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Anelastic Structure and Evolution of the Continental Crust and Upper Mantle from Seismic Surface Wave Attenuation

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Abstract

Regional variations of the shear-wave quality factor (Q_μ) in both the upper crust and upper mantle of continental regions are large, with values in old stable cratons exceeding those in tectonically active regions in both depth ranges by as much as an order of magnitude or more. The frequency dependence of Q_μ in the upper crust also shows large regional variations, being low in tectonically active regions and relatively high in stable regions. Because of the large variations in Q_μ from region to region, it is easy to map regional variations of both upper crustal Q_μ and Lg coda Q, even though both measurements may be marked by large uncertainties. Both upper crustal Q_μ and Lg coda Q values correlate with the time which has elapsed since the most recent tectonic activity in continental regions. A tomographic image of the variation of Lg coda Q values across Africa clearly shows reduced Q values which correspond to recent tectonic activity in the East African system and other regions of Mesozoic or younger age. Reductions of Lg coda Q can also be detected which correlate with tectonic activity that occurred in the early Paleozoic during the coalescence of the cratons which formed that continent.

Lg coda Q values have been mapped for almost all of Eurasia. Low Q values span from southern Europe to regions of southeast Asia where recent tectonic activity has taken place. The highest values occur in the east European platform which has been relatively stable since the Precambrian. The

Siberian platform is characterized by lower values which may be related to tectonic activity which occurred in the Mesozoic era.

Q_μ values increase rapidly at mid-crustal depths, in a range which appears to coincide with the transition to the plastic lower crust. At lower crustal and upper mantle depths, Q_μ values decrease with increasing depth, possibly by progressive unpinning of dislocations with increasing temperature at those depths. Observed regional variations in upper mantle Q_μ values at depths of about 150 km can be explained by differences in temperature affecting a single atmomic attenuatin mechanism, but those at shallower depths cannot.

Regional variations of Q_μ in the upper crust are most easily explained by differences in the density of fluid-filled fractures in the upper crust which allow movements of fluids during propagation of seismic waves. Studies of the regional variation of Q_μ and Lg coda Q values indicate that crack density is greatest during and immediately following tectonic activity in a region and that it decreases with time following that activity. Permeability determinations in deep wells show that the fluid movement which reduce Q_μ values are restricted to major fracture zones, a situation which will produce widely differing values of Q in local studies, depending on the locatoin of the study relative to the fractures. The fluid volume in those cracks decreases with time by retrograde metamorphism or loss to the surface, causing a reduction in the number of open cracks and a concomitant increase in Q_μ.

Introduction

Seismic waves in the Earth are known to attenuate with distance of travel at rates greater than those predicted by geometrical spreading of the wave fronts. That excess attenuation might be caused by either intrinsic anelasticity of rock through which the waves travel or by inhomogeneities which refract, reflect, or scatter seismic energy. The diminution in seismic wave amplitudes produced by those factors must be taken into account when determining magnitudes of earthquakes or yields of explosions. This practical need spawned much of the research on seismic wave attenuation for the past four decades. Another rationale for studying seismic wave attenuation is the need to determine the dispersion produced by anelasticity. Attenuative dispersion is required to satisfy causality [*Lomnitz*, 1957; *Futterman*, 1962; *Strick*, 1967] and must be accounted for in order to reconcile velocity models of the Earth obtained from body-wave travel-times at high frequencies with those obtained from long-period surface wave or free oscillation measurements [*Jeffreys*, 1967; *Liu et al.*, 1976].

The loss of elastic energy which a seismic wave experiences when it traverses an imperfectly elastic material is commonly described by the inverse of the quality factor, or internal friction, Q^{-1}, which can be defined as

$$Q^{-1} = \frac{\Delta E}{2\pi E_{max}} \qquad (1)$$

where ΔE is the amount of energy lost per cycle and E_{max} is the maximum amount of elastic energy contained in a cycle. *O'Connell and Budiansky* [1978] recommended that average stored energy, rather than maximum energy be used in this definition, in which case the integer 2 in the denominator is replaced by 4. The latter definition has the advantage that Q can be written as the ratio of the real to the imaginary part of a complex elastic modulus.

Previous reviews of seismic Q in this journal have emphasized laboratory results, mechanisms, or radially symmetric Q models of the whole Earth [*Knopoff*, 1963; *Gordon and Nelson*, 1966; *Jackson and Anderson*, 1970; *Kanamori and Anderson*, 1977; *Karato and Spetzler*, 1990]. The present review

will discuss the regional variation of shear wave Q (Q_μ) throughout the crust and upper mantle of the Earth as inferred from the attenuation of seismic surface waves, including both fundamental-mode waves and higher modes. Fundamental-mode studies at periods between about 5 and 100 s (0.01-0.2 Hz) will provide average attenuation values over broad regions and allow us to determine Q_μ as a function of depth in the crust and upper mantle. Higher-mode studies will mainly be restricted to frequencies near 1 Hz and will concentrate on that superposition of higher modes which comprises the Lg phase and its coda. Information on Lg coda will allow us to place constraints on the frequency dependence of Q_μ and to study its regional variations in much greater detail than is possible with fundamental-mode data. The higher-mode data, because it encompasses a higher frequency range than does the fundamental mode, however, leads to the additional possibility of scattering effects obscuring those of intrinsic attenuation. The effects of scattering must either be taken into account or must be shown to be much less important than intrinsic Q in the types of study to be discussed. If we can isolate that component of attenuation which is produced by intrinsic anelasticity we can study properties such as temperature, state, fluid content of permeable rock, and movement of solid state defects, factors which are closely tied to the tectonic history of the crust and upper mantle, and which are not easily amenable to investigation using seismic velocities.

If we were to study intrinsic seismic wave attenuation on the moon, rather than on the Earth, there would clearly be a problem with scattering because that effect is so large there. Lunar seismograms exhibit very long and prominent codas (Figure 1) which dominate the recordings. *Dainty et. al* [1981] showed that the lunar coda is best explained by intense scattering in a near-surface layer which is characterized by low intrinsic absorption.

Absorption is small because there are no volatiles present in the moon, at least in those near-surface regions [*Tittman et al.*, 1976]. On the other hand, geological evidence [*Cathles*, 1990; *Torgersen*, 1990] and information from deep drill holes [*Morrow and Byerlee*, 1992; *Durham and Bonner*, 1993] indicate that the upper crust of the Earth contains abundant interstitial fluids which can migrate through cracks in igneous rock. Intrinsic absorption caused by that movement can drastically reduce seismic wave amplitudes [*O'Connell and Budiansky*, 1977; *Winkler and Nur*, 1979], an effect which will be especially severe for scattered waves which travel long distances as they bounce back and forth between scatterers. The smaller importance of scattered phases on terrestrial, as compared to lunar seismograms, is apparent from the seismogram comparison of Figure 1. The record from station CCM, in a high-Q region of the central United States, exhibits sharp onsets which are suggestive of direct wave arrivals, rather than scattering, and any significant coda is of brief duration, whereas the lunar coda is still prominent one hour after the onset of the wave motion.

This and other observations of seismic wave attenuation, discussed in a later section, lead us to interpret our results in terms of intrinsic Q in the Earth over the frequency range from about 0.01 to 1.0 Hz. It would be incorrect, however, to accept all attenuation observations as providing information on intrinsic Q in the Earth. It is well known that measurements of seismic wave amplitudes are susceptible to numerous types of error, both random and systematic in nature. Those errors can be large and, if not recognized, can lead to models of intrinsic Q of the Earth with spurious features that could erroneously be taken as real. For that reason, the following sections will discuss various methods of measuring surface wave attenuation and possible sources of error for each of them.

Although the possibility for error in amplitude measurements is large, there are also very large regional variations in intrinsic attenuation throughout the continental crust and upper mantle. The span of values reported in numerous studies exceeds two orders of magnitude. For that reason, even though there may be large uncertainties in Q models derived from surface-wave attenuation data, we may still be able to obtain much information which pertains to regional variations of Q and its relation to crustal and upper mantle structure and evolution.

Figure 2 dramatically illustrates the large differences which regional variations in of Q can have on seismograms recorded along high-Q and low-Q paths in continental regions of the Earth. The two broad-band IRIS stations both recorded wave motion produced by a magnitude 4.6 earthquake which occurred in southeastern New Mexico. The epicentral distance to CCM, along a predominently high-Q path, is 1256 km and that to PAS, along a predominently low-Q path is 1417 km. If the anelastic properties along the path to PAS were the same as those to CCM we would expect the amplitudes recorded at PAS to be about 80-90% as large as those recorded at CCM. The attenuation is so much greater along the path to PAS, however, that neither the P wave (arriving just before 3 minutes at CCM) nor the S wave (arriving at about 5 minutes at CCM) can be observed at PAS. The maximum amplitude of Lg, the largest phase on both records, is reduced by a factor of about 7 at PAS relative to that at CCM. Moreover, the predominant frequencies for signals recorded at PAS are significantly lower than those recorded at CCM. These observations indicate that attenuation along the path through the western United States to PAS is much greater than that along the path through the central United States to CCM. Similar observations have been reported for fundamental-mode Rayleigh waves generated by two nuclear explosions in Colorado [*Mitchell*, 1975]. Earlier evidence for differences in attenuative properties between the eastern and western United States came from studies of the relative fall-offs of Pg and Lg energies in the Basin and Range province and the central United States [*Sutton et al.*, 1967].

Fundamental-mode Attenuation Coefficients

The decay of surface wave amplitudes, in excess of that produced by geometrical spreading, can be described by $e^{-\gamma x}$ where γ is termed the attenuation coefficient. It is related to the quality factor, Q, by the expression $\gamma_{R,L}=\pi f/U_{R,L}Q_{R,L}$ where f is frequency, U is group velocity, and the subscripts R and L refer respectively to Rayleigh or Love waves.

Inversions of surface wave attenuation values lead to models of intrinsic shear-wave Q (Q_μ) or its inverse, internal friction (Q_μ^{-1}), and usually assume that the medium of travel can be described by a laterally homogeneous layered structure. If that were true for the real Earth, measurements of γ would yield values which are not contaminated by effects such as lateral refraction, multipathing, or focusing/defocusing which are produced by lateral complexities in crustal or upper mantle structure. Experience has shown, however, that determinations of surface-wave attenuation coefficients contain large uncertainties produced by those effects. For that reason, attenuation measurements require great care and careful selection of sources, station locations, and paths.

One way to minimize the effect of lateral structural variations is to use paths which are as short as possible. Unfortunately, however, surface waves attenuate very slowly, especially in high-Q regions. Since amplitude decay is small for short distances of travel it is difficult to obtain precise and reliable measurements of the attenuation coefficients for those paths. To obtain

precise measurements of attenuation it is preferable to use long paths, but with long paths we increase the chance of introducing systematic errors into our measurements.

The best method to use for a particular region will depend on several factors, including the availability of two-station paths, suitably located sources and, in the case of single-station measurements, the availability of sources with known focal mechanisms. Because we cannot always count on the occurrence of earthquakes in places which would be suitable for the application of a particular method, several methods for determining surface wave attenuation coefficients have been developed, some being suitable for some situations and some for others. These will be briefly described in the following paragraphs.

The Two-station Method

The groundwork for modern surface-wave attenuation research was laid by *Satô* [1958] when he applied Fourier analysis to the study of surface wave amplitude decay and Q determinations. *Solomon* [1972] applied Fourier analysis to the decay of surface wave amplitudes between two stations in a study of upper mantle Q and *Tsai and Aki* [1969] applied it in a study of crustal Q in western North America. It has formed the basis of several subsequent studies in both continental and oceanic regions [*Canas and Mitchell*, 1978, *Hwang and Mitchell*, 1986; *Al-Khatib and Mitchell*, 1991]. The method uses the equation

$$\gamma(\omega) = -\frac{\ln\{[A_2(\omega,r)/A_1(\omega,r)]\sqrt{\sin\Delta_2/\sin\Delta_1}\}}{r_2-r_1} \qquad (2)$$

where $A_{1,2}$ are observed spectral amplitudes at stations 1 and 2 at distances $r_{1,2}$ (in km), and $\Delta_{1,2}$ (in radians) from the source, and ω is angular frequency. An ideal situation in which to apply the method would be one in which the two stations are separated by a relatively large distance, across which elastic and anelastic properties to not vary laterally, and the earthquake source lies on the great circle path which passes through the stations. In addition, the path between the source and nearest station should, ideally, also be free of major lateral complexities which might cause the surface waves to refract laterally or produce focusing or defocusing at one station or the other.

If surface waves are refracted laterally between the source and stations, it then becomes likely that the waves recorded at one station will have left the source at a different azimuth from those recorded at the other station. When that happens variations in the source radiation pattern can produce very large amplitude differences at the two stations which will not be related to anelastic properties of the crust or upper mantle along the path between the stations. This effect is illustrated schematically in Figure 3 which shows a path to the near station which has left the source near a node in the radiation pattern while the path to the far station has left the source from a portion of the radiation pattern where the amplitudes are larger. If the difference between the source amplitudes is sufficiently large, negative attenuation coefficient values could be obtained. The possible arrival of waves from azimuths which depart from the great-circle direction can be checked by plotting the particle motion observed at each station. Polarization analyses can be performed in the time domain [*Vidale*, 1986; *Mitchell et al.*, 1993], in the frequency domain [*Lerner-Lam and Park*, 1989], or in a frequency-time domain [*Levshin et al.*, 1992].

Advantages of the two-station method are that it does not require knowledge of the earthquake source mechanism and, in the absence of effects produced by non great-circle path propagation, it should produce average values for the attenuation coefficients along the path between the two stations. Average values should be obtained, even if the anelastic properties

along that path vary with position. If two suitably located stations are available and if non-great-circle propagation can be avoided, it is therefore the method of choice.

Single Source - Many Station Methods

Tsai and Aki [1969] developed a method for simultaneously determining source spectral amplitudes and average surface wave attenuation coefficient values for a broad region surrounding an earthquake. It uses spectral amplitudes of surface waves recorded at several azimuths and distances from an earthquake with a known fault-plane solution. Tsai and Aki equalized observed spectral amplitudes to a selected reference distance r_0 using

$$A_{r_0} = A_r \left(\frac{R_e \sin \Delta_r}{r_0} \right)^{\frac{1}{2}} \exp\left(\frac{\pi f r}{UQ} \right) \qquad (3)$$

where A_r and A_{r0} are spectral amplitudes at the epicentral distance r and the reference distance r_0, respectively, R_e is the radius of the Earth, Q is the quality factor for Rayleigh or Love waves, and U is the corresponding group velocity taken directly from the seismograms used. Taking the logarithm of both sides leads to an expression which can solved by linear least-squares to obtain attenuation coefficient values over the region of interest, as well as the seismic moment of the event.

An advantage of the method is that it yields values for the standard errors of the unknown parameters in a straightforward way. A disadvantage is that it can lead to systematic errors in the determination of surface wave attenuation coefficients if it is used in a region in which anelastic properties vary laterally [*Yacoub and Mitchell*, 1977]. This is illustrated schematically in Figure 4 for a hypothetical case in which an earthquake occurs near a boundary which separates two regions with differing values of shear wave Q (Q_μ). The slope of the amplitude-distance curve would lead to an attenuation coefficient value which is lower (and could be negative) than that which would be produced if either of the Qμ values in the model were uniform throughout the entire model.

A variation of the *Tsai and Aki* [1969] method which can be used when a source mechanism is not available was developed by *Mitchell* [1975]. Following *Toksöz and Kehrer* [1972], he expressed the radiation pattern of an earthquake source as a superposition of a circular pattern (from an explosion source) and the pattern produced by a vertical fault with pure strike-slip motion. It requires a non-linear least-squares solution of the expression

$$W(\omega) = W_e(\omega)[1 + F(\omega)\sin 2[\{\theta - \theta_0(\omega)\}]\exp[-\gamma(\omega)r] \quad (4)$$

where the unknowns are the surface-wave attenuation coefficient, γ, the apparent source spectral value given by W_e, the apparent strike of the fault, θ_0, and a factor, F, which expresses the relative effect of fault motion to that of the explosion. Stations are located at various distances, r, and azimuths, θ, from the source. Variations in the value of F can produce a wide variety of radiation patterns which mimic those for real faults. Increasing values produce radiation patterns which can progress from being circular (when F=0) to being oblong, two-lobed, and four-lobed. In most inversions, values obtained for F, θ_0, and W_e have no physical meaning and only the value for the attenuation coefficient (γ) is realistic.

The method of *Mitchell* [1975], like that of *Tsai and Aki* [1969], is most useful in regions which are relatively laterally uniform in their anelastic properties. Application of either method in regions of laterally varying Qμ can produce attenuation coefficient values which are systematically biased either too high or too low, depending the station configuration with respect to the Q distribution and the source. It is possible to reduce these biases by using a regionalized approach to the inversion [*Yacoub and Mitchell*, 1977; *Patton and Taylor*, 1984], but there must be some *a priori* knowledge of the manner in which the Q values vary laterally.

Single Source - Single Station Methods

One goal of regional attenuation studies is to characterize the anelastic properties of single tectonic provinces. This is often not possible using either the two-station method or single source - many station methods. Two stations may not be available which lie along great-circle paths to a useful earthquake, or the source-station configuration may be adequate, but paths between the source and station pair may be long and traverse complex structure which produces lateral refraction and/or focusing of surface waves. Single source - many station methods are only useful in situations where several stations surround a source and all lie within a single province, or if it is possible to subdivide the region of study before determining attenuation coefficients. Single-source single-station methods avoid the necessity of having two stations which lie along a single great-circle path through the source and have the great advantage that it is often possible to find earthquake or explosion sources which lie within the same tectonic province as that for the station, and therefore may provide a relatively uniform path of travel for surface waves.

With single-source single-station methods, however, we must either know the source depth and focal mechanism for each earthquake used, or must be able to determine those parameters from the data available. At least two approaches have been used in this class of methods. *Cheng and Mitchell* [1981] used combined determinations of higher-mode amplitude spectra and fundamental-mode spectra along with a forward modeling approach to obtain Q models of the upper crust. *Chan et al.* [1989] and *Seber and Mitchell* [1992] used spectral shapes of the fundamental mode in their studies but, by doing so, had to use events which were small enough so that the source time function did not adversely affect their results.

Observed Fundamental-mode Attenuation Coefficient Values

Surface wave attenuation coefficients have now been determined for several continental regions. Figure 5 presents many of those determinations for Rayleigh waves which have propagated across stable portions of continents. The results of several studies, which include some paths across stable regions, have been excluded from that plot because they span more than a single tectonic province [*Tsai and Aki*, 1969; *Solomon*, 1972; *Yacoub and Mitchell*, 1977]. In addition, the plot includes only those values for which the standard deviations are less than 0.2 x 10^{-3}. Several values from studies of *Chen* [1985] and *Hwang and Mitchell* [1986] were not plotted for that reason. The plot indicates that at periods between about 20 s and 60 s, the attenuation coefficient determinations fall within a narrow range of values centered at about 0.15 x 10^{-3} km^{-1}. At longer periods the values appear to decrease slightly, although the standard deviations for those points would allow higher values which are comparable to those obtained in the 20-60 s period range. At periods less than about 15 s the attenuation coefficient values increase with decreasing period and reach values of about 0.5 x 10^{-3} km^{-1} at periods of 5-7 s.

Figure 6 presents results from several studies in tectonically active regions. As in Figure 5, most of the data points have standard deviations of 0.2 x 10^{-3} or less. Two exceptions have, however, been made in the short-period portions of the plot [*Patton and Taylor*, 1984; *Lin*, 1989] where, because of the shorter wavelengths at those periods and the laterally complex structure in active regions, those waves are more likely to be adversely affected by lateral changes in structure. For those two cases, where the attenuation coefficient values and their standard deviations are both larger than those at longer

periods, data have been included with standard deviations as large as 0.3 x 10^{-3}. At periods in the 20-100 s range the mean attenuation coefficient value stays relatively constant at about 0.25 x 10^{-3} km^{-1}. At shorter periods the attenuation increases sharply with decreasing period and reaches values as great as 3.0 x 10^{-3} km^{-1} at periods near 5 s. Three sets of values stand out as deviating systematically from the mean values on the plot. Those are values from the western margin of the United States [*Al-Khatib and Mitchell*, 1991], from a path across a broad portion of the western United States [*Solomon*, 1972], and from a study which included a broad portion of the Basin and Range province of the western United States and adjacent regions [*Patton and Taylor*, 1984]. The low values at periods in the 20-30 s period range in the study of *Solomon* [1972] may be due to focusing because the paths used are long and cross many major structural features at oblique angles. The high values of Patton and Taylor [1984] at 15-40 s periods in the Basin and Range province may occur because of the application of a single event-many station method in a region where Q is apt to vary laterally (see the previous section). Alternately, there may be regions of low Q in the part of the Basin and Range studied by Patton and Taylor which are not present along paths used by *Solomon* [1972], *Chen* [1985], or *Lin* [1989]. The values for the western margin of the United States [*Al-Khatib and Mitchell*, 1991] may reflect real low-Q material in the upper mantle, but the possibility that defocusing of energy occurred at the second of the two stations cannot be ruled out.

Figure 7 summarizes the results of Figures 5 and 6 by grouping the data into that which corresponds to stable regions and that which corresponds to tectonically active regions. The two data sets are also compared to a set of oceanic fundamental-mode attenuation coefficient values obtained over the period range 20-100 s for three age regions of the Pacific [*Canas and Mitchell*, 1978]. The highest Pacific values correspond to that part of the Pacific which formed between 0 and 50 M.y. ago, the intermediate values correspond to lithosphere which is between 50 and 100 M.y. in age, and the lowest values correspond to those portions of the Pacific lithosphere which formed more that 100 M.y. ago. Tectonically active regions are characterized by higher attenuation coefficient values than are regions which are stable. Attenuation coefficients for periods greater than about 45 s for stable regions lie within the error bounds for the span of values for the three oceanic regions. At periods between 25 and 45 s the lowest attenuation coefficients which are observed are those for the age region of the Pacific > 100 s, but there is overlap in the error bars for the data from the two types of region.

At periods shorter than about 20 s the values obtained in stable regions and those obtained in tectonically active regions diverge; those in tectonically active regions show a sharper increase with decreasing period, achieving values which are over three times higher than those in stable regions at periods near 5 s. That increase is evident both in the values obtained by *Lin* [1989] for one path within the Basin and Range province and by *Patton and Taylor* [1984] for a broader region of that province.

Lg Coda Q

The preceding section described difficulties in the measurement of surface wave attenuation for the fundamental mode. Those measurements are usually made at frequencies less than 0.2 Hz and since the factors which give rise to errors in those measurements are dominantly systematic in nature, their effects can sometimes be reduced by careful selection of sources, stations, and procedures used. The higher-frequency waves which comprise Lg and its coda can be adversely affected, like the lower-frequency

fundamental mode, by systematic errors, but also can be affected by random perturbations from an assumed laterally homogeneous layered structure [*Xie and Mitchell*, 1990b]. The systematic deviations from a laterally homogeneous layered structure may occur as a random, but laterally varying statistical distribution, or may be coherent over a sub-region of interest [*Cara et al.*, 1981]. Stationary random perturbations can be treated as a statistical distribution which is laterally stationary [*Aki and Richards*, 1980]. Such a distribution may be used to characterize random lateral variations in velocity and layer thickness [*Kennett*, 1989] as well as other scatterers. Systematic deviations include large-scale disruptions of crustal wave guides and near-receiver site effects, as well as other conditions, all of which can also affect lower-frequency fundamental-mode waves. An example of such a deviation was documented by *Xie and Mitchell* [1990b] in the western United States where coda Q determinations using coda for which scattering ellipses overlapped a significant portion of oceanic crust were systematically different from those where scattering ellipses were largely confined to continental crust.

Methods used to determine Lg coda Q include spectral amplitude decay of band-pass filtered records [*Aki and Chouet*, 1975], the change of predominant frequency with time in the coda [*Herrmann*, 1980], and stacking of spectral amplitudes observed for several time windows along the coda [*Xie and Nuttli*, 1988]. The last of these has permitted coda Q determinations which have sufficient stability and precision to permit the application of coda Q tomography where event and station coverage is sufficient.

Lg Coda Q Tomography

Lg coda Q tomography permits study of regional anelasticity variations on continental scales. It can be applied if a sufficient number of earthquakes and recording stations are well-distributed across the region of interest. We divide that region into a number of rectangular grids and parameterize the unknown lateral variation of Lg coda Q by assuming it to have a constant value, Q_m, inside the mth grid. We take Q_n to be the Q value obtained using the stacked spectral ratio method and assume that it represents the areal average of Lg coda Q in the elliptical area sampled by coda waves at the maximum lapse time at which the measurement is made. If we denote the area of the portion of the nth ellipse which overlaps the mth grid by s_{mn}, we obtain

$$\frac{1}{Q_n} = \frac{1}{S_n}\sum_{m=1}^{N}\frac{s_{mn}}{Q_m} + \varepsilon_n \qquad n=1,2,\ldots,N \tag{5}$$

where

$$S_n = \sum_{j=1}^{N} s_{jn}$$

and ε_n is the residual which results from the errors in modeling Lg coda and in the measurement of Lg coda Q. Numerical considerations in applying this equation are discussed by *Xie and Mitchell* [1990a]. Because the results of coda Q tomography are presented as maps which show the variation of Lg coda Q and its frequency variation over large portions of continents, it well-suited for studying the relationship of Lg coda Q to the evolution of continents. It was first applied to Africa [*Xie and Mitchell*, 1990a] and has since been applied to Eurasia [*Pan et al.*, 1992]. A summary of both sets of results is presented later in this section. The African summary includes a new interpretation in terms of the tectonic evolution of that continent.

Is Lg Coda Q a Measure of Intrinsic Absorption or Scattering?

Shear waves recorded locally and Lg waves recorded at regional distances exhibit a coda which cannot be explained by deterministic modeling with plane layered structures [*Aki and Chouet*, 1975]. A curious aspect of Lg coda is

that, even though it consists of scattered waves, measured values of Q_{Lg} are not necessarily low in regions (such as those with much topographic relief) where severe scattering might be expected to occur. For instance, a map of Lg coda Q derived for the United States [*Singh and Herrmann*, 1983] exhibits values of about 600 in the Rocky Mountains where relief is high and only 150-300 in the Basin and Range where relief is much smaller. Similarly, Lg coda Q values in the Appalachian Mountains are higher than those obtained for the Atlantic coastal plain. These types of qualitative observations led *Mitchell* [1980] to infer that Lg coda Q values were determined by intrinsic Q_μ in the crust and to conclude that it would be valid to combine observations of Lg coda Q with attenuation coefficient data to infer frequency-dependent Q_μ models for the continental crust. There are other types of observational evidence which also support that inference. First, the rapid fall-off of coda amplitudes of waves traveling in the Earth relative to those traveling in the moon (Figure 1) is due to a higher intrinsic Q and greater degree of scattering in the moon than in the Earth [*Dainty and Toksöz*, 1981]. It is well known that intrinsic Q is high in lunar material because the volatile content is very low there [*Tittman, et al.*, 1976]. The presence of volatiles in crustal rock in the Earth, on the other hand, will drastically reduce intrinsic Q values [*O'Connell and Budiansky*, 1977; *Tittmann*, 1977]. Second, as discussed in the following section, patterns of regional Lg coda Q variation are the same as those of Q_μ in the upper crust inferred from studies of fundamental-mode surface wave attenuation. Third, measurements of attenuative body-wave dispersion in the Pyrenees [*Correig and Mitchell*, 1989] were found to be consistent with those which would be predicted by a Q model inferred from the decay of spectral amplitudes. That consistency would not occur if the crust in that region were characterized by high intrinsic Q values and high scattering. Fourth, the equality of Q and its frequency dependence for Lg waves and Lg coda in the Basin and Range province [*Xie and Mitchell*, 1990b] suggests that intrinsic attenuation is high and is the predominant mechanism affecting both types of wave.

Recent computational [*Frankel and Wennerberg*, 1987; *Hoshiba*, 1991; *Mayeda et al.*, 1992] and theoretical [*Zeng*, 1991; *Zeng et al.*, 1991] studies also indicate that intrinsic attenuation far outweighs scattering attenuation in contributing to the decay of coda waves. *Wennerberg* [1993] summarized the results of much recent work and concluded that if the scattering model of *Zeng* [1991] is correct, changes in intrinsic attenuation are 5 to 15 times as significant for determining coda attenuation than are comparable changes in scattering attenuation. Contrary interpretations have been presented, based on the agreement of computed frequency dependence for scattering models with that observed [*Dainty*, 1981; *Wu*, 1982; *Dainty*, 1984]. Intrinsic Q models of the crust are, however, also characterized by being frequency dependent and can also explain that agreement.

Observed Lg Coda Q Values

Numerous single-path determinations of values for Lg coda Q and its frequency dependence have been made in continental regions, mostly at frequencies of about 1 Hz. Measurements Q_{Lg} usually assume that those values can be represented as $Q_{Lg}=Q_0f^{\eta}$ where Q_0 is the value of Q_{Lg} at 1 Hz and η indicates its frequency dependence. Lg coda Q is similarly represented. Regions which have been stable for a long time are usually characterized by relatively high values of Lg coda Q (800-1200) and relatively low values of frequency dependence (0.0-0.3) at frequencies of 1 Hz. These ranges of values occur in the portion of North America east of the Rocky Mountains [*Nuttli*, 1973; *Singh and Herrmann*, 1983], in the stable cratons of South America [*Raoof and Nuttli*, 1985], and in the Indian Shield [*John*, 1983]. Lower Lg coda Q

values (150 - 400) and higher values of frequency dependence (0.3 - 1.0) characterize tectonically active regions, such as western North America [*Sutton, et al.*, 1967; *Singh and Herrmann*, 1983; *Chávez and Priestley*, 1986; *Nuttli*, 1986; *Xie and Mitchell*, 1990b], western South America [*Raoof and Nuttli*, 1985], and the Himalaya [*John*, 1983]. This is also true for Q measurements of the direct Lg phase because Lg Q and Lg coda Q are usually found to be nearly the same whenever they are both determined in the same regions [*Nuttli*, 1988]. Q values for body waves measured at relatively short distances are also high in the stable craton of eastern North America [*Al-Shukri et al.*, 1988] and can be low in tectonically active regions such as California [*Li et al.*, 1994], China [*Jin and Aki*, 1988], and the Pyrenees Mountains [*Correig et al.*, 1990].

The tomographic method, described earlier, allows us to determine lateral variations of Lg coda Q over broad regions. Figure 8 displays a tomographic image of Lg coda Q for Africa which is slightly modified from that of *Xie and Mitchell* [1990a]. The most conspicuous feature of the map is the band of low Q values which lies along the East African Rift Zone, a region known to currently be tectonically active. Because the resolving power of these measurements is limited to be between about 800 and 2000 km, depending on data coverage [*Xie and Mitchell*, 1990a], it is possible that the low Q values in the crust which produce that band are really restricted to a much narrower region, characterized by lower values, than that shown on the map. Other regions of low Q values are the Cape Fold Belt, at the southern tip of Africa, which is part of an orogenic belt of late Paleozoic - early Mesozoic age, the region of the Atlas Mountains in Morocco which are of Alpine age, and the Cameroon Line which proceeds northeastward from the bend in the western coast and is the site of recent volcanic activity [*Cahen et al.*, 1984].

Almost all of the rest of Africa has been stable since the Pan African orogeny which ended about 550 My ago [*Clifford*, 1970]. Within Africa there are three regions where Lg coda Q is higher than in the rest of the continent, being 800 or more. These are the West African Craton in western Africa, the center of the East Sahara Craton in northeastern Africa, and the northern portion of the Kalahari craton in southern Africa. These three cratons, along with the Congo craton, just to the west of the southern part of the East African rift system, coalesced to form the African continent in a vast tectono-thermal event which ceased about 550 My ago [*Clifford*, 1970]. Q values obtained for the Congo craton are lower, being about 600. These lower values may be real and may occur because of the craton's proximity to the East African Rift System and/or to Paleozoic folding which occurred in the Congo fold belt near the eastern edge of the craton. It is also possible that the limited resolution of the regionalization method prevents us from delineating a small region of higher Q values which may lie there.

Figure 9 plots the frequency dependence exponents (ζ) of Lg coda Q at a frequency of 1 Hz. The values in western Africa have recently been redetermined and those in the northwestern part of that continent differ from those obtained by *Xie and Mitchell* [1990a]. The reason for the redetermination is that instruments in that region are part of the Ivory Coast network and differ in their responses from the WWSSN instruments used throughout the rest of Africa for coda Q determinations. The redetermination leads to new values of frequency dependence which provide a better fit to the slopes of the stacked spectral ratios in that region. Note that low frequency dependence values correlate with high values of Q.

With the exception of the Congo craton, Lg coda Q values within the major cratons of Africa are higher than values associated with the fold belts between

them. This suggests that the variations of Lg coda Q values in the stable portions of Africa reflect tectonic activity associated with the coalescence of those cratons and the formation of the African continent and that Lg coda studies can be used to study the episodes in the tectonic evolution of continents at least as early as the early Paleozoic era.

Although, as indicated above, Lg coda Q values, as well as Lg Q values are usually high in stable cratons, there are at least three exceptions to that rule. They are the Arabian peninsula where Lg coda Q values range between 157 and 300 [*Ghalib*, 1992], central Australia where Lg values are about 230 [*Bowman and Kennett*, 1991], and the Siberian craton where Q values range between about 400 and 600. Low Q values for the Arabian peninsula have also been inferred from the spectra of fundamental-mode Rayleigh waves [*Seber and Mitchell*, 1992]. Those low values correlate with recent tectonic activity which is known to have occurred in the region [*Chazot and Bertrand*, 1993]. The low Q values in the Siberian craton occur in a region which underwent extensive extension and volcanism during the Mesozoic era and those in the central portion of the Australian shield have been attributed to a large positive velocity gradient in the lower crust of that region [*Bowman and Kennett*, 1991] which reduces Lg Q by increasing the effect of geometrical spreading. There is, however, geological evidence for tectonic activity there during the Devonian and Carboniferous eras [*Plumb*, 1979], so it is possible that reduction of Q values in the upper crust and enhanced geometrical spreading both contribute to reductions in Q in that region.

Our Lg coda Q map of Eurasia (Figure 10) shows low values which stretch from southern Europe to southeast Asia through regions of current or recent tectonic activity. High values (~700) occur through a broad region from Scandinavia to the Urals which roughly corresponds to the East Eruopean craton. In contrast, the Siberian craton which we might expect to also have high values has low values (~400) throughout much of its extent. The lowest values within that craton correspond in position with a zone of extensive tectonism which occurred during the Mesozoic era. Frequency dependence values (Figure 11) show no consistent relation to coda Q_0.

Inversion of Fundamental-mode Attenuation data

The methodology for inverting surface-wave attenuation for anelastic properties was developed three decades ago [*Anderson and Archambeau*, 1964; *Anderson et al.*, 1965] and was used to obtain upper mantle models of shear-wave Q (Q_μ) assuming that the Earth was radially symmetric and that Q is >>1 and independent of frequency. A more complete treatment simultaneously inverts for Q and velocity as functions of depth [*Lee and Solomon*, 1978], an approach which takes full account of the dependence of surface wave attenuation and velocity on both the elastic and dissipitative parts of Earth structure. Although that procedure improves the uncertainties in resulting Q_μ models, it produces models which are essentially the same as those produced by methods which solve only for Q. Since we are only interested in broad-scale features of the distribution of Q_μ with depth all the models which we present are derived using the formulation of *Anderson et al.* [1965]. The equations have been modified to permit study of the frequency dependence of Q assuming that that dependence is constant with depth [*Mitchell*, 1980], and more recently to include the possibility that frequency dependence can vary with depth [*Mitchell and Xie*, 1994]. In this last case, shear wave Q in layer l ($Q_{\mu l}$) can be expressed as $Q_{0l} f^{\zeta_l}$, where the subscript l is a layer index, Q_{0l} is the Q_μ value at 1 Hz for layer l, and the exponent ζ denotes the degree of frequency dependence. Note that ζ is an intrinsic frequency dependence, whereas η,

defined earlier as the frequency dependence of Q_{Lg}, is not an intrinsic property of the crust, but can be affected by both velocity and Q_μ structure [*Mitchell*, 1991]. The Rayleigh wave attenuation coefficient, γ_R, in this formulation is

$$\gamma_R = \frac{\pi}{C_R^2 T}\sum_{i=1}^{N}\left[\left(\beta_i \frac{\partial C_R}{\partial \beta_i}\right)_{\omega\rho\alpha} + \frac{1}{2}\left(\alpha_i \frac{\partial C_R}{\partial \alpha_i}\right)_{\omega\rho\beta}\right] Q_{\mu_i}^{-1} \qquad (5)$$

where the subscripts R, α, and β identify Rayleigh, compressional, and shear waves, respectively, ω is angular frequency, and C_R is Rayleigh wave phase velocity. The subscripts, ω, α, and β indicate that those quantities are held constant when the partial derivatives are computed. The factor 1/2 occurs in the right-hand term because we assume that compressional wave Q is twice as large as Q_μ for the inversions. Previous studies, such as *Mitchell* [1980], indicate that inversion results are affected only slightly by reasonable assumed values of compressional wave Q. When ζ_l is zero at all depths, the inversion yields a frequency-independent model.

In order to study the frequency dependence of Q_μ we must separate it from the effect of depth dependence. This is not possible using only fundamental mode data, but is feasible if higher mode attenuation data are available [*Mitchell*, 1980]. Figure 12 shows that higher-mode surface waves at a given period sample a larger depth interval in the crust than does the fundamental mode. Eigenfunctions of vertical displacement in that figure were computed for a model of the eastern United States which is identical to that of *Mitchell and Herrmann* [1979] except that it does not include a sedimentary layer. They show that the amplitude of the fundamental mode at a period of 1 s is insignificant below depths of 4 or 5 km. If a sedimentary layer had been included, significant amplitudes would have been restricted to the upper 2 or 3 km of the crust. The first higher mode at 1 s, however, samples about the same depth interval as that sampled by the fundamental mode at a period of 5 s (about 20 km) and the fourth higher mode samples somewhat deeper. Adequate synthesis of the Lg phase to periods as short as 1 s by mode summation requires at least 20 to 30 additional higher modes, all of which sample the crust between the surface and depths of 30 km or more [*Mitchell and Xie*, 1994]. The first higher mode at a period of 5 s samples depths similar to those sampled by the fundamental mode at 20 s (50-60 km).

Mitchell and Xie [1994] used (5) in a combined formal inversion and forward modeling procedure to obtain frequency-dependent Q_μ models for the Basin and Range province. The procedure begins by assuming a depth distribution for ζ_1 and inverts fundamental-mode attenuation coefficient data to obtain a Q_μ^{-1}. If the resulting model is acceptable (i.e. it predicts attenuation coefficient values which agree with observed values and does not fluctuate wildly) we determine the values of Lg Q and its frequency dependence at 1 Hz and compare them with observed values. Mitchell and Xie computed a set of synthetic seismograms at several distances and used the stacked spectral ratio method of *Xie and Nuttli* [1988] to obtain those values. This process can be repeated for various depth distributions of ζ. Because of the large uncertainties in attenuation and Q data as well as the non-uniqueness inherent in Q inversions, there will be many models which will satisfy the fundamental-mode and Lg data. As found by Mitchell and Xie in the Basin and Range, however, there may be features of the models which are independent of the depth distribution assumed for ζ.

Figure 13 shows a model obtained by inverting fundamental-mode Rayleigh-wave attenuation data, obtained in the Basin and Range, for the case when ζ = 0.0. Low Q_μ values ($\approx$ 50) characterize the upper crust, high values ($\approx$

1000) occur at mid-crustal depths, and values decrease gradually at greater depths. The resolving kernels which are associated with the model indicate that resolution is best in the upper crust and degrades with increasing depth. The value obtained for Q_μ at any depth is an average which is smeared out over the depth interval encompassed by the resolving kernel. Those kernels indicate that only gross features of the models can be resolved and that any changes which occur over small depth ranges have no real meaning. The standard deviation bars in Figure 13 are small enough so that we can attribute significance to the major features of the model. A preset parameter allows us to trade resolution against standard deviation values in the inversion process. Every inversion involves several trials to find a value for that parameter so that we avoid models with negative Q_μ^{-1} values or values which fluctuate wildly from layer to layer. The resolution kernels and standard deviation values for the model in Figure 13 are fairly typical and can be taken to be roughly representative of the models shown in ensuing figures.

Upper Mantle Models

As mentioned earlier, the first shear wave Q (Q_μ) models of the mantle assumed that the anelastic properties of the Earth were radially symmetric. Since, however, most mechanisms for explaining Q are thermally activated, and since it is known that temperatures in the Earth may vary drastically from region to region [*Pollack et al.*, 1993], it is likely that there will be regional variations of Q_μ. The variations of the Rayleigh wave attenuation coefficient values for various regions of the world shown in Figures 5, 6, and 7 also suggest that, in order to be adequately explained, Q_μ must vary from place in both the crust and upper mantle.

Models, obtained from the inversion of surface wave attenuation data, which primarily bear on the regional variation of upper mantle Q_μ over broad continental regions have been obtained for several regions. These include eastern (ENA) and western (WNA) North America [*Solomon*, 1972; *Lee and Solomon*, 1978; *Chen*, 1985; *Hwang and Mitchell*, 1986], as well as eastern (ESA) and western (WSA) South America, the Indian shield (IS), and the Himalaya (HI) [*Hwang and Mitchell*, 1986]. In addition, regional variations of Q_μ have been found to occur in the upper mantle of the western Cordillera of North America [*Al-Khatib and Mitchell*, 1991]. Figure 14 summarizes models obtained in the last ten years for both stable and tectonically active continental regions. The broader-scale models (solid lines) indicate that Q_μ decreases with increasing depth for the entire interval between 50 and 200 km. The models for stable regions (ESA and ENA) have higher Q_μ values at all depths than do models for tectonically active regions (WSA, HI, and WNA). At depths between 100 and 200 km *Hwang and Mitchell* [1986] found Q_μ values in the stable regions to be 125-150 and those in tectonically active regions in the same depth range to be 40-70.

Al-Khatib and Mitchell [1991] divided the western United States into three regions which they called the eastern Rockies which was largely confined to the Rocky Mountains, the Intermountain region which stretches from the western margin of the Rocky Mountains westward through the Basin and Range, and the Western Margin which consists mostly of California and parts of western Oregon. With this finer regionalization they found the Q_μ values in the uppermost mantle of all three regions to be lower than those obtained for tectonically active regions by *Hwang and Mitchell* [1986]. At depths greater than about 125 km, values for the eastern Rockies are similar to those in tectonically active regions but lower values occur in the Intermountain region (20-40) and for the Western Margin region (15-50). These comparisons indicate that Q_μ in the upper mantle varies tremendously from region to

region and that models derived from data acquired for long paths over broad regions can lead to models which differ substantially from models derived from data obtained over smaller areas within that broad region. The gross consistency of the models and of the patterns of the Q_μ distributions indicate that models obtain from surface wave attenuation data reflect real changes of shear wave internal friction in the upper mantle. As indicated in an earlier section, however, amplitude data are subject to systematic errors which often cannot be taken into account; conclusions regarding the distribution of Q_μ in the Earth should be based only the gross features of models and fine-scale features should be ignored.

The differences in Q_μ at upper mantle depths between stable regions and tectonically active regions affect Q values and t^* (travel-time/Q) values observed for teleseismic body waves both at long periods [*Solomon and Toksöz*, 1970] and short periods [*Der et al.*, 1982]. Q determinations of those waves consistently find that Q values in the upper mantle beneath the western United States must be lower than beneath the central and eastern United States.

Currently available methods and data do not allow any inferences to be made concerning the frequency dependence of Q_μ in the mantle from surface waves. Body-wave studies, however, indicate that Q in the upper mantle may be frequency dependent, at least in some regions. *Der* [1982] has summarized available information on that topic.

Crustal Models

Studies of crustal Q_μ require data at shorter periods than those used for mantle studies. Those waves are therefore more likely to be affected by lateral variations in elastic and anelastic structure than are the longer period waves used to study the mantle. It is advantageous therefore to obtain data over regions which are as laterally uniform as possible and to make multiple observations within each region. The number of such regions is still very limited.

Although there are difficulties in determining Q_μ at shallow depths in the crust, it is important to do so, because observations of the attenuation of Lg can be combined with fundamental-mode surface-wave attenuation to study the frequency dependence of Q at crustal depths. The displacement eigenfunctions in Figure 12 show that higher mode surface waves (which comprise Lg) at a period of 1 s sample roughly the same depth range in the crust as do fundamental-mode surface waves in the period range 5-20 s. For that reason, 1 s Lg attenuation data can be combined with that obtained for fundamental surface waves to study the frequency dependence of Q_μ in the crust. Short-period stations of the WWSSN network (with a peak response at 1 s) began operating in the early 1960's and many recordings of the Lg phase have been amassed since then. There is therefore a large data base for studies of 1 s Lg waves.

Regions where both fundamental-mode and Lg data have been used to obtain crustal models of Q_μ are still very limited and I will restrict discussion to models obtained for the eastern United States between the Great Plains and Appalachians [*Cong and Mitchell*, 1988] and for the Basin and Range province [*Mitchell and Xie*, 1994]. Fundamental mode surface waves used for recent attenuation determinations in both regions can be considered to be relatively free of focusing effects produced by lateral variations in structure. I infer that that is the case for the central United States because of its relatively uniform velocity structure and for the Basin and Range because of careful determinations of surface wave particle motion for the data used in that study [*Mitchell, et al.*, 1993].

Figure 15 shows models obtained for both regions. The eastern United States model requires that Q_μ vary with frequency, at least in the range of about 0.3-1.0 Hz, in order to explain both fundamental-mode and Lg attenuation data [*Mitchell*, 1980; *Cong and Mitchell*, 1988]. It could also be frequency dependent at longer periods, but that is not required by the data [*Mitchell*, 1980]. The EUS model in Figure 15 is plotted for the case where $\zeta = 0.5$. That number is not well determined and could vary with depth as well as frequency.

Crustal Q_μ models for the Basin and Range are also highly non-unique. The model in Figure 15 assumes that $\zeta = 0.0$ at depths between the surface and 15 km and that $\zeta = 0.5$ at greater depths. *Mitchell and Xie* [1994] derived several other models which also satisfactorily explained fundamental-mode Rayleigh wave and Lg attenuation data (Lg Q and its frequency dependence 1 s). Common features of all of those models are low Q_μ values in the upper crust, a rapid increase in value at mid-crustal depth, and a low average frequency dependence throughout the crust. If it is assumed that ζ is uniform with depth, a value of 0.1 explains all of the data well and a value of 0.0 (a frequency-independent model) explains the data within its uncertainties. *Mitchell* [1991] had shown earlier that it was possible to explain the frequency dependence of Lg Q (about 0.5 in the Basin and Range) by frequency--independent models in which Q_μ increased rapidly from low to high values at mid-crustal depths. Several frequency-independent models had previously been derived for the Basin and Range [*Cheng and Mitchell*, 1981; *Patton and Taylor*, 1984; *Lin*, 1989] which were not constrained by Lg Q information.

A rapid increase in Q_μ at mid-crustal depths was detected in an inversion of Rayleigh wave attenuation in the central United States [*Mitchell*, 1973]. The same feature has since been found to occur at similar depths beneath the western United States [*Mitchell*, 1975; *Mitchell and Xie*, 1994] and has been a feature of all models derived from fundamental-mode surface wave attenuation when sufficient data at short periods are available. Observed increases of Q with lapse time in coda waves [*Roecker et al.*, 1982; *Pulli*, 1984; *Phillips and Aki*, 1986; *Del Pezzo et al.*, 1990], as well as increases of P-wave Q values with distance from the source [*Al-Shukri, et al.*, 1988] also indicate an increase of Q within the crust. The increase in Q with depth based on coda observations has usually been inferred to indicate an increase in scattering Q at greater depths. *Wennerberg* [1993], however, argues that an increase in intrinsic Q with depth is a more likely explanation. *Mitchell* [1991] found that the large frequency dependences of Lg Q and Lg coda Q found at 1 Hz in the Basin and Range province can readily be explained by a severe increase in Q_μ at mid-crustal depths without the need for frequency-dependent Q_μ values in the crust.

Overviews of Lg coda Q studies indicate that Q_0 is high and its frequency dependence is low in most stable regions whereas that Q_0 is low and its frequency dependence is high in most tectonically active regions [*Nuttli*, 1988]. Because of these similarities, it is likely that the types of model obtained for North America which explain Lg coda results there are applicable to much of the world's continental regions. A rapid increase in Q_μ at mid-crustal depths is therefore likely to be a prevalent feature of the world's continents.

A Peak in Q^{-1} at 0.5 Hz?

Aki [1980] speculated, but did not consider it firmly established, that there may be a peak at a frequency of about 0.5 Hz in Q^{-1} values for the lithosphere. Available crustal values of Q_μ^{-1} make it possible to further investigate that possibility. Figure 16 presents values of Q_μ^{-1} obtained and inferred for the upper crust of stable and tectonically active regions over the frequency range from 0.05 to greater than 1 Hz and compares them with Q values measured for fundamental-mode surface waves and for Lg. For the values in stable regions

it is assumed that the upper crust takes on values obtained from frequency-dependent inversions of fundamental-mode Rayleigh wave data at frequencies less than about 0.6 Hz and has value of about 0.001 at 1 Hz as inferred from frequency-dependent inversions [*Mitchell*, 1980; *Cong and Mitchell*, 1988]. Q_μ^{-1} in the upper crust of tectonically active regions is taken to be independent of frequency [*Mitchell*, 1981; *Mitchell and Xie*, 1994] at frequencies to just above 1 Hz. Studies of shear-wave Q at higher frequencies [*Hough and Anderson*, 1988; *Leary and Abercrombie*, 1994] in California, however, suggest that Q_μ may become frequency dependent at some point, as indicated by the dashed line in Figure 16. The vertical bars in the figure indicate ranges for Q_R^{-1} (at 0.05 and 0.2 Hz) and for Q_{Lg}^{-1} (1 Hz) for stable (narrow bars) and tectonically active (wide bars) regions. Q_μ^{-1} is very low in the lower crust (0.002 or less at 1 Hz) and those low values have a strong effect on measured Q_R^{-1} and Q_{Lg}^{-1} values. For that reason, at a frequency of 0.05 Hz Q_R^{-1} in stable regions is about 0.0025 (Q_R = 400) and in tectonically active regions is centered near 0.005 (Q_R = 200) even though Q_μ^{-1} in the upper crust is about 0.0033 (Q_μ = 300) in stable regions and 0.017 (Q_μ = 60) in tectonically active regions. If we were to use Q_R^{-1} instead of Q_μ^{-1} to infer the frequency dependence of Q in the lithosphere we would infer a peak in Q^{-1}; such a peak is, however, not necessary if we use Q_μ^{-1} and remember that those values vary tremendously from region to region.

Correlation of Upper Crustal Q_μ with Lg Coda Q

Available Lg coda Q and upper crustal Q_μ information indicate that they both vary regionally in the same manner; they are high in stable cratons and low in regions where tectonic activity is occurring or has recently occurred. This is clear from the following comparisons which refer to frequencies of 1 Hz for both Q_μ and Q_{Lg}; the same conclusions, however, also hold for fundamental-mode results at lower frequencies. In stable portions of eastern North America, both upper crustal Q_μ [*Mitchell*, 1980] and Q_{Lg} [*Singh and Herrmann*, 1983] have been found to be high (~1000 and 800-1200, respectively), whereas both Q_μ values [*Cheng and Mitchell*, 1981; *Mitchell and Xie*, 1994] and Lg coda Q values [*Singh and Herrmann*, 1983; *Chávez and Priestley*, 1986; *Nuttli*, 1986] in the Basin and Range are much lower (50-80 and 140-400, respectively). Similarly, Q_μ values in stable and active portions of South America [*Hwang and Mitchell*, 1986] correlate with observed values of Lg coda Q [*Raoof and Nuttli*, 1985] as do Q_μ values in India with measured Lg coda Q [*John*, 1988]. Those Q_μ values are inferred from the attenuation of fundamental-mode Rayleigh waves obtained over the period range of about 5 to between 30 and 100 seconds. Wavelengths for these waves are roughly between 10 and 300 km as compared to wavelengths of Lg coda waves which may range between about 1.5 and 3.0 km. The similarity of the patterns of Lg coda Q variation with those of Q_μ variation therefore suggests that both Lg coda and fundamental-mode Rayleigh waves are predominantly being affected by intrinsic anelastic properties of the Earth.

Local Q Studies

Although coda Q tomography permits the study of regional variations of Q in greater detail than does fundamental-mode surface-wave attenuation, results are still averaged over fairly large areas. Studies of local features require smaller scale studies using body waves. Body-methods have been applied to the study of several small-scale features such fault zones and geothermal source areas [*Jin and Aki*, 1988; *Al-Shukri and Mitchell*, 1990; *Fehler et al.*, 1992; *Mayeda, et al.*, 1992; *Li, et al.*, 1994]. Although body-wave studies are outside the scope of this review, I mention them here because they

provide constraints which any model proposed to satisfy surface wave attenuation must satisfy. Local studies, for instance, can provide information on the distribution and density of regions of locally low or high Q values in the crust. A model which includes consists of adjoining regions of low Q and high Q is discussed in a later section.

An interesting possibility which has been raised on the basis of local Q studies is that temporal variations in Q can be observed which occur over time scales of a few years. Chouet [1979] made the first observations which suggested this possibility and their plausibility has been supported by several subsequent studies [*Del Pezzo et al.*, 1983; *Rhea*, 1984; *Novelo-Casanova et al.*, 1985; *Jin and Aki*, 1986; *Sato*, 1986; *Peng et al.*, 1987]. Although short-term temporal changes in Q have not found wide-spread support, they are not inconsistent with the conclusions expressed in this review, that Q_μ can change drastically, depending on the density of cracks and volume of fluids present in the upper crust. If those parameters can change over geological time, it is also possible that they can change rapidly in tectonically active regions over relatively short periods of time. Increased attenuation, possibly observed prior to some earthquakes could be interpreted as being due to enhanced cracking and increased fluid content in the upper crust prior to an earthquake. That fluid, in turn, could reduce effective stress in the region and lead to a greater possibility of earthquake occurrence.

Effect of Temperature on regional Q_μ Variations

The models shown in Figures 14 and 15 indicate that Q_μ varies by as much as a factor of 10 in the upper mantle at depths reaching 200 km and by factors between 3 and 12 in the upper crust, depending upon the the frequency at which Q is determined. The low values, both in the upper crust and the upper mantle, occur in regions characterized by high heat flow and elevated temperatures at depth. It is pertinent to ask therefore whether variations in temperature alone can explain the differences in Q_μ which have been determined. Laboratory studies on internal friction have revealed a "high temperature background" [*Chang*, 1961; *Gueguen et al.*, 1989] which is often the dominant factor in the measurements. Increasingly higher temperatures are thought to produce progressive unpinning of dislocations and, consequently, lower Q_μ values in upper mantle rock.

Gueguen [1989] described the effect of the "high-temperature background" on their measurements of internal friction by $Q^{-1} \sim \omega^{-\zeta} \exp(-\zeta E/RT)$ where ω is angular frequency, ζ is the frequency dependence parameter discussed earlier, E is activation energy, and T is absolute temperature. They found that E ~ 440 kJ mol^{-1} for olivine single crystals. *Berckhemer et al.* [1982] found values for E of the order of 500 to 700 kJ mol^{-1} for polycrystalline mantle rocks. If temperature is the only factor which causes differences in Q from region to region, the above relation can be used to write an expression for the ratio of Q at a particular depth in two regions i.e.

$$\frac{Q_{\mu_2}}{Q_{\mu_1}} = \omega^{\zeta_2 - \zeta_1} \exp\left[\frac{E}{R}\left(\frac{\zeta_2 T_1 - \zeta_1 T_2}{T_1 T_2}\right)\right] \qquad (6)$$

We can solve this equation for E if it is assumed to be the same in both regions for a particular depth of interest and if the Q_μ, ζ, and T values are known for those regions. At mantle depths, ζ is not well known; consequently, determinations of E using equation 6 should be made for ranges of possible values. Using the continental temperature values of *Pollack et al.* [1993] at a depth of 150 km which would be appropriate for the central United States and

for the Basin and Range province, there are several combinations of ζ values which predict E values within the range measured by *Berckhemer et al.* [1982] for upper mantle rocks. If the solidus occurs at temperatures appropriate for a volatile-free mantle [*Pollack, et al.*, 1993], then E values in the 500-700 range, as measured by *Berckhemer et al.* [1982], will result if ζ values are in the range 0.2 to 0.3. Values in this range have been estimated as appropriate averages for the Earth from observations over a broad frequency range [*Anderson and Minster*, 1979]. If the solidus is appropriate for a mixed volatile upper mantle, then ζ values in the range 0.4 to 0.6 predict E values in the proper range. This range of frequency dependence values is also reasonable and has been suggested as being appropriate for the lower crust and upper mantle in the Basin and Range [*Mitchell and Xie*, 1994]. Although these calculations do not rule out factors other than temperature as the mechanism for causing regional variations in upper mantle Q, they do show that it is at least plausible that temperature alone could produce the variations inferred to occur at depths of 150 km and greater.

The same calculation was performed for depths in the upper crust at a depth of 10 km. At that depth ζ was taken to be 0.5 in the eastern United States and 0.0 in the Basin and Range [*Mitchell*, 1991]. Q_μ values at the same depth were taken as 800 and 70, respectively (Figure 15). For those values and appropriate temperatures for the eastern United States and Basin and Range [*Pollack, et al.*, 1993], the activation energy, E, is calculated to be about 10 kJ/mole, a value which is much too low to be realistic if temperature in the only factor which contributes to the difference in Q_μ between those regions. We must therefore appeal to another mechanism to explain regional Q differences in the upper crust. The one which appears to be most viable is the motion of fluids in cracks [*Mitchell*, 1975; *Mitchell and Xie*, 1994] and will be discussed in the following section.

Over the depth range from the lower crust through the upper mantle to about 100 km, it is also difficult to obtain reasonable values of activation energy using published values for temperature, Q, and frequency dependence. The only way that it is possible to obtain reasonable values is when both ζ_1 and ζ_2 are less than 0.1. Observed frequency dependence values for teleseismic body waves [*Der, et al.*, 1982] seem to preclude that possibility. Another possible reason for that difficulty may be that whereas Q_μ at a depth of 150 km can be explained by a single atomic attenuation mechanism with a corresponding value of E and may vary as $e^{-E/RT}$, that other mechanisms with different values of E may become important at shallower depths and Q may vary as $e^{E_1/RT} + e^{E_2/RT} + \ldots$

A Model for Regional Q_μ Variations

Several studies have emphasized the role of interstitial fluids in producing regional variations of shear wave internal friction in the upper crust of continental regions [*Mitchell*, 1975; *Mitchell*, 1980; *Mitchell and Xie*, 1994]. A model in which the motion of fluids in a network of cracks causes reduced values of Q_μ is consistent with both laboratory measurements [*Gordon and Davis*, 1968; *Winkler and Nur*, 1979] and with theoretical predictions [*O'Connell and Budiansky*, 1977]. Although limited resolution prevents a precise statement regarding the depth that fluids in the crust penetrate, our models indicate that that depth is in the range 10 to 15 km. Higher Q_μ values at greater depths imply that fluids are absent or, at least that they are much less abundant, at those depths which correspond to the plastic lower crust. Petrological evidence [*Yardley*, 1986] and fluid transport calculations [*Bailey*, 1990] argue against the presence of fluids in the lower crust. The high Q_μ values which are required there to explain observed surface wave attenuation appear to support those views.

Studies of both fundamental-mode surface wave attenuation and Lg coda Q indicate that Q_μ is lowest is regions which have most recently undergone tectonic deformation. This suggests that faults, cracks, and interstitial fluids which produce reduced Q_μ values, are abundant in those regions at the time of, and shortly following, periods of crustal deformation. The fluids may be due to metamorphic reactions which occur as result of elevated temperatures from frictional heating or internal deformation [*Edheridge et al.*, 1984; *Newton*, 1989]. After the cessation of deformation, fluids will be slowly lost, either by retrograde metamorphism or by migration to the surface, cracks will close, and Q_μ will increase in the upper crust. Figure 17 schematically illustrates that process. Results from the continent of Africa (Figure 8) suggest that reductions in observed Lg coda represent the effect of tectonic activity which occurred in the early Paleozoic era.

Surface wave results alone cannot say anything about the distribution of the fractures and cracks which cause the reductions in Q_μ discussed in this review. Permeability determinations in deep wells [*Morrow and Byerlee*, 1992; *Durham and Bonner*, 1993], however, indicate that fluid motion is not uniform in the crust, even in tectonically active regions. Rather, fluid circulation is restricted to major fracture zones. Fluids may occur in clusters of dikes and faults which exist within brittle portions of the crust as proposed by *Hill* [1977] for California. If that is the case, local studies may produce widely disparate values of Q for the upper crust; they will be low in fractured regions and high in regions which lack significant fractures. Surface waves traversing a non-uniformly fractured region will be affected by fluid movement over a broad region and thus yield low values of Q_μ.

Mid-crustal Q_μ values are much higher than those in the upper crust and show much smaller regional variations. At depths greater than about 25 km, however, Q_μ values decrease with depth, slowly in stable regions and more rapidly in tectonically active regions, so that significant regional differences in Q_μ again occur. Calculations in the previous section, assuming a constant activation energy, indicated that differences in temperature could easily explain regional differences in Q_μ at depths of 150 km. At shallower depths in the mantle and at lower crustal depths, it is possible that Q_μ may be governed by more than a single attenuation mechanism. Increasing temperatures serve to unpin dislocations, a process which decreases Q and produces a transition to viscoelastic behavior [*Karato and Spetzler*, 1990]. Partial melt will also decrease Q_μ [*Shankland et al.*, 1981] but experimental results show that the onset of decreasing Q_μ occurs well before temperature reaches liquidus levels [*Sato et al.*, 1989].

Summary and Conclusions

Measurements of surface-wave attenuation coefficients and Lg coda Q values have many pitfalls, but modern processing methods now allow us to look at their regional variations in some detail. Using those processes, and careful screening of data to ensure that it pertains only to the portion of the crust which is of interest, allows us to study, not only the anelastic structure of the crust, but its tectonic evolution as well.

Q_μ and its frequency dependence in the upper of crust of continents varies with geographic region, with observed regional differences of Q_μ being greater than an order of magnitude. The regional differences are related to

the tectonic evolution of the continental crust with regions of recent deformation exhibiting the lowest values of Q_μ and old stable cratons the highest. The regional variations of Q_μ are easily discerned on broadband seismograms, and dramatically affect values of fundamental-mode attenuation coefficient values as well as the Lg phase and its coda. These regional variations of Q_μ in the upper crust are most easily explained by the movement of fluids in cracks which are abundant during and immediately following periods of deformation and which decrease in number with time following the deformation. Fluid volumes continually decrease, by retrograde metamorphism or loss to the surface, producing higher and higher values of Q_μ with time. Tomographic maps of Lg coda Q suggest that we are able to see reduced values of Q caused by tectonic activity which occurred as long ago as the early Paleozoic era.

Circulating fluids are restricted to major fracture zones which traverse the brittle portion of the upper crust; thus local studies of Q will lead to disparate values for the upper crust depending upon where the study was done relative to the location of fractures, whereas surface wave attenuation will sense fluid movement in fractures over a broad region and infer low values for Q_μ in the upper crust.

Q_μ in the lower crust is relatively high, both in stable and in tectonically active regions. This can be explained by the short residence times of fluids in mobile, plastic regions of the crust. The rapid transition from low to high Q_μ values at mid-crustal depths may be a wide-spread feature of continental regions.

At upper mantle depths, it is possible to entirely explain regional variations in Q_μ by different values of temperatures at those depths. Higher temperatures produce a greater degree of unpinning of dislocations, leading to lower Q_μ and eventually to a transition to viscous behavior.

Although much progress in understanding shear-wave internal friction, its regional variation, and frequency dependence has been made in recent years, there are many unanswered questions. Constraints on Q_μ and its frequency dependence in the lower crust and upper mantle are weak and we do not know anything about possible regional variations of frequency dependence at those depths. In addition, our knowledge of Q_μ in the lithosphere of oceanic regions, which comprise the greatest portion of the Earth's surface, is rudimentary. Almost nothing is known of the nature of the frequency dependence of Q_μ there. These and other questions related to crustal and upper mantle anelasticity should be fruitful research areas for years to come.

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Figure Captions

Figure 1. Top - A seismogram recorded by station CCM at Cathedral Cave, Missouri, generated by an earthquake of m_b 4.7 which occurred 183 km away. Bottom - A seismogram recorded by the Apollo 12 seismometer and which was produced by the impact of the S4B Saturn booster of Apollo 14 at distance of 147 km.

Figure 2. Seismograms recorded at station CCM and at PAS (Pasadena, California) from the same event in southeastern New Mexico which occurred on January 2, 1992 at 11 45 35.6 GMT and had a magnitude of 4.6. The epicentral distance to CCM is 1256 km and that to PAS is 1417 km.

Figure 3. The amplitude radiation pattern which would be produced by a vertical strike-slip fault with a strike of about N10°E and schematics of possible paths to two recording stations if the waves are laterally refracted in media with laterally varying elastic properties.

Figure 4. Top Map view of a simple laterally varying model where intrinsic Q is uniformly low to the left of vertical boundary and uniformly high to the right. **x** denotes a source of seismic waves and the numbered circles denote the locations of seismic stations. Bottom Possible amplitude variations with distance if surface travel between two stations (1 to 2 and 3 to 4) or if amplitudes are recorded at two stations, one in the low-Q region (station 1) and one in the high Q region (station 3).

Figure 5. Summary of fundamental-mode Rayleigh-wave attenuation coefficient determinations in stable continental regions at periods between 5 and 90 s.

Figure 6. Summary of fundamental-mode Rayleigh-wave attenuation coefficient determinations in tectonically active regions at periods between 6 and 103 s.

Figure 7. Comparison of fundamental-mode Rayleigh-wave attenuation coefficient determinations for tectonically active and stable continental regions (Figures 5 and 6), with those for oceanic regions [*Canas and Mitchell*, 1978]. The three oceanic values at each period correspond to three different age ranges (0-50 My, 50-100 My, and >100 My) of lithosphere formation.

Figure 8. Tomographic map of Lg coda Q values at 1 Hz for the African continent.

Figure 9. Tomographic map of the frequency dependence (η) of Lg coda Q at 1 Hz for the African continent.

Figure 10. Tomographic map of Lg coda Q values at 1 Hz for Eurasia

Figure 11. Tomographic map of the frequency dependence of Lg coda Q at 1 Hz for Eurasia.

Figure 12. Rayleigh wave eigenfunctions for the fundamental, first, and fourth higher modes at a period of 1 s; for the fundamental and first higher mode at a period of 5 s; and for the fundamental mode at a period of 20 s. All eigenfunctions were computed for a velocity model of the eastern United States [*Mitchell and Herrmann*, 1979].

Figure 13. Left - A Q_μ^{-1} model for a path across the Basin and Range province obtained from an inversion in which it was assumed that Q_μ is independent of frequency (ζ=0) at all depths. The model adequately satisfies both fundamental-mode attenuation coefficient data and observed values for Q_{Lg} (267±56) and its frequency dependency (0.37±0.06) at 1 Hz. Horizontal bars indicate one standard deviation. Right - Normalized resolving kernels determined for depths of 3.0, 9.0, 19.5, and 32.5 km [*Mitchell and Xie*, 1994].

Figure 14. Upper mantle models of Q_μ determined for various stable and tectonically active regions. ESA (eastern South America), ENA (eastern North America), IS (Indian shield), WSA (western South America), HI (the Himalaya), and WNA (western North America) models were determined from the inversions two-station measurements of Rayleigh wave attenuation coefficients on three continents [*Hwang and Mitchell*, 1986] and the ER (eastern Rocky Mountains), IM (Intermountain region), and WM (western margin) models were determined from the same type of data for three regions of the western United States [*Al-Khatib and Mitchell*, 1991].

Figure 15. Frequency-dependent crustal models of Q_μ for the eastern United States [*Cong and Mitchell*, 1988] and the Basin-and-Range province [*Mitchell and Xie*, 1994]. The EUS model assumes a uniform frequency dependence (ζ=0.5) at all depths and the B&R model assumes that Q is independent of frequency to depths of 15 km and that frequency dependence at greater depths is 0.5.

Figure 16. Values of Q_μ^{-1} inferred for the upper crust of stable [*Mitchell*, 1980] and tectonically active [*Mitchell and Xie*, 1994] regions. The vertical bars toward the left of the figure denote observed ranges of

values for Rayleigh wave internal friction (Q_R^{-1}) and those toward the right denote observed ranges of values for Q_{Lg}^{-1}. Wide bars indicate values which were obtained in tectonically active regions and narrow bars indicate those which obtained in stable regions.

Figure 17. Schematic of a fluid-filled crack model which illustrates why intrinsic Q in the upper crust of continents increases with time following periods of tectonic activity. The model on the left represents crust which has recently undergone tectonic deformation and the one on the right represents crust after sufficient time has passed since deformation so that fluids are absorbed or lost and many of the cracks have closed. The shading represents the plastic lower crust.

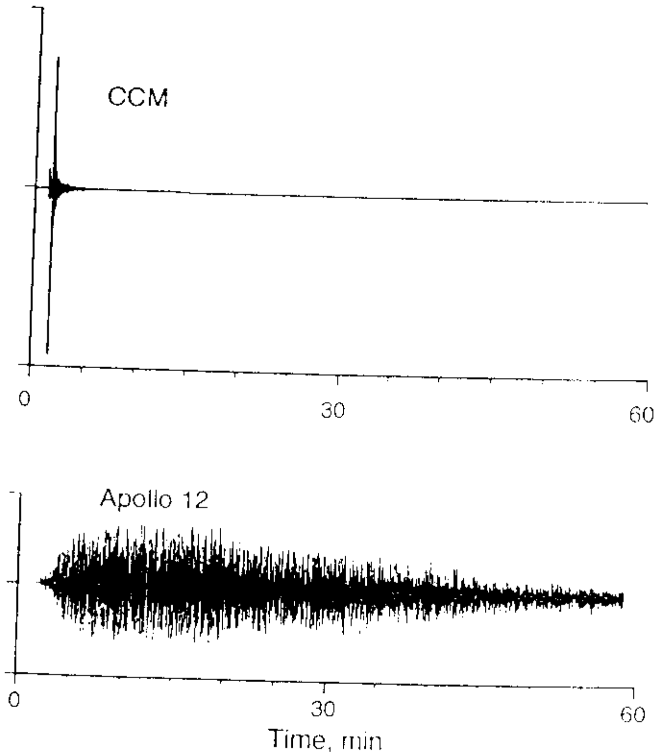
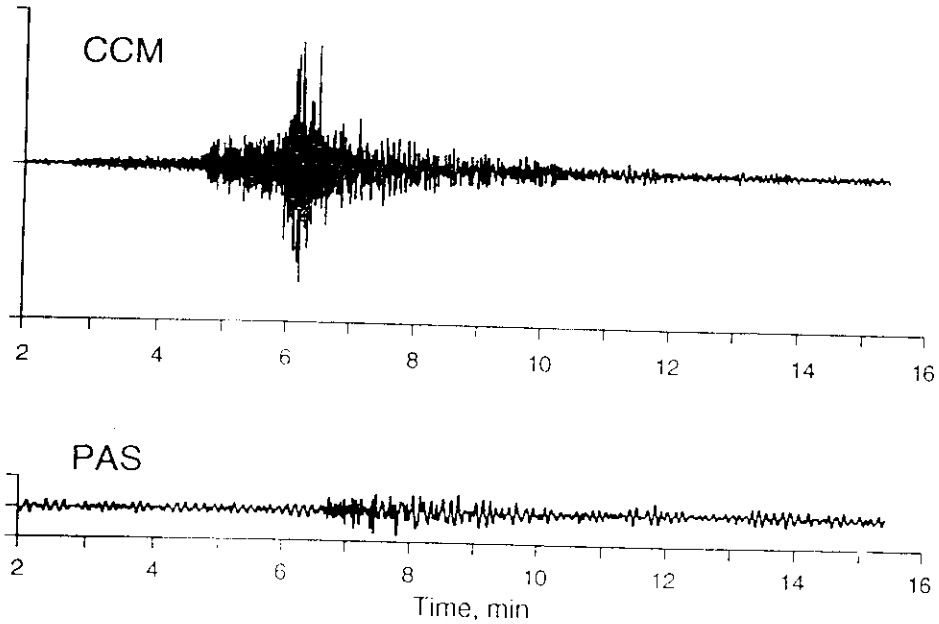


Figure 1



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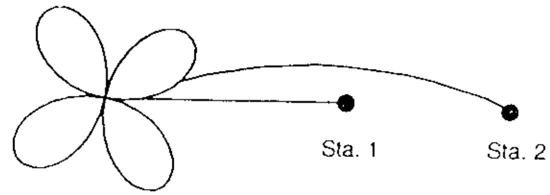

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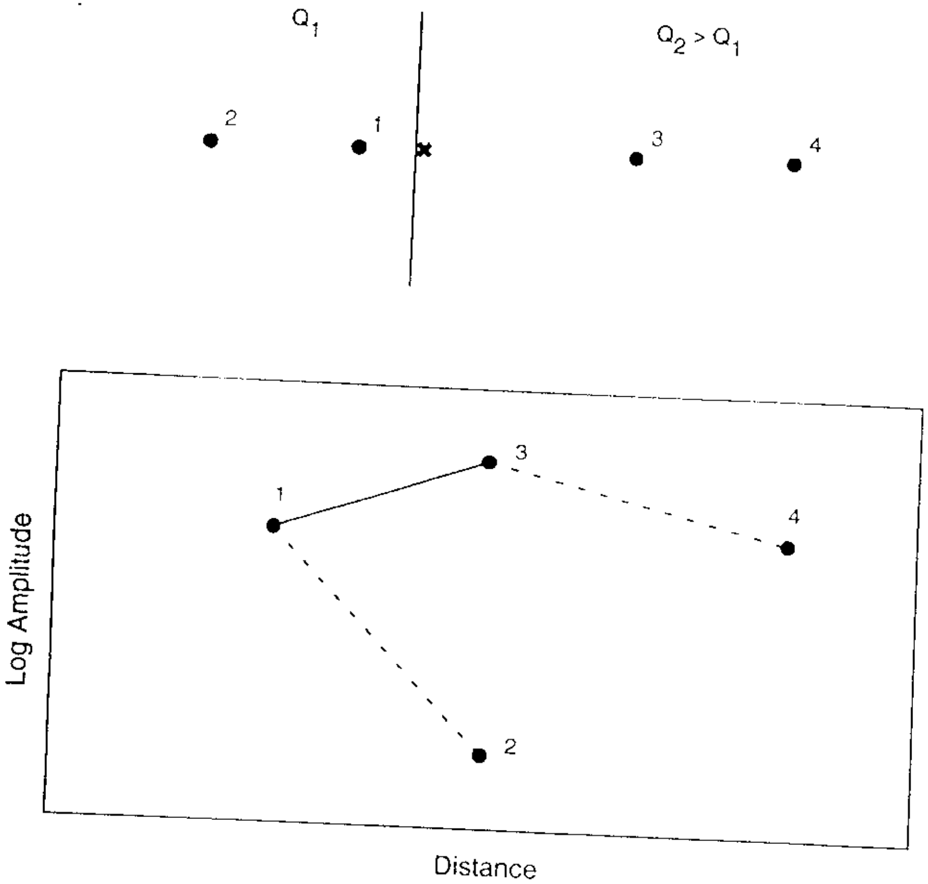

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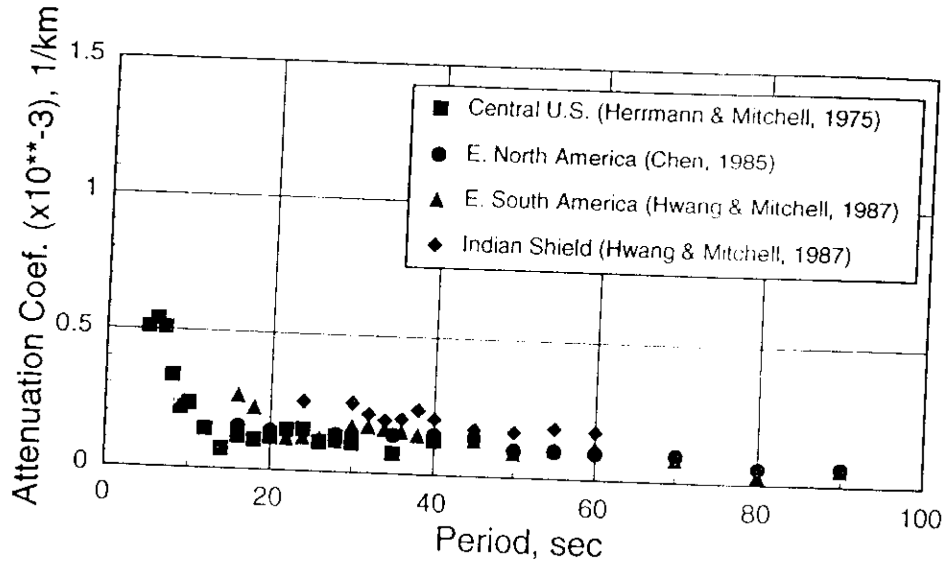

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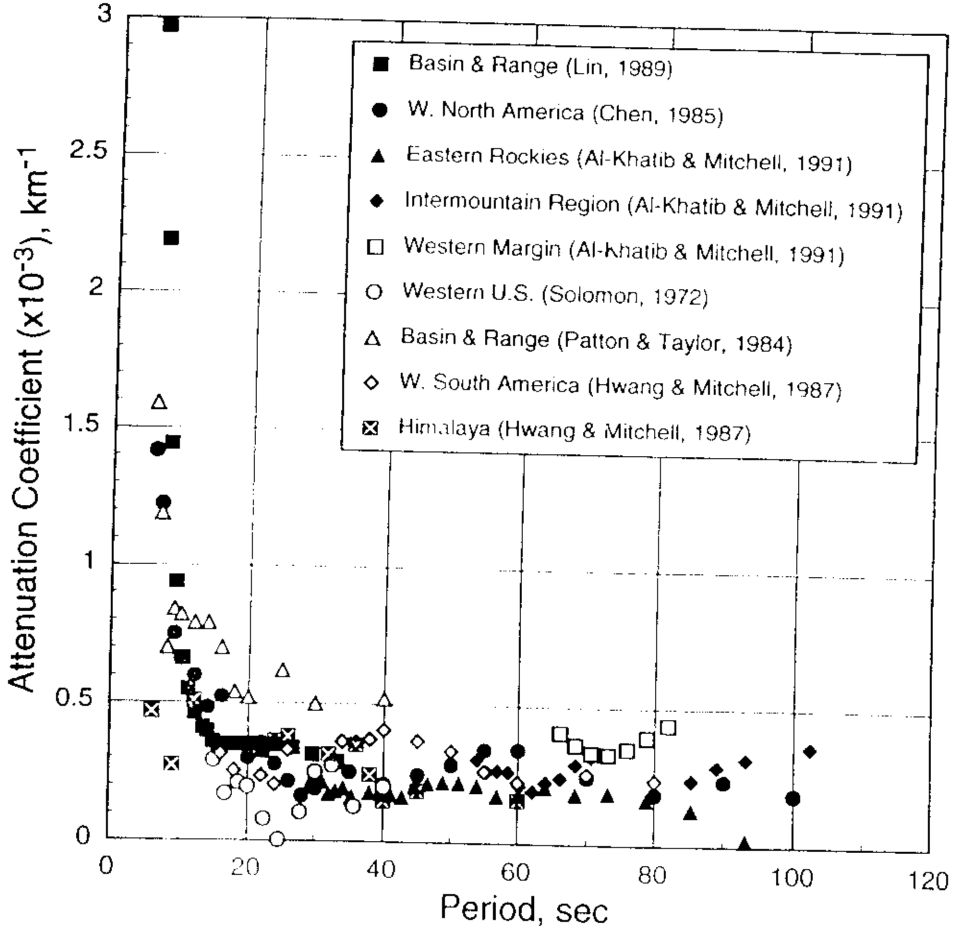

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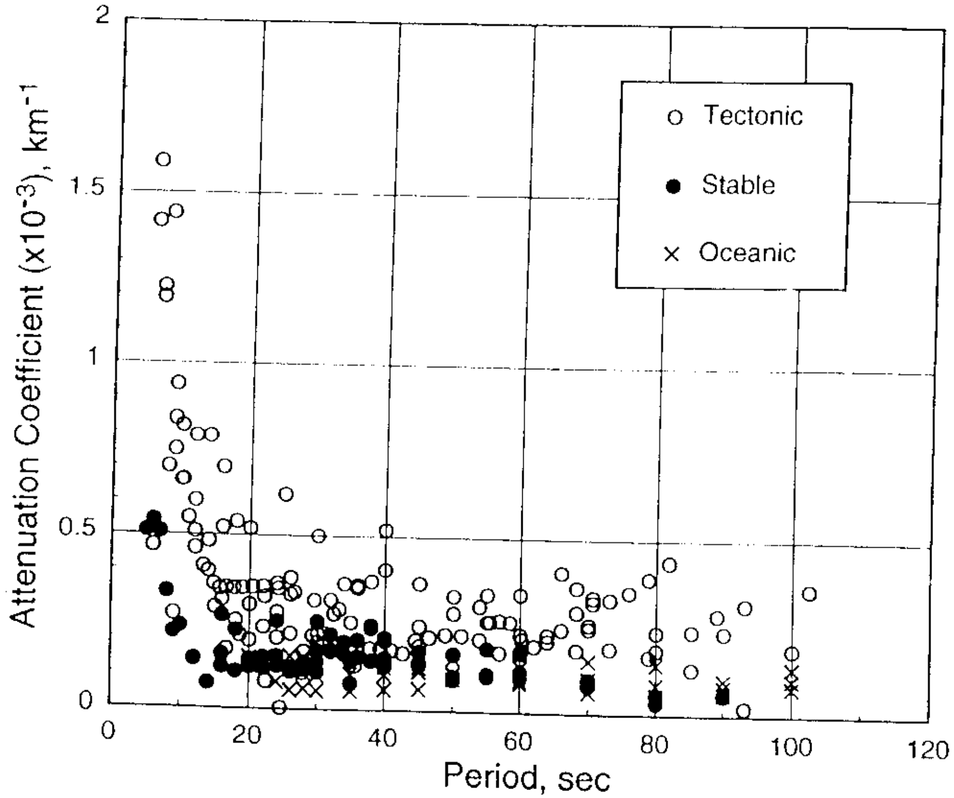

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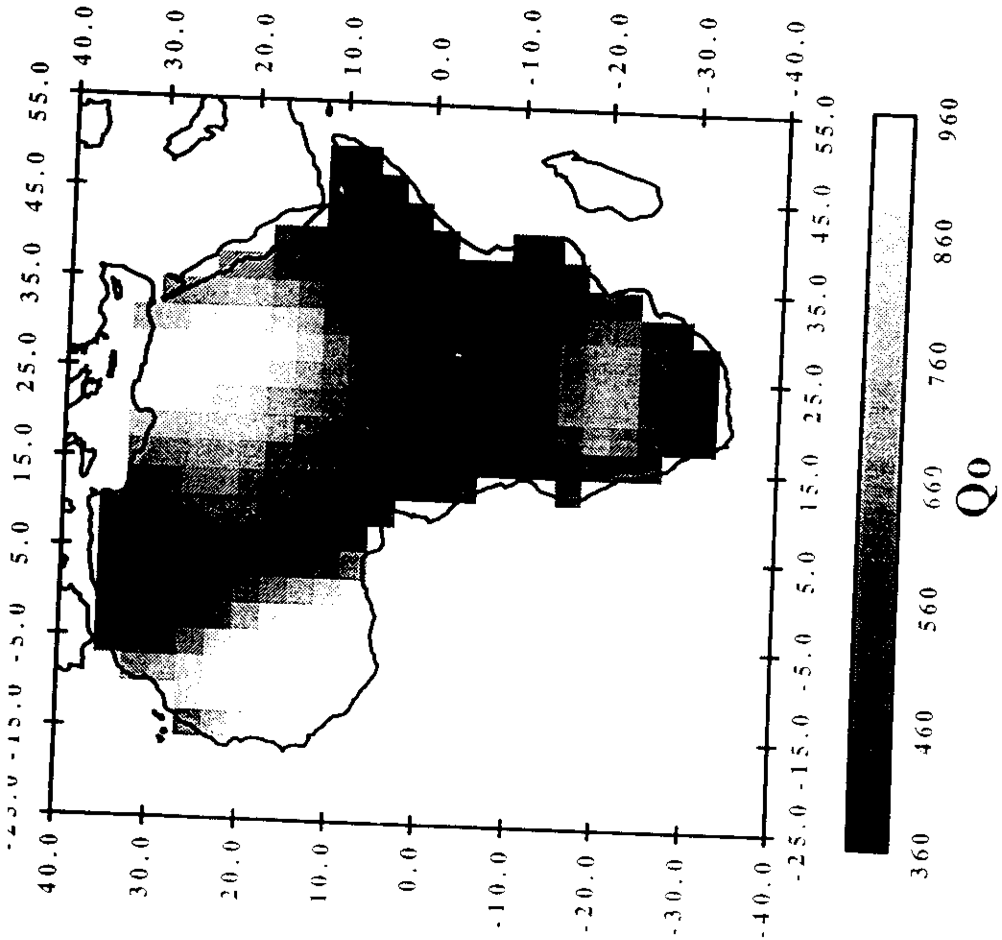

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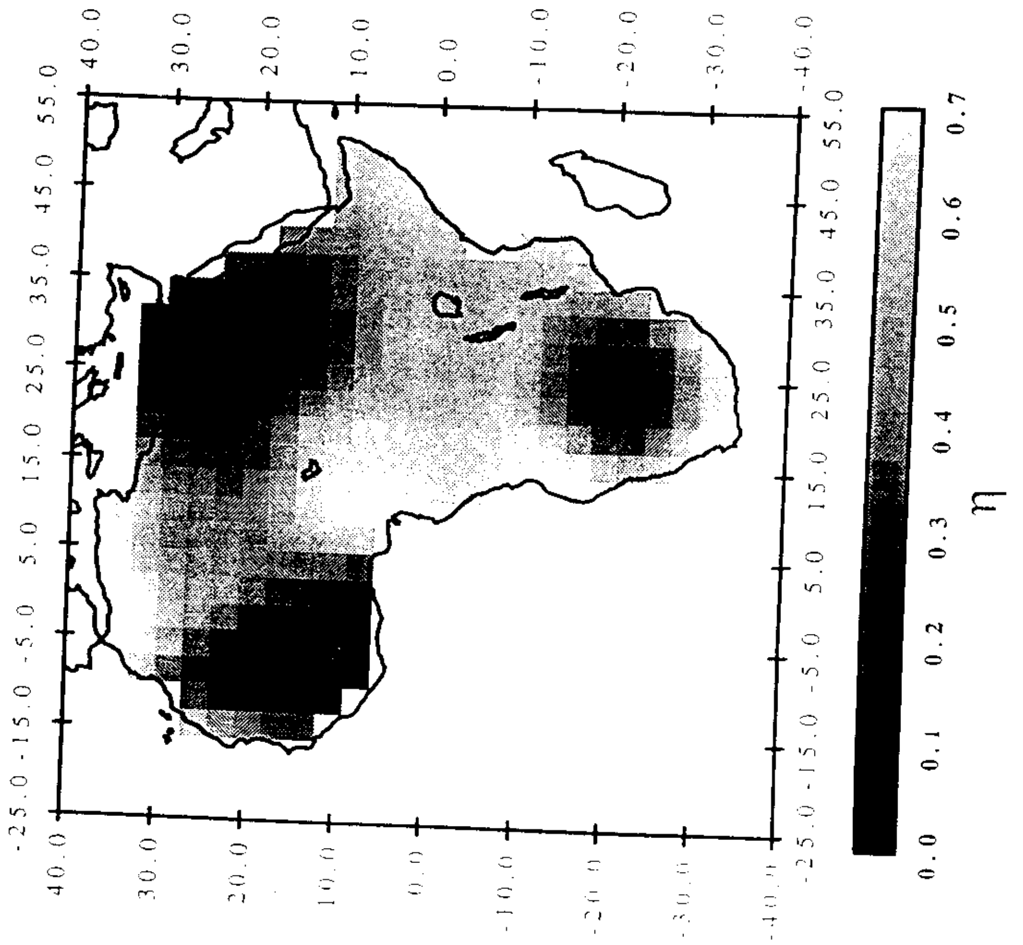

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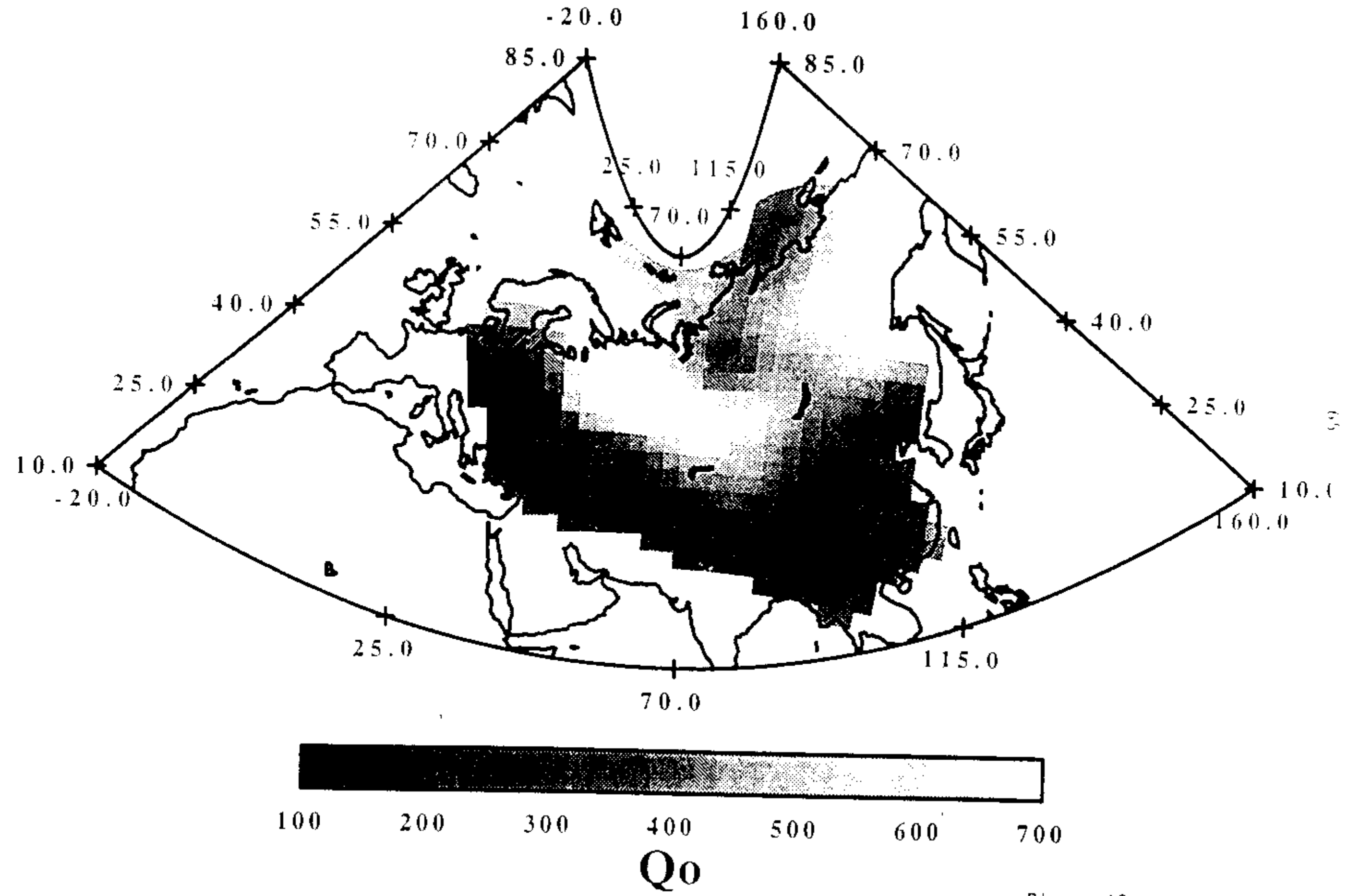

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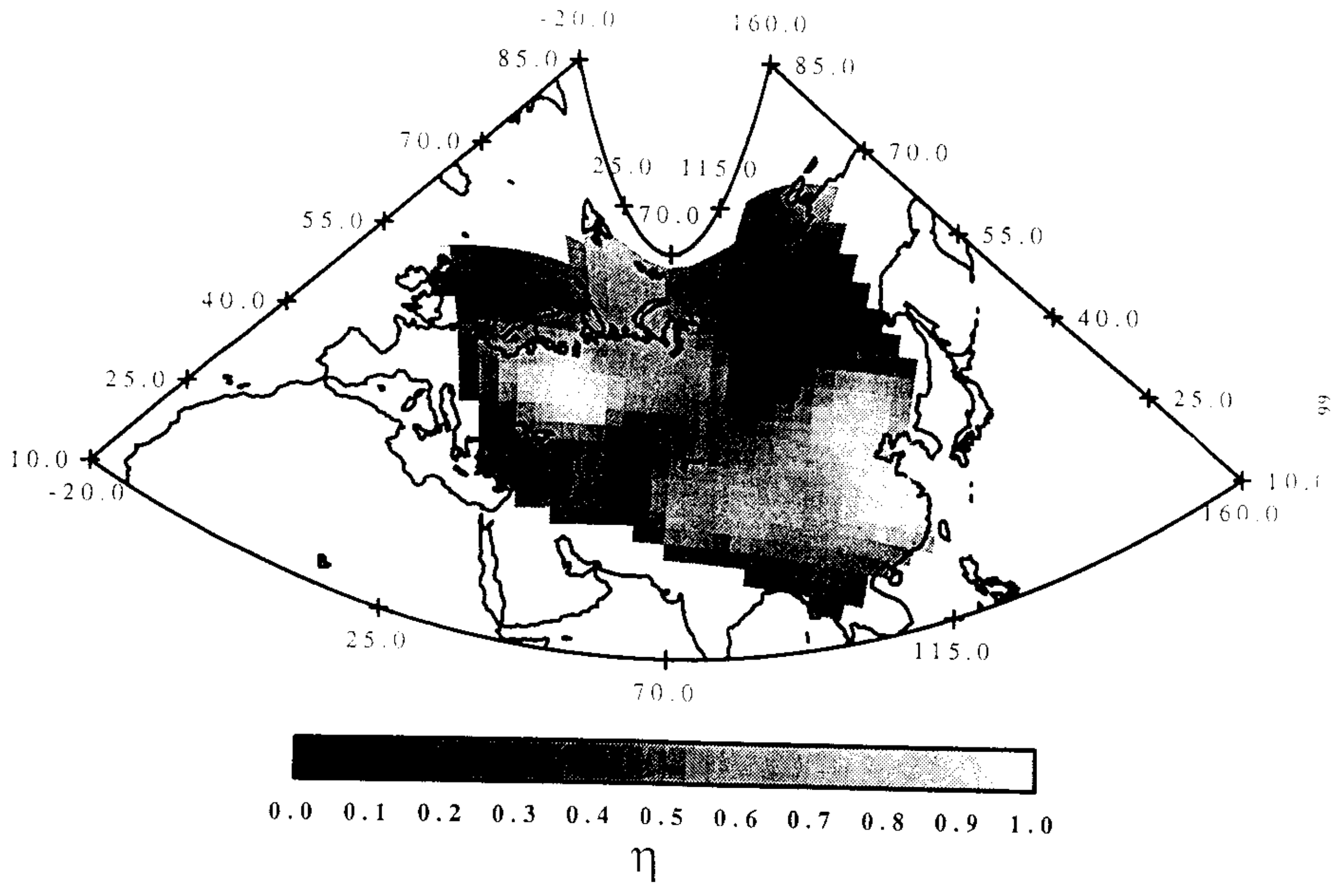

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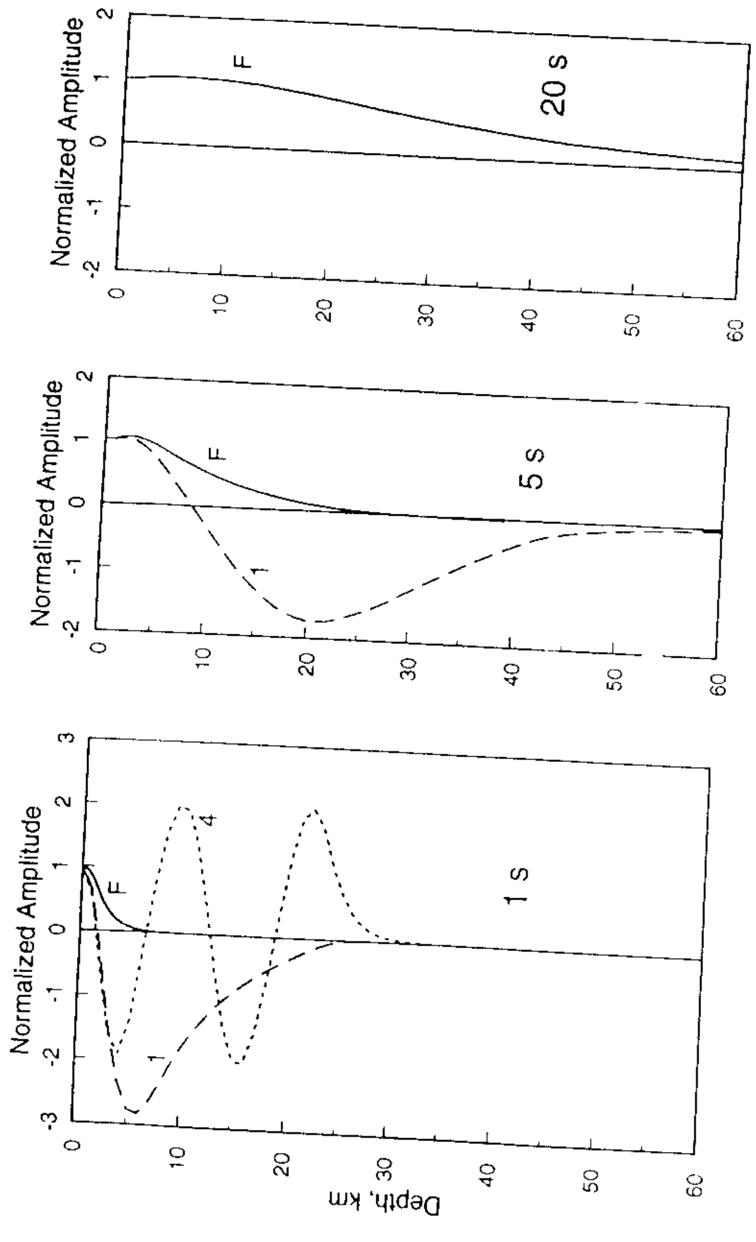

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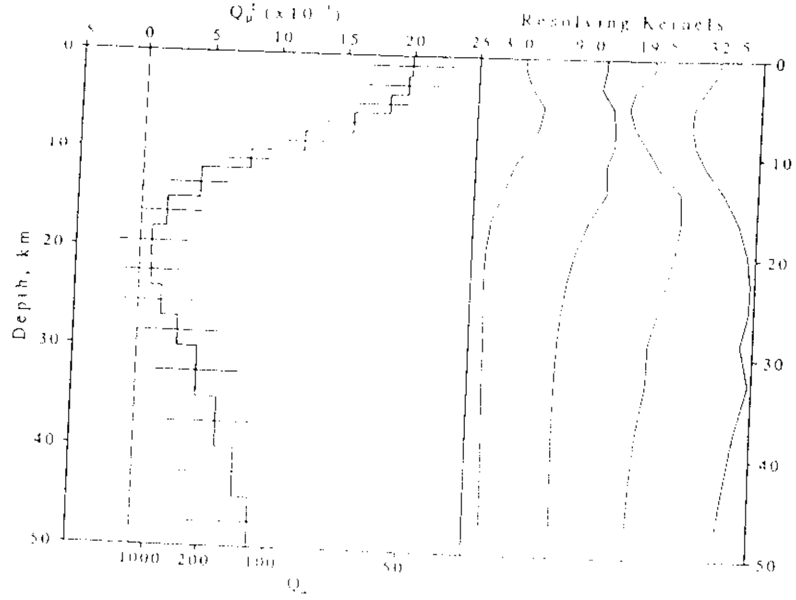

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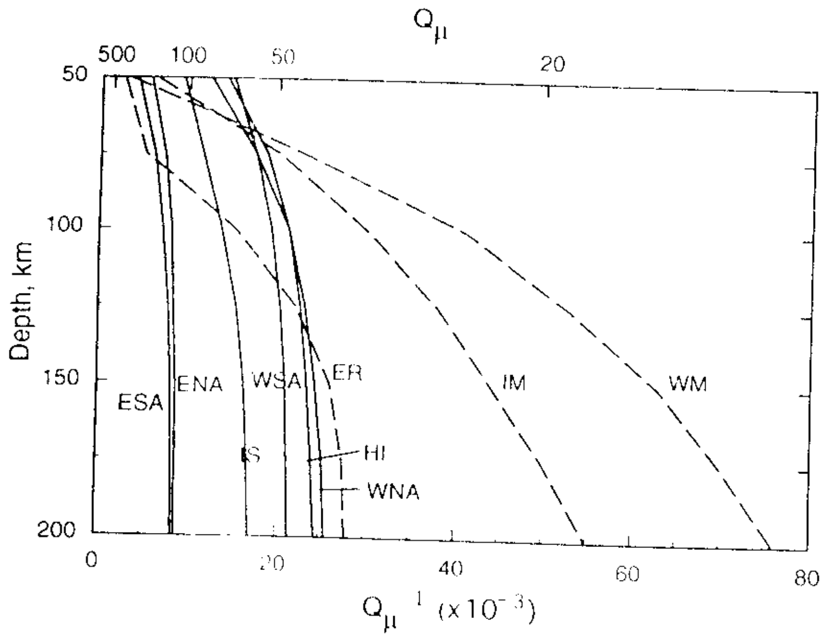

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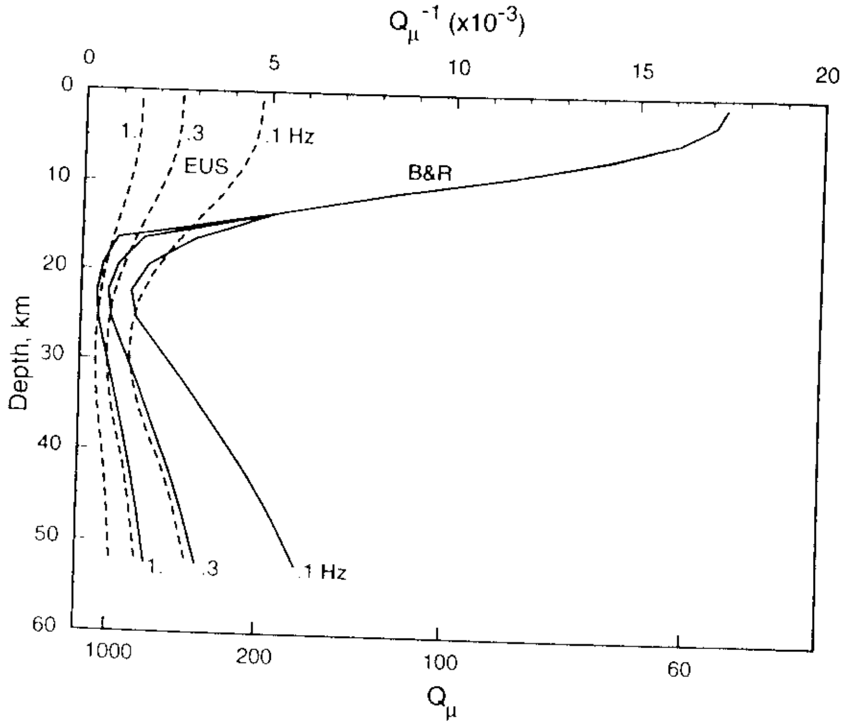

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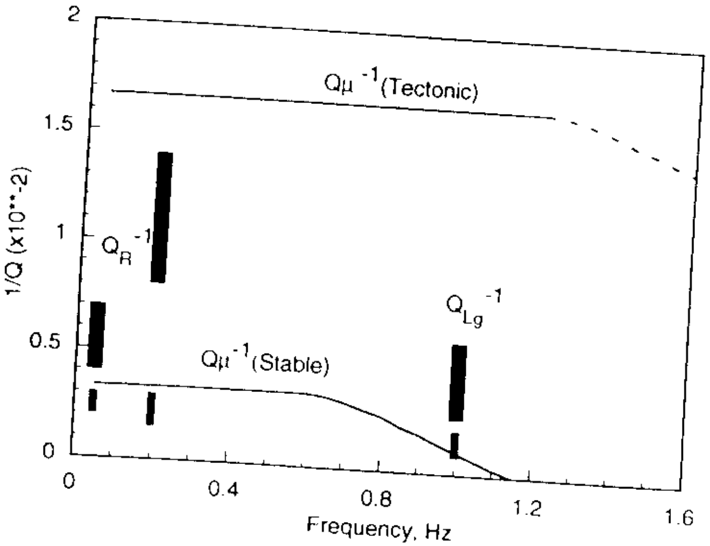

Figure 16

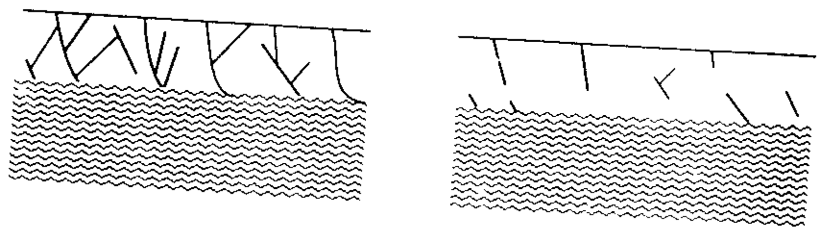

Figure 17