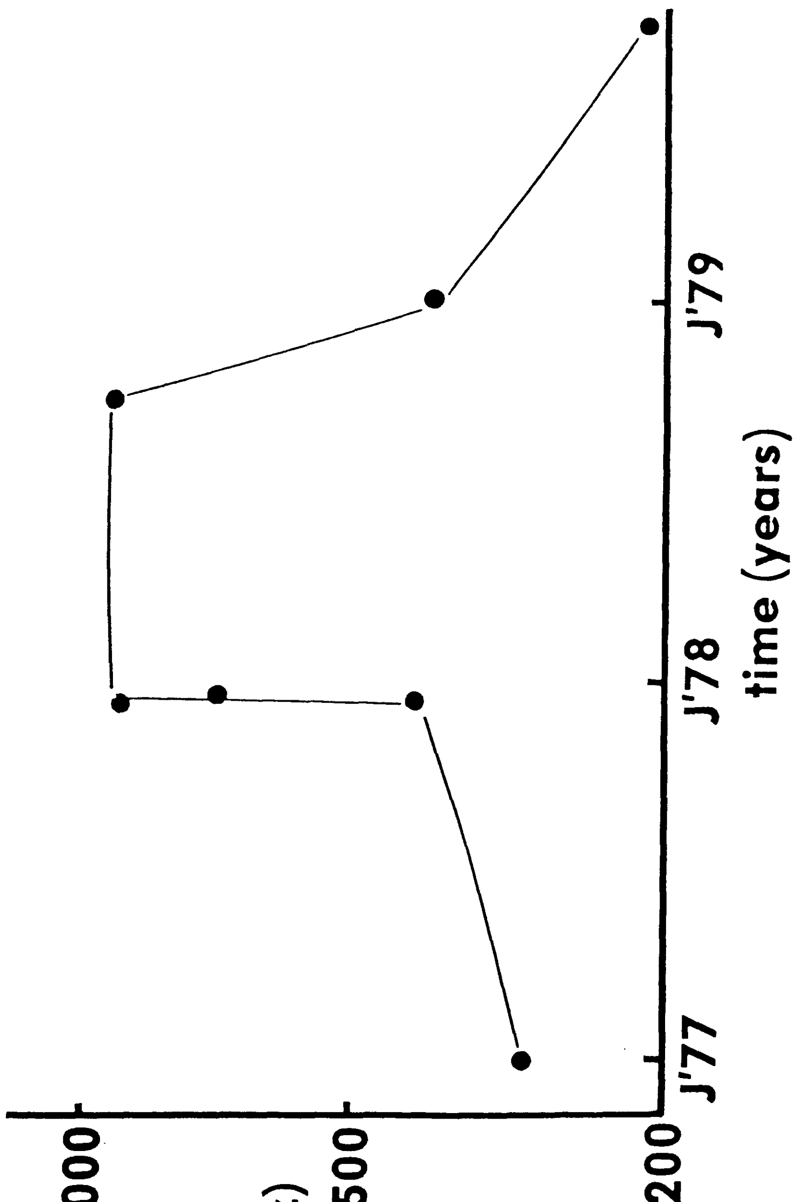

Figure 18. Qc at 3 Hz versus time in years for earthquakes near Charleston. Values plotted are from line fit to data for each earthquake.

increased. The value stayed high for almost a year, then dropped over the next year to a pre-increase level. The same observation results from plotting Qc at any frequency versus time. The increase in December, 1977, is coincident with an unusually high seismic activity rate in that there were six felt earthquakes in one week at that time. In no other single week or month, during 10 years of seismograph network operation, were there more than two earthquakes felt.

In making a statement regarding temporal variation in Qc, the earthquake location parameters must be carefully examined to rule out a systematic cause for change in Qc. Looking over table 1, the high values from the earthquake on 771215b appear to result from it being a larger earthquake. A contradicting observation is from the earthquake on 790127, at a similiar magnitude and depth, but with lower values at all frequencies. Also, the 790811 earthquake is of similiar magnitude with still lower values. The change in values, then, does not seem to be correlated with magnitude.

The change in values also does not appear associated with depth or location. All earthquakes were in the same 15 kilometer depth range and within a 20 kilometer radius of each other. It seems acceptable, then, to assert that the temporal variations observed in figure 18 are independent of source depth, magnitude, or location.

Temporal changes in Qc have a precedent. In Chouet (1979), a monotonic decrease in Qc was observed, determined from earthquakes that occurred during a one year aftershock study. This aftershock swarm occurred along the San Andreas fault zone, in central California. Qc in the frequency range between 1.5 and 24 Hz was studied. Chouet showed that the change in Qc was far in excess of possible errors on any individual estimate of Qc, and was independent of earthquake depth, magnitude, or distance from the recording stations. He did note that the temporal variations appear strongest at the higher frequencies (greater than 6 Hz). He concluded that the different shapes of the coda, hence coda decay, result from a combined effect of Qc variations in earth layers and of frequency dependent composition of the waves that compose the coda. Aki (1983, pers. comm.) indicated that other instances of similiar temporal variation in Qc, measured during an aftershock swarm, have been demonstrated in Chinese work.

A factor that can influence the value of Qc is the amplitude of ambient stress in a region. A change in the stress field would alter the shape and volume of void spaces in a material, thus altering the way frictional energy is dissipated. Stewart, et al, (1983) tested Q in laboratory samples, varying pressure and strain amplitude. They measured Q for compressional and shear waves in a dry sandstone using the bar resonance technique in the megahertz frequency band. The absolute value of changes they observed cannot be applied to our study, but trends in behavior can be examined.

One result of Stewart's, et al, (1983) shear wave experiments showed a ten-fold increase in Q when confining pressure was increased from 10 to 600 bars. For an increasing microstrain field from .3 to .75, there was imperceptible change in Q under high pressure, but up to a three-fold decrease in Q under low pressure.

In our study, the time changing Qc is correlated with time changing rate of seismic activity. In Stewart, et.al. (1983), study, changes in Q were correlated with changes in pressure and strain. Perhaps in the earth, changes in Qc can be correlated with changes in the pressure and strain fields, and hence with seismic activity rate.

Spatial Variation

Having demonstrated a temporal variation in Qc, it is reasonable to examine the data for a spatial variation also. We do so by first removing the assumed temporal variation, normalizing the data to the average of the time varying data. We worked with Qc from the unfiltered signal since the most data was available there.

To get a temporal variation correction factor, we averaged the value of Qc at 10 Hz from the least squares fitted lines (Qc(ave) = 1730). For each earthquake, we subtracted its Qc(10) value from the average value, and used this correction factor to reduce station data (example, 1730 - 1560 = 170 for 770120). Each earthquake had different correction factor, ranging from +930 (790811) to -520 (771215b and 780907). This value was added to the station data for each earthquake. These values are plotted in figure 19.

Of the 28 available path dependent values, 10 are greater than 2100. Two of these have very short path-lengths and may have error in them from this condition, and one other has a high error margin (21%). Six of the high values cluster in the eastern region of figure 19, while low values cluster in the western part. There are no

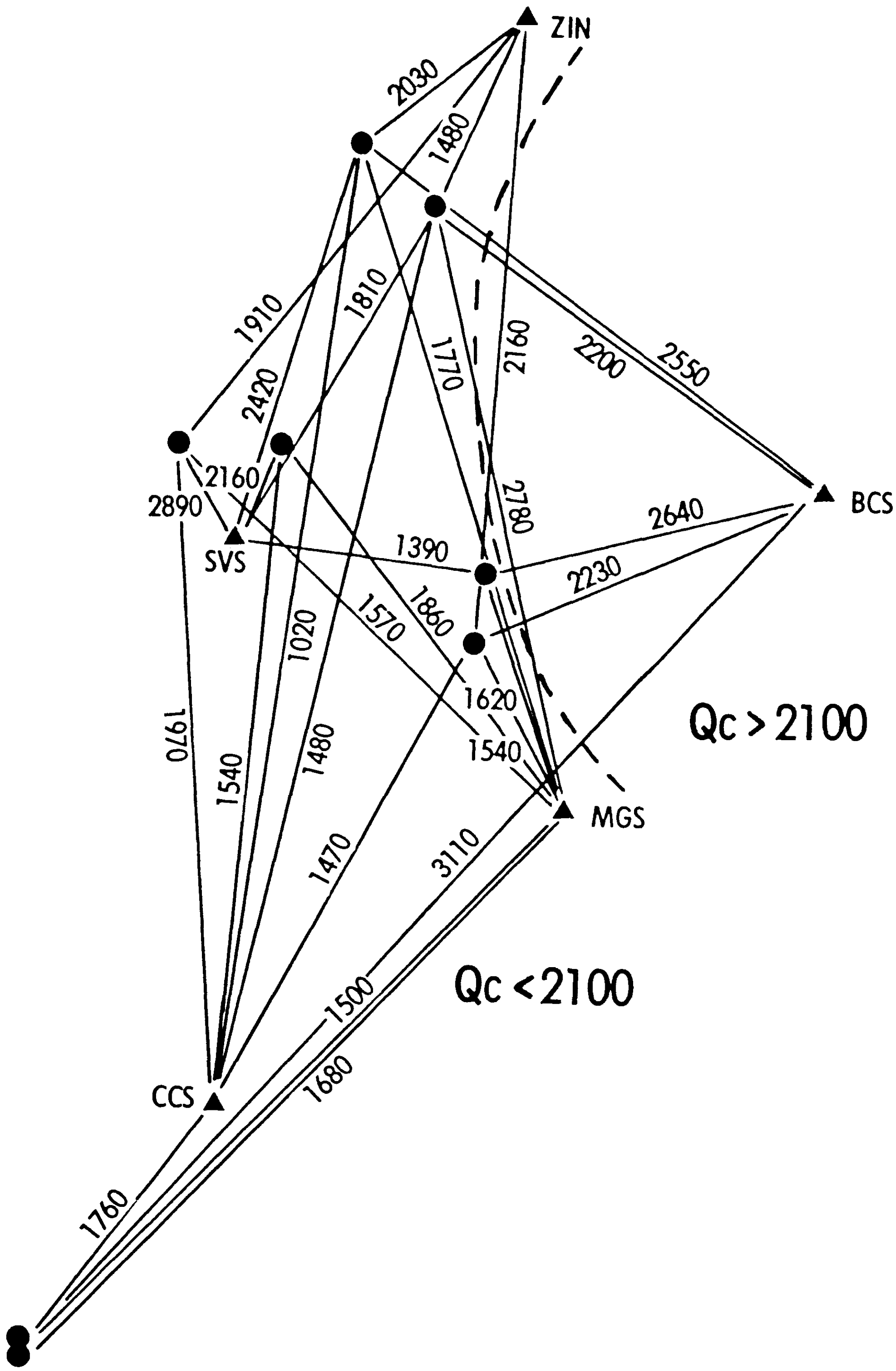

Figure 19. Unfiltered Qc, with temporal effect removed, by station to earthquake path.

low valued paths in the eastern region. Four of the six high value observations come from station BCS, but two other paths are also high. These are MGS and ZIN. Low valued observations come from these two stations for paths in the western part of the region. This suggests that station dependent characteristics are not the source for the spatial variation. We are also encouraged about the validity of all our numerical results in that there is not completely random behavior in the spatial variation.

Drawing a line between the high valued and low valued Qc zones, and the zone of even lower valued results for the 770605 earthquake (western part of the state), we see a steady increase in Qc (unfiltered) from the west to the east (figure 20). We are further supported in making a statement regarding spatial variation because this corresponds to a steady increase across the coastal plain.

The region with highest value Qc indicated from this study should also have been apparent from damage reports from the 1886 Charleston earthquake. Sloan's intensity map from this earthquake (Dutton, 1890), is reproduced, in part, in figure 21. The western edge of the innermost isoseismal, from Summerville to Middleton Hall, roughly corresponds to a line from the northernmost earthquakes in

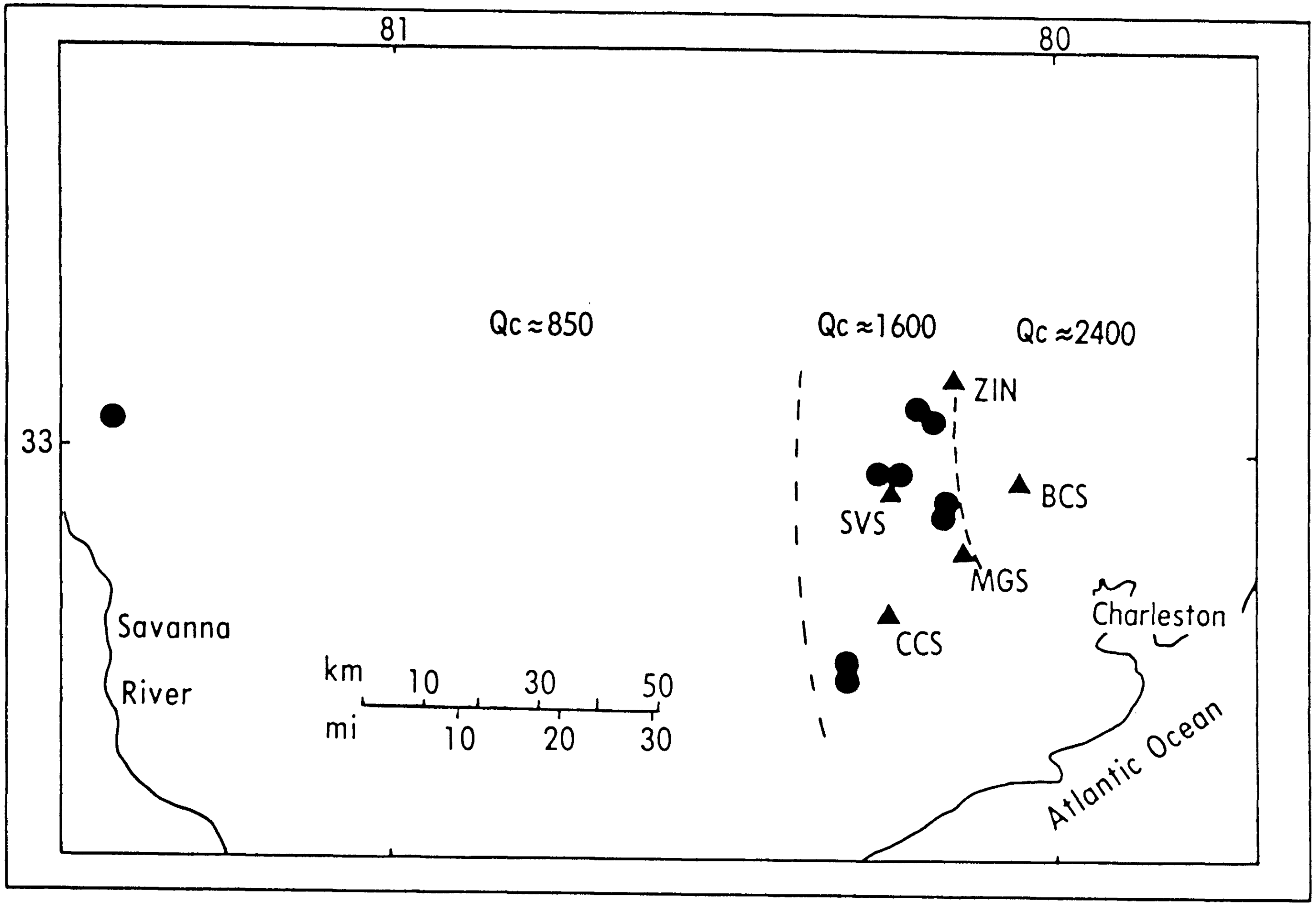

Figure 20. Regionalized Qc zones for the South Carolina Coastal Plain, showing region of low Qc to west, and higher regions to east.

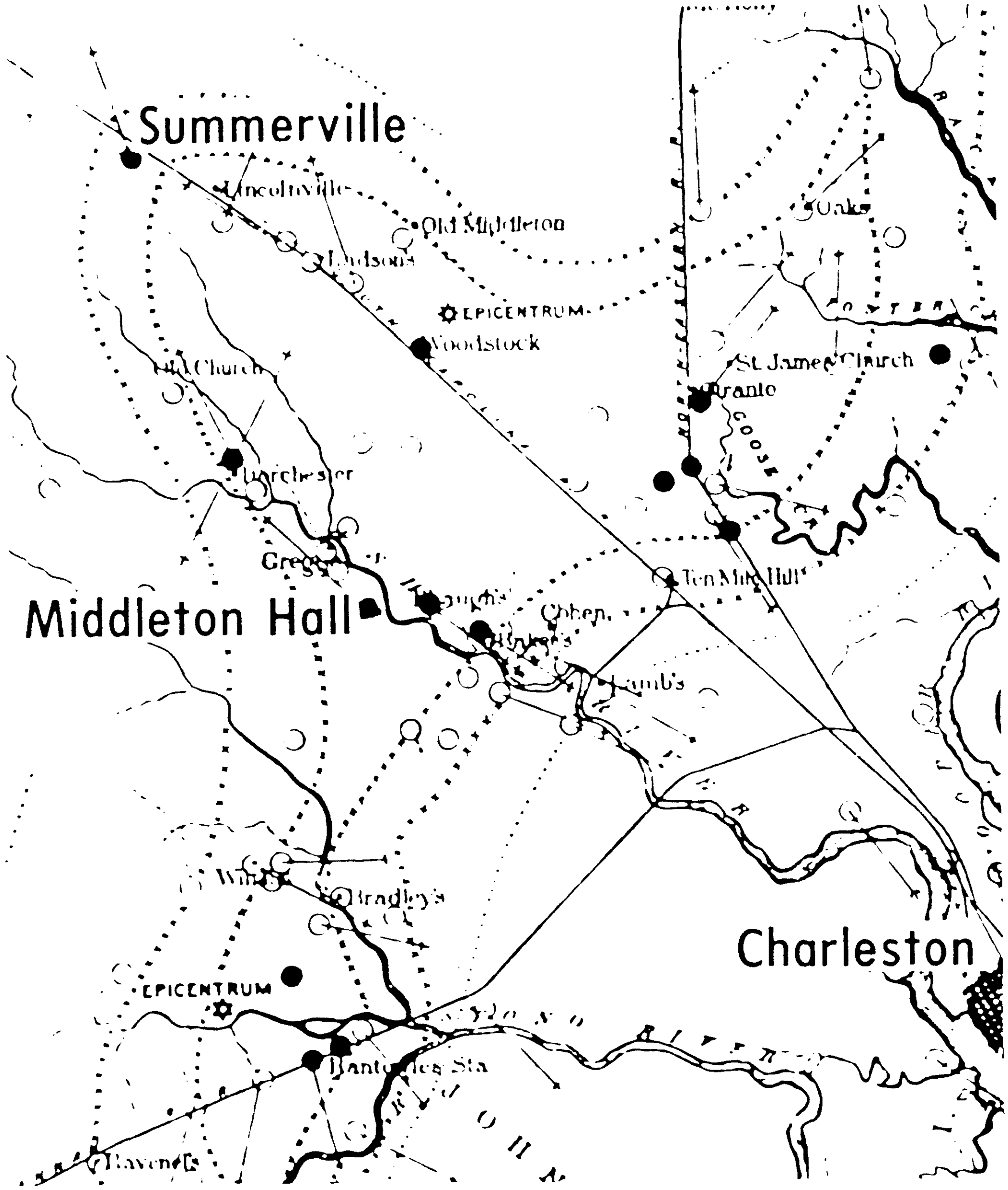

Figure 21. Part of Sloan's intensity map of the 1886 earthquake damage (Dutton, 1890).

figure 19 to station MGS. This roughly corresponds to the demarcation between high and low Qc zones in figure 20. Highest damage from the 1886 earthquake occurred east of this demarcation, or in the high Qc zone. The epicenter of the 1886 earthquake was chosen by Sloan and Dutton (Dutton, 1890) towards the center of the highest intensity zone, near the present location of station BCS. This location is also east of current seismic activity. It is possible that the location of the 1886 earthquake was misidentified because of spatial variations in energy transmission and decay.

CONCLUSIONS

Qc was found to be strongly frequency dependent between 2 and 10 Hz for small earthquakes in the South Carolina Coastal Plain. Averaging all Qc values results in the equation $Qc = Qo(f/fo)^n$, where fo is 1 Hz, frequency ranges from 2 to 10 Hz, Qo ranges from 70 to 420 with an average of 190, and n ranges from .73 to 1.29 with an average of .94. This relationship agrees with the results from Singh and Hermann (1983) for New England. In their study, the equation $Qc = (200\text{-}400)f^{.4}$, where frequency ranges from .2 to 5 Hz, describes frequency dependent Qc in the coastal regions of New England. At 3 Hz, their values range from 300 to 550, while the result from this study is 490.

An apparent temporal variation in the measured parameter, over a 2½ year interval, is suggested by the data. Qo apparently changed from 80 to 400 and down to 70 from January, 1977 to August, 1979. The increase was coincident with an unusual increase in the seismic activity rate. An explanation for this phenomenon comes from laboratory results, in which shear wave Q strongly increased with increasing pressure, and decreased with increasing strain (Stewart, et.al., 1983). The association of increased Qc and seismic activity rate in South Carolina may be related to the association of increased shear wave Q and increased pressure in the laboratory.

An apparent spatial variation in Qc was also observed, Qc at 10 Hz increasing from 800-900 in the western part of the state, to 1600 in the western part of the coastal area, to more than 2000 in the eastern part of the coastal area. The demarcation between high and low Qc in the area near Charleston coincides with the western edge of the highest intensity zone from the 1886 earthquake. The 1886 earthquake may be mislocated because of spatial variations in energy decay that were not identified at the time of that earthquake.

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