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 intrinsic shear wave quality factor Q_μ in both the upper crust and upper mantle of continents 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. Q_μ depends upon frequency, at least near 1 Hz, and that frequency dependence also varies regionally in the upper crust. It is typically low in tectonically active regions and higher 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 Q estimated from the coda of Lg waves (Q_{Lg}^c), even though both measurements may be marked by large uncertainties. Although coda Q of direct body waves may be strongly affected by scattering, Q_{Lg}^c appears to be primarily governed by intrinsic Q_μ in the upper crust. Both upper crustal Q_μ and Q_{Lg}^c values correlate with the time that has elapsed since the most recent tectonic activity in continental regions. A tomographic image of the variation of Q_{Lg}^c values across Africa shows reduced Q values which correspond to recent tectonic activity in the East African rift system and other regions of Mesozoic or younger age. Reductions of Q_{Lg}^c that correlate with tectonic activity that occurred in the early Paleozoic during the coalescence of the cratons which formed that continent can also be detected. Q_μ increases rapidly at midcrustal depths, in a range which appears to coincide with the transition to the plastic lower crust. In the lower crust and upper mantle, Q_μ decreases with increasing depth, possibly by progressive unpinning of dislocations with increasing temperature. Observed regional variations in upper mantle Q_μ at depths of about 150 km can be explained by differences in temperature alone, but those at crustal 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 which fluids can move during the propagation of seismic waves. Studies of the regional variation of Q_μ and Q_{Lg}^c indicate that crack density is greatest during and immediately following tectonic activity in a region and that it decreases with time. Permeability determinations in deep wells show that fluid movements in those cracks may be largely restricted to zones of crustal fracturing. That situation will produce widely differing values of Q in local studies, depending on the location of the study relative to the fractures. The fluid volume in cracks appears to decrease with time by loss to the surface or by retrograde metamorphism, causing a reduction in the number of open cracks and a concomitant increase in Q_μ.

INTRODUCTION

Seismic waves in 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 either by intrinsic anelasticity of the 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 4 decades. Another rationale for studying seismic wave attenuation is the need to determine the dispersion produced by anelasticity. Velocity dispersion must be associated with intrinsic attenuation to satisfy causality [*Lomnitz*, 1957; *Futterman*, 1962; *Strick*, 1967]. That dispersion must be accounted for in order to reconcile velocity models of 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 that 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

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

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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*, 1964; *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 Earth as inferred from the attenuation of seismic surface waves, including both fundamental and higher modes. Fundamental-mode studies at periods between about 5 and 100 s (0.01–0.2 Hz) 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 measurements in this review are mainly restricted to frequencies near 1 Hz and concentrate on the *Lg* phase and its coda. *Lg* is a superposition of higher modes for which the onset travels at group velocities between 3.3 and 3.6 km/s, and the coda consists of scattered waves which follow *Lg* and decay slowly with time, being visible up to several minutes later. *Lg* and its coda are prominent on regional short-period and broadband seismograms recorded in continental regions. Information on *Lg* coda allows us to place constraints on the frequency dependence of Q_μ and to study its regional variation in greater detail than is possible with fundamental-mode data.

The higher modes, because they encompass a higher frequency range than does the fundamental mode, however, lead to the additional possibility of scattering effects obscuring those of intrinsic attenuation. When these waves are used to constrain the distribution of intrinsic anelasticity, the effects of scattering must either be taken into account or be shown to be unimportant. 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 in the Moon, rather than in 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 and Toksöz* [1981] showed that the lunar coda is best explained by intense scattering in a near-surface layer with low intrinsic absorption. Absorption is small because there are no volatiles present in the Moon, at least near the surface [*Tittman et al.*, 1976]. On the other hand, geological evidence [*Cathles*, 1990; *Torgersen*, 1990] and information from deep drill holes [*Borevsky et al.*, 1984; *Morrow and Byerlee*, 1992] indicate that the upper crust of Earth contains abundant interstitial fluids which can migrate through cracks in igneous rock. Intrinsic absorption caused by that fluid movement can significantly 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 lesser importance of scattered phases on terrestrial, as opposed to lunar seismograms, is apparent from the seismograms compared in Figure 1. The record from station CCM at Cathedral Cave, Missouri in a high-Q region of the central United States, exhibits sharp onsets and brief coda duration, whereas the lunar seismogram has an emergent onset and the coda is still prominent 1 hour after the beginning of the wave motion.

This observation, as well as others discussed in a later section, leads us to interpret our results in terms of intrinsic Q in Earth over the frequency range from about 0.01 to 1.0 Hz. It would be incorrect, however, to accept all attenuation estimates as providing information on intrinsic Q in Earth. Attenuation determinations 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 Earth with spurious features. The following sections will discuss various methods of measuring surface wave attenuation and possible sources of error for each of them.

Although the potential for error in amplitude measurements is large, regional variations in intrinsic attenuation throughout the continental crust and upper mantle are also large, the span of reported values exceeding 2 orders of magnitude. Therefore even though Q models derived from surface wave attenuation data may have large uncertainties, we may still be able to obtain information about the variability of Q and its relation to crustal and upper mantle structure and evolution.

Figure 2 illustrates the variability of Q in continents by showing seismograms recorded along a high-Q and a low-Q path. The ground motion measured at the two broadband stations was produced by a magnitude 4.6 earthquake which occurred in southeastern New Mexico. The epicentral distance to CCM, along a predominantly high-Q path, is 1256 km, and that to station PAS (Pasadena, California), along a predominantly low-Q path, is 1417 km. If the anelastic properties along the two paths were the same, we would expect the amplitudes recorded at PAS to be comparable to those recorded at CCM. The attenuation is so much

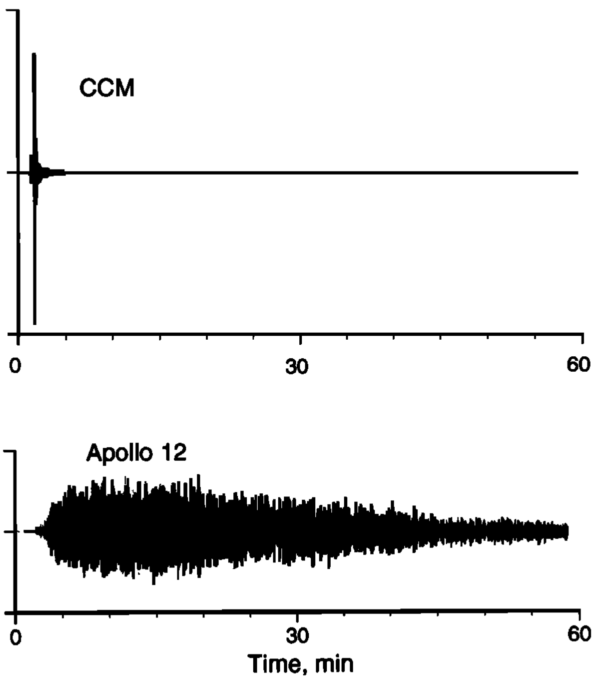

Figure 1. (top) A seismogram recorded by station CCM at Cathedral Cave, Missouri, generated by an earthquake of m_b 4.7 which occurred 183 km away. (bottom) A seismogram recorded by the Apollo 12 seismometer and which was produced by the impact of the S4B Saturn booster of Apollo 14 at a distance of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

greater along the path to PAS, however, that neither the P wave (arriving just before 3 min at CCM) nor the S wave (arriving at about 5 min at CCM) is observable 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. Although the absence of P and S waves on the PAS record could be caused by an upper mantle low-velocity zone, as well as by low Q values, the reduced Lg amplitudes can only be caused by greater seismic absorption along that path. 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 decay of Pg and Lg energy 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 of surface waves 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 to Rayleigh or Love waves, respectively. Q_R and Q_L are not intrinsic properties of Earth but refer to the attenuation of surface waves. That attenuation is affected by intrinsic anelasticity, as well as by seismic velocities over a range of depths in the crust and upper mantle. Intrinsic anelasticity can be described by the shear wave quality factor Q_μ and the bulk quality factor Q_κ. Q_R is a weighted average of those values and is also affected by intrinsic velocities over a range of crustal and upper mantle depths. The inversion of Rayleigh wave attenuation for Q_μ is discussed in a later section.

Because of the relative insensitivity of surface wave attenuation to Q_κ, inversions of surface wave attenuation values lead to models of Q_μ or its inverse Q_μ^{-1}. It

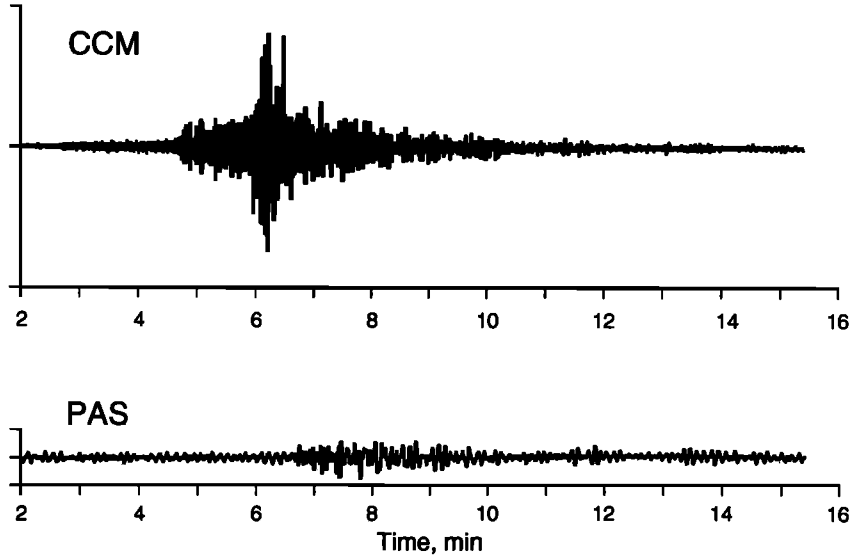

Figure 2. Seismograms recorded at stations CCM and PAS (Pasadena, California) from the same event in southeastern New Mexico which occurred on January 2, 1992, at 1145:35.6 UT and had a magnitude of 4.6. The epicentral distance to CCM (along a high-Q path) is 1256 km, and the epicentral distance to PAS (along a low-Q path) is 1417 km.

is usually assumed, in these inversions, 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 structure. Experience has shown, however, that determinations of surface wave attenuation coefficients can be uncertain because of all of those effects. Attenuation measurements therefore require great care and a careful selection of sources, station locations, and paths.

One way to minimize the effects of lateral structural variations is to use paths which are as short as possible. Surface waves, however, attenuate very slowly with distance, 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 short 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 great circle paths and two in-line stations, 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 that would be suitable for the application of a particular method, several methods for determining surface wave attenuation coefficients have been developed. These are briefly described in the following paragraphs.

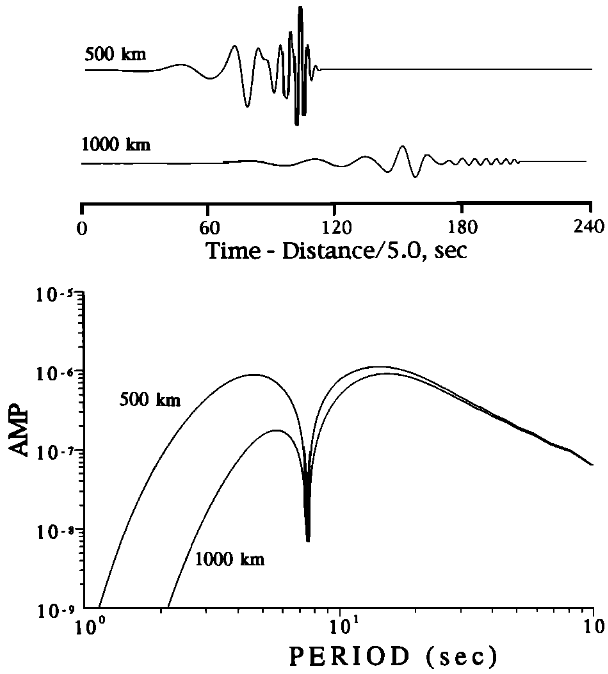

Figure 3. (top) Fundamental-mode Rayleigh waves computed at distances of 500 km and 1000 km along an azimuth of 45° from the strike direction of a shallow (5 km) vertical strike-slip fault for a Basin and Range Province velocity and Q_μ model (Figure 9). (bottom) Amplitude spectra computed for the two seismograms.

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 [e.g., *Canas and Mitchell*, 1978; *Hwang and Mitchell*, 1987; *Al-Khatib and Mitchell*, 1991]. The method uses the equation

$$\gamma(\omega) = -\frac{\ln \{[A_2(\omega, r)/A_1(\omega, r)][\sin \Delta_2/\sin \Delta_1]^{1/2}\}}{r_2 - r_1} \tag{2}$$

where $A_{1,2}$ are observed spectral amplitudes at stations 1 and 2 at distances $r_{1,2}$ (in kilometers) and $\Delta_{1,2}$ (in radians) from the source and ω is angular frequency.

An ideal situation in which to invert surface wave attenuation data is one in which two stations are separated by a relatively large distance, across which elastic and anelastic properties do not vary laterally and in which the earthquake source lies on the great circle path that passes through the stations. 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.

Figure 3 illustrates the process by showing two theoretically expected seismograms from a strike-slip earthquake at a depth of 5 km in a laterally homogeneous velocity model with a low-Q upper crust and a high-Q lower crust (Figure 9). One station is 500 km and the other is 1000 km from the source, and both lie along the same azimuth (45° from the strike direction of the fault). Because of the high-Q lower crust, there is only a slight difference in spectral amplitudes at the two stations for periods greater than the spectral hole at 8 s. At shorter periods, however, the amplitudes at 1000 km are greatly reduced compared to those at 500 km because of the highly attenuating material in the upper crust. Using these amplitudes and (2), we can easily determine γ for the path between the two stations.

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

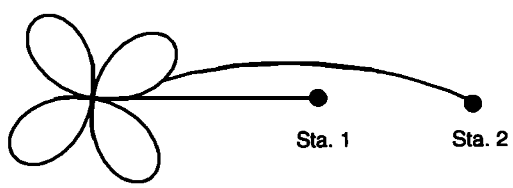

Figure 4. The amplitude radiation pattern which would be produced by a vertical strike-slip fault with a strike of about N10°E and schematics of possible paths to two recording stations if the waves are laterally refracted in media with laterally varying elastic properties.

at a different azimuth from those recorded at the other station. When this 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 along the path between the stations. This effect is illustrated schematically in Figure 4, where a path to the near station has left the source near a node in the radiation pattern, while a path to a more distant 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, a negative attenuation coefficient could be obtained. The possible arrival of waves from azimuths which depart from the great circle direction can be checked by examining the particle motion 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 and that in the absence of effects produced by lateral refraction, 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 lateral refraction can be avoided, it is the method of choice.

Single Source–Many Station (SM) 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

$$A_{r_0} = A_r(R_E \sin \Delta_r/r_0)^{1/2} \exp(\pi f r/UQ), \quad (3)$$

where A_r and A_{r_0} are spectral amplitudes at the epicentral distance r and the reference distance r_0, respectively, R_E is the radius of Earth, Q is the quality factor for Rayleigh or Love waves, and U is the corresponding group velocity measured from the seismograms. Taking the logarithm of both sides leads to an expression which can be 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 used in a region where anelastic properties vary laterally [*Yacoub and Mitchell*, 1977]. That effect can be illustrated by a hypothetical case in which an earthquake occurs near a vertical boundary which separates two regions with differing values of Q_μ. The measured attenuation coefficient will depend on the station distribution. If stations in the low-Q region are predominantly nearer the source than are those in the high-Q region, the resulting amplitude-distance plot would lead to an attenuation coefficient which is lower (and could be negative) than that which would be produced if either of the Q_μ values in the model were uniform throughout the entire model.

A variation of the *Tsai and Aki* [1969] method that can be used when a source mechanism is not available was developed by *Mitchell* [1975]. Following *Toksöz and Kehrer* [1972], Mitchell 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. The method solves the nonlinear expression

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

by a least squares procedure. 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 that can progress from being circular (when $F = 0$) to being oblong, two-lobed, and four-lobed. In many inversions, values obtained for F, θ_0, and W_e will have no physical meaning and serve only to orient, shape, and scale the radiation pattern. The attenuation coefficient γ is, however, always realistic because it is the value which produces the smoothest fit between observed and theoretical radiation patterns.

The method of *Mitchell* [1975], like that of *Tsai and*

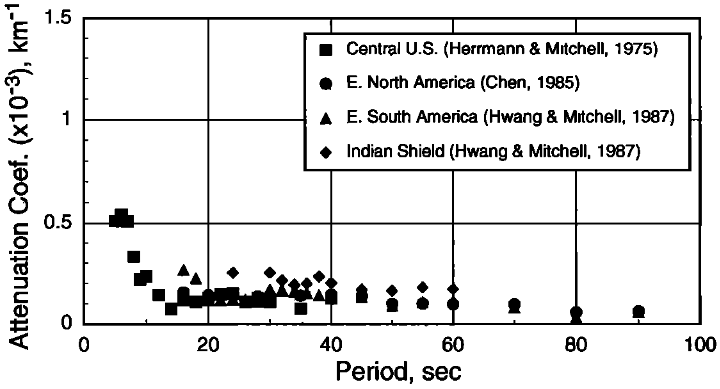

Figure 5. Summary of fundamental-mode Rayleigh wave attenuation coefficient determinations in stable continental regions at periods between 5 and 90 s. Standard deviations for all data points are smaller than 0.2×10^{-3} km^{-1}.

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 Q_μ can produce attenuation coefficients which are systematically biased. Such biases may be reduced 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 (SS) 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 SM methods. Two stations may not be available that 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 structures that produce focusing of surface waves. SM methods are useful only in situations where several stations surround a source and all lie within a single tectonic province or if it is possible to subdivide the region of study before determining attenuation coefficients. SS methods avoid the necessity of having two stations that lie along a single great circle path through the source. They have the advantage that it is often possible to find earthquake or explosion sources that lie within the same tectonic province as that of the station and therefore may provide a relatively uniform path of travel for surface waves.

With SS methods, however, either we must know the source depth and focal mechanism for each earthquake used, or we must be able to determine those parameters from the available data. 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. In so doing, they had to use events that were small enough that the source time function did not adversely affect their results.

Attenuation Coefficient Values From Fundamental-Mode Observations

Surface wave attenuation coefficients have now been determined for several continental regions. Figure 5 presents many of the results for Rayleigh waves that have propagated across stable portions of continents. The results of several studies that include some paths across stable regions have been excluded because they span more than a single tectonic province [*Tsai and Aki*, 1969; *Solomon*, 1972; *Yacoub and Mitchell*, 1977]. In addition, Figure 5 includes only those values for which the standard deviation is less than 0.2×10^{-3} km^{-1}. Several values from the studies of *Chen* [1985] and *Hwang and Mitchell* [1987] were not plotted for this reason. Figure 5 indicates that at periods between about 20 s and 60 s, the attenuation coefficients fall within a narrow range of values centered at about 0.15×10^{-3} km^{-1}. At longer periods the values appear to decrease slightly, although the uncertainty would allow higher values, comparable to those obtained in the 20- to 60-s period range. At periods less than about 15 s the attenuation coefficients increase with decreasing period and reach about 0.5×10^{-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 have standard deviations of 0.2×10^{-3} km^{-1} or less. Data with larger standard deviations (up to 0.3×10^{-3} km^{-1}) have, however, been included at short periods [*Patton and Taylor*, 1984; *Lin*, 1989]. Because of their shorter wavelengths and because they have traversed a laterally complex region, those waves are more likely to have been adversely affected by the laterally complex structure. At periods between 20 and 100 s the mean attenuation coefficient stays relatively constant at about 0.25×10^{-3} km^{-1}. At shorter periods the attenuation increases sharply with decreasing periods and reaches values as great as 3.0×10^{-3} km^{-1} at periods near 5 s. Three sets of values stand out from the main trend in Figure 6. Those are values from the

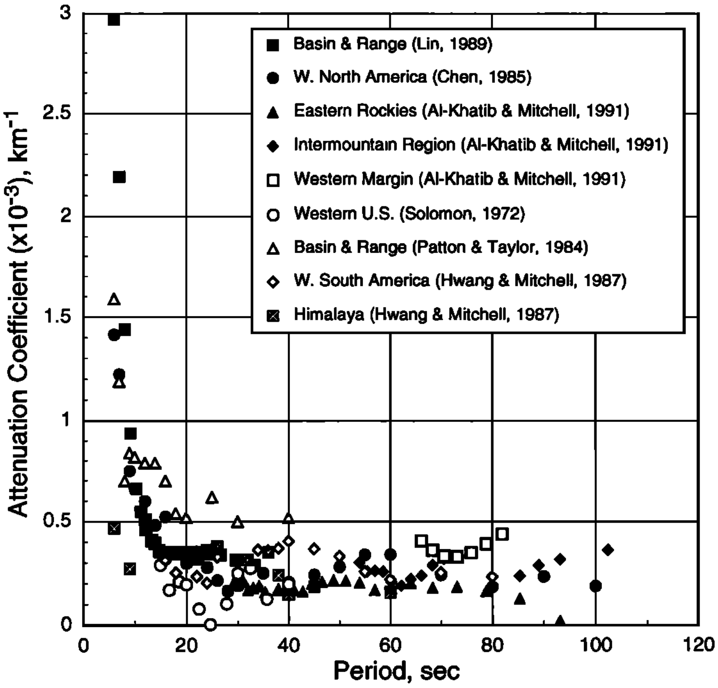

Figure 6. Summary of fundamental-mode Rayleigh wave attenuation coefficient determinations in tectonically active regions at periods between 6 and 103 s. Standard deviations are smaller than 0.2×10^{-3} km^{-1} for all data points except those of *Patton and Taylor* [1984] and *Lin* [1989] for which standard deviations may as large as 0.3×10^{-3} km^{-1}.

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 that 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- to 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 periods of 15–40 s in the Basin and Range province may occur because of the application of an SM method in a region where Q is apt to vary laterally. Alternatively, there may be regions of low Q in that 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 contributes to those high attenuation coefficients cannot be ruled out.

Figure 7 summarizes the results of Figures 5 and 6. The two continental data sets are compared to each other, as well as 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

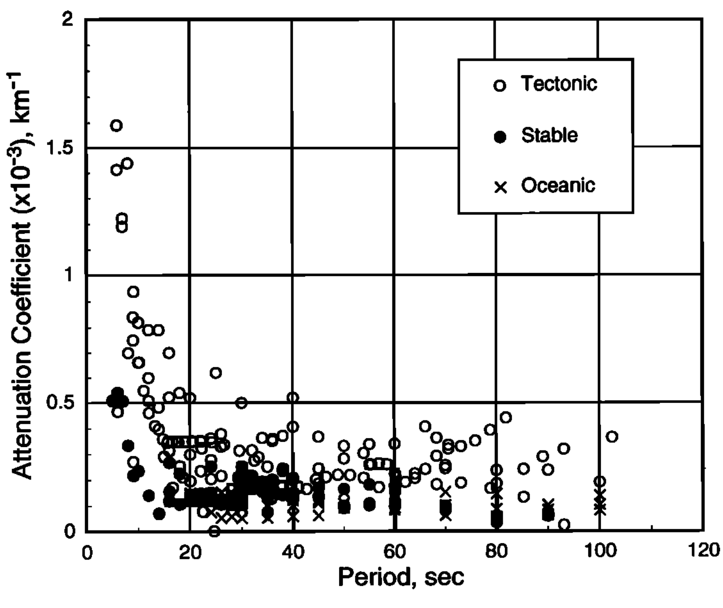

Figure 7. Comparison of fundamental-mode Rayleigh wave attenuation coefficient determinations for tectonically active and stable continental regions (Figures 5 and 6) with those for oceanic regions [*Canas and Mitchell*, 1978]. The three oceanic values at each period correspond to three different age ranges (0–50 m.y., 50–100 m.y., and >100 m.y.) of lithosphere formation. Standard deviations for the continental data are described in the captions for Figures 5 and 6. Standard deviations for the oceanic data average about 0.2×10^{-4} km^{-1} and are all smaller than 0.4×10^{-4} km^{-1}.

correspond to that part of the Pacific that formed between 0 and 50 m.y. ago, intermediate values correspond to lithosphere that is between 50 and 100 m.y. in age, and the lowest values correspond to those portions of the Pacific lithosphere that formed more that 100 m.y. ago. Tectonically active continental regions are characterized by higher attenuation coefficients than are stable continental regions through the entire period range. Attenuation coefficients for stable regions at periods greater than about 45 s lie within the error bounds for the span of values for the three oceanic regions. At periods between 25 and 45 s the lowest observed attenuation coefficients are the oceanic values for the region of the Pacific which is greater than 100 m.y. in age, while the younger oceanic values are comparable to those of stable continental regions. There is, however, overlap between the error bars for the measurements in the stable continental and all oceanic regions.

At periods shorter than about 20 s the values obtained in tectonically active regions show a sharp increase with decreasing period, reaching values that are over 3 times larger than those in stable regions at periods near 5 s. The rapid increase is evident in the results of *Lin* [1989] for one path within the Basin and Range province and *Patton and Taylor* [1984] for a broader region of the same province.

Lg CODA *Q*

Lg coda is that oscillatory portion of a short-period regional seismogram which follows the *Lg* wave and which may continue for several minutes. Q for *Lg* can be represented as $Q_0 f^{\eta}$ and that of *Lg* coda Q by $Q'_0 f^{\eta'}$ where Q_0, and Q'_0 are Q values at 1 Hz while η and η' represent frequency dependence. Although Q_0 and η may, in general, differ from Q'_0 and η', several studies find that those values are similar [*Der et al.*, 1984; *Nuttli*, 1988; *Xie and Nuttli*, 1988; *Xie et al.*, 1995]. The present paper will refer to Q values for *Lg* and *Lg* coda, respectively, as Q_{Lg} and Q^c_{Lg}. Measurements of attenuation of *Lg* and its coda, like those for fundamental-mode surface waves, are often contaminated by errors which are dominantly systematic in nature. The severity of those errors can, however, sometimes be reduced by careful selection of sources, stations, or procedures.

Because *Lg* and its coda are characterized by higher frequencies than are fundamental-mode data, they can also be affected by random perturbations from an assumed laterally homogeneous layered structure [*Xie and Mitchell*, 1990b]. Stationary random perturbations can be treated as a statistical distribution that 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]. Systematic deviations from a laterally homogeneous layered structure may occur as a random but laterally varying statistical distribution or may be stationary over a region of interest [*Cara et al.*, 1981]. 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 *Lg* 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 Q^c_{Lg} include 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 led to coda Q determinations with sufficient stability and precision to permit Q^c_{Lg} tomography.

Q^c_{Lg} Tomography

Q^c_{Lg} 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 first divide the region into a number of rectangular cells, and the unknown Q^c_{Lg} is assumed to have a constant value Q_m inside each cell. We take Q_n to be the Q value obtained from the nth seismogram using the stacked spectral ratio method and assume that it represents the areal average of Q^c_{Lg} 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 that overlaps the mth grid by s_{mn}, we obtain

$$\frac{1}{Q_n} = \frac{1}{S_n} \sum_{m=1}^{N} \frac{s_{mn}}{Q_m} + \varepsilon_n \qquad n = 1, 2, \cdots, N \quad (5)$$

where

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

and ε_n is the residual misfit. This set of linear equations can be solved for $1/Q_m$ by minimizing the misfit ε_n, as discussed by *Xie and Mitchell* [1990a]. Because coda Q tomography produces maps that show the spatial variation of Q^c_{Lg} and its frequency variation over large portions of continents, it is well suited for studying the relationship between attenuation and 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]. The African results, along with a new interpretation of those results in terms of the tectonic evolution of that continent, are presented later in this section.

Is Q_{Lg}^c 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}^c 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 Q_{Lg}^c 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, Q_{Lg}^c in the Appalachian Mountains is higher than that obtained for the Atlantic coastal plain. These types of qualitative observations led *Mitchell* [1980] to infer that Q_{Lg}^c reflects intrinsic Q_μ in the crust and to conclude that it would be valid to combine observations of Q_{Lg}^c with attenuation coefficient data to infer frequency-dependent Q_μ models for the continental crust. Several types of evidence support that conclusion. First, as discussed in the following section, patterns of regional Q_{Lg}^c variation are the same as those of Q_μ in the upper crust inferred from studies of fundamental-mode surface wave attenuation. Second, the equalities of Q and the frequency dependence of Q for Lg waves with those for Lg coda in the Basin and Range province [*Xie and Mitchell*, 1990b] suggest that intrinsic attenuation is high there and is the predominant dissipation mechanism affecting both types of wave. Third, a recent study of Lg coda propagation in the western United States which separated the effects of intrinsic Q and scattering found that the contribution of scattering was weak [*Mayeda and Walter*, 1994; K. M. Mayeda, personal communication, 1994]. Fourth, a recent computational study using realistic crustal models found that lateral crustal complexity has only a small effect on Lg amplitudes [*Bouchon*, 1995]. Since Q_{Lg} and Q_{Lg}^c are similar in most regions, we can infer that both Q_{Lg} and Q_{Lg}^c are governed by intrinsic Q_μ in the crust.

Information pertaining to the relative contributions of intrinsic anelasticity and scattering to body wave attenuation at short distances presents a less consistent picture. Some computational [*Frankel and Wennerberg*, 1987; *Hoshiba*, 1991] and theoretical [*Zeng*, 1991; *Zeng et al.*, 1991] studies 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, then intrinsic anelasticity is 5–15 times more significant for coda attenuation than is scattering. Several observational studies of shear wave attenuation at short distances, however, have found that the contribution of scattering to attenuation is higher than that of intrinsic anelasiticity [*Aki*, 1980a, b; *Mayeda et al.*, 1992; *Jin et al.*, 1994].

Observed Q_{Lg}^c Values

Numerous single-path determinations of Q_{Lg}^c and its frequency dependence have been made in continental regions, mostly at frequencies of about 1 Hz. Tectonically stable regions are usually characterized by high values of Q_{Lg}^c (800–1200) and a weak frequency dependence (η = 0.0–0.3). These ranges 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 Q_{Lg}^c (150–400) and stronger frequency dependence (η = 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 measurements of Q_{Lg}, since Q_{Lg} and Q_{Lg}^c are usually found to be similar [*Nuttli*, 1988]. Q for body waves measured at relatively short distances is also high in the stable craton of eastern North America [*Al-Shukri et al.*, 1988] and 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 Q_{Lg}^c over broad regions. Plate 1 displays a tomographic image of Q_{Lg}^c for Africa that is slightly modified from *Xie and Mitchell* [1990a]. The most conspicuous feature of the map is the band of low Q which lies along the East African Rift Zone, a tectonically active region. Because the resolving power of the inversion is limited (500–2000 km), it is possible that the low Q values in the crust that produce that band are really restricted to a much narrower region, characterized by lower values than those shown on the map. Other regions of low Q are the Cape Fold Belt at the southern tip of Africa, part of an orogenic belt of late Paleozoic–early Mesozoic age; the Atlas Mountains in Morocco, of Alpine age; and the Cameroon Line, the line of recent volcanic activity extending northeastward from the sharp bend in the western coast [*Cahen et al.*, 1984].

Almost all of the rest of Africa has been stable since the Pan-African orogeny, which ended about 550 m.y. ago [*Clifford*, 1970]. There are three regions where Q_{Lg}^c is higher than 800: the western portion of the West African Craton, the eastern portion of the East Sahara Craton, and the north central portion of the Kalahari Craton. These three cratons, along with the Congo Craton, just to the west of the southern part of the East

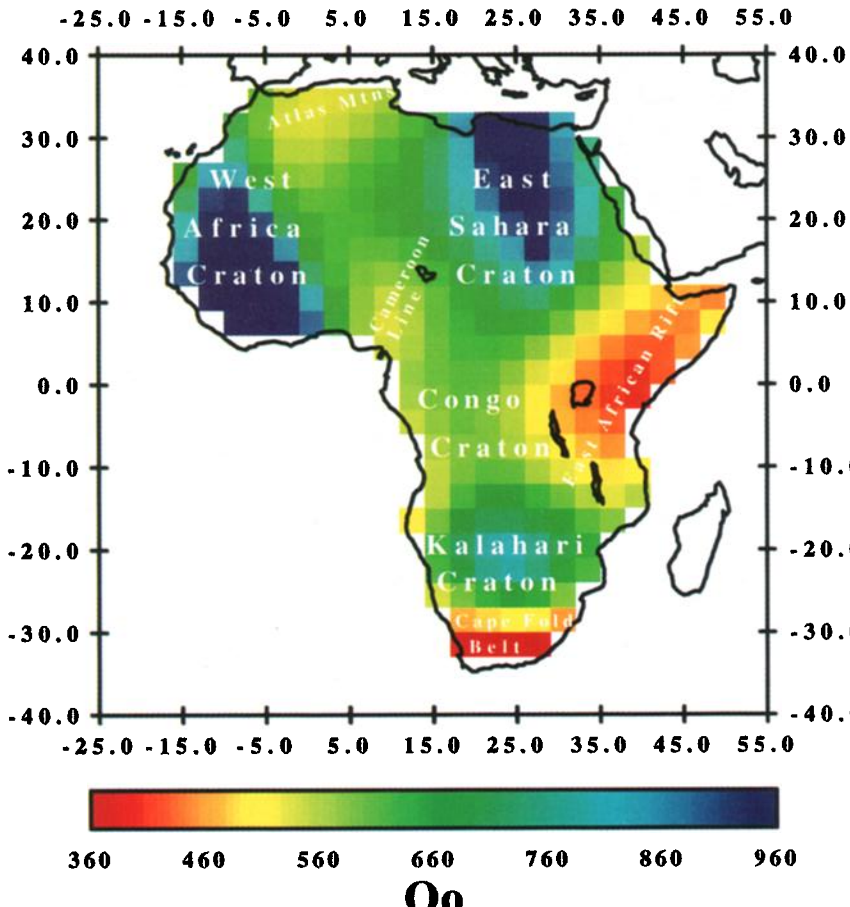

Plate 1. Tomographic map of *Lg* coda *Q* values at 1 Hz for the African continent with names of some major tectonic features superposed (slightly modified from *Xie and Mitchell* [1990a]). The terminology and locations of the cratons are adopted from *Kröner* [1979], and those of other features are from *Clifford* [1970] and *Cahen et al.* [1984].

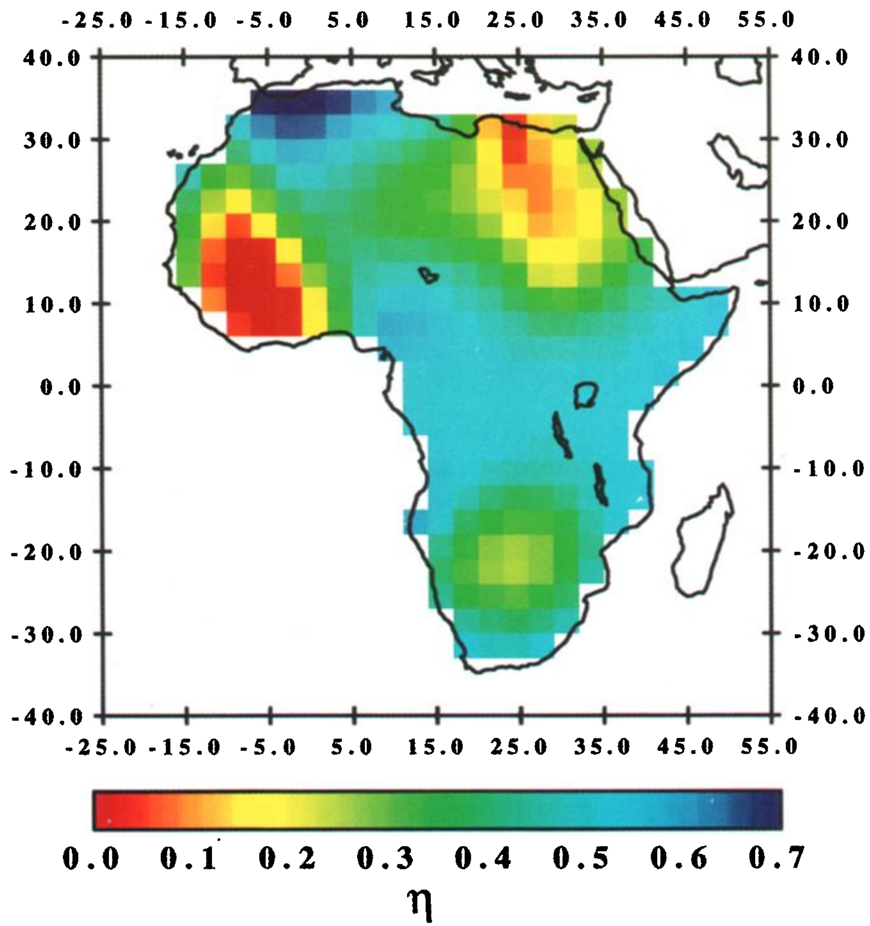

Plate 2. Tomographic map of the frequency dependence η of *Lg* coda *Q* at 1 Hz for the African continent.

African rift system, coalesced to form the African continent in a vast tectono-thermal event that ceased about 550 m.y. ago [*Clifford*, 1970]. Q values obtained for the Congo Craton are lower than those obtained for the other cratons of Africa, being about 600. These lower values may be real and may result from 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 inversion prevents us from delineating a small region of higher Q.

Plate 2 shows the frequency dependence exponents η of Q_{Lg}^c at a frequency of 1 Hz. The η values in western Africa have recently been redetermined, and those in the northwestern part of the continent differ from those obtained by *Xie and Mitchell* [1990a]. They were redetermined because the instruments of the Ivory Coast network, which were used for measurements in that region, differ in their responses from the World-Wide Standard Seismograph Network (WWSSN) instruments used throughout the rest of Africa for Q_{Lg}^c determinations. The redetermination finds frequency dependences that better fit the stacked spectral ratios in that region. Note that weak frequency dependences correlate with high Q values.

With the exception of the Congo Craton, Q_{Lg}^c within the major cratons of Africa is higher than that in the fold belts between them. This suggests that the variations of Q_{Lg}^c in the stable portions of Africa reflect tectonic activity associated with the coalescence of the cratons into the African continent and that Lg coda studies might be used to study episodes in the tectonic evolution of continents as early as the early Paleozoic era.

Although Q_{Lg}^c, as well as Q_{Lg}, is usually high in stable cratons, there are at least three exceptions to that rule. They are the Arabian peninsula, where Q_{Lg}^c values range between 157 and 300 [*Ghalib*, 1992], central Australia, where Q_{Lg} is about 230 [*Bowman and Kennett*, 1991], and the Siberian Craton, where Q values range between about 400 and 600. Low Q for the Arabian peninsula has also been inferred from the spectra of fundamental-mode Rayleigh waves [*Seber and Mitchell*, 1992]. Those low values correlate with recent tectonic activity that is known to have occurred in the region [*Chazot and Bertrand*, 1993]. The low Q values in the Siberian craton occur in a region that underwent extension and volcanism during the Mesozoic era [*Zonenshain et al.*, 1990].

The lower than expected Q_{Lg} observed in the central part of the Australian shield has been attributed to a large positive velocity gradient in the lower crust of that region [*Bowman and Kennett*, 1991] which reduces Q_{Lg} by increasing the effect of geometrical spreading. The possibility that crustal velocity structure could significantly affect Q_{Lg} in other regions was investigated in a study of the Basin and Range province [*Mitchell and Xie*, 1994]. They found that changes in velocity structure which might be expected from various seismic refraction studies in the region, change Q_{Lg} by only 11% in the worst case and are more typically less than 7%. Departures from assumed velocity structure therefore appear to be an important factor affecting Q_{Lg} determinations only in cases where the departures are much larger than those typically observed. In Australia there is also geological evidence for tectonic activity there during the Devonian and Carboniferous periods [*Plumb*, 1979], which may also contribute to low Q_{Lg}.

INVERSION OF FUNDAMENTAL-MODE ATTENUATION DATA

The methodology for inverting surface wave attenuation coefficients to determine anelastic properties was developed 3 decades ago [*Anderson and Archambeau*, 1964; *Anderson et al.*, 1965] and was used to obtain upper mantle models of Q_μ, assuming that Earth was radially symmetric and that Q is $\gg 1$ and independent of frequency. A more complete treatment simultaneously inverts for Q and velocity as functions of depth [*Lee and Solomon*, 1978]. Although that procedure improves the uncertainties in resulting Q_μ models, it produces models that are essentially the same as those produced by methods which solve only for Q [*Lee and Solomon*, 1978]. We are only interested in broadscale features of the distribution of Q_μ with depth; therefore all the models that 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 frequency dependence is constant with depth [*Mitchell*, 1980], and more recently to allow frequency dependence to vary with depth [*Mitchell and Xie*, 1994]. In this case, Q_μ in layer l (Q_{μ_l}) is expressed as $Q_{0l}f^{\zeta_l}$, where Q_{0l} is the Q_μ value at 1 Hz for layer l and the exponent ζ describes 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 Earth and may be affected by both velocity and Q_μ structure [*Mitchell*, 1991]. The Rayleigh wave attenuation coefficient γ_R in this formulation is

$$\gamma_R = \frac{\pi}{C_R^2 T} \sum_{i=1}^{N} \left[\left(\beta_l \frac{\partial C_R}{\partial \beta_l} \right)_{\omega\rho\alpha} + \frac{1}{2} \left(\alpha_l \frac{\partial C_R}{\partial \alpha_l} \right)_{\omega\rho\beta} \right] Q_{\mu_l}^{-1} \tag{6}$$

where R, α, and β identify Rayleigh, compressional, and shear waves, respectively; ρ is density; ω is angular frequency; C_R is Rayleigh wave phase velocity; and T is period. The subscripts, ω, ρ, 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 compres-

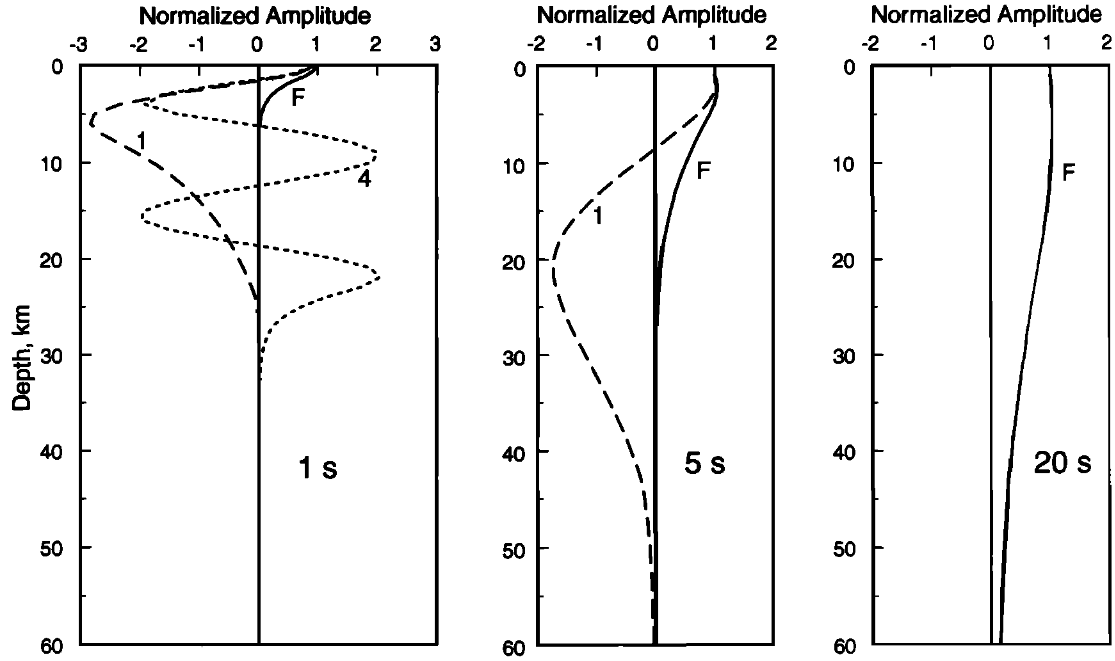

Figure 8. Rayleigh wave eigenfunctions for the fundamental, first, and fourth higher modes at a period of 1 s, for the fundamental and first higher mode at a period of 5 s, and for the fundamental mode at a period of 20 s. All eigenfunctions were computed for a velocity model of the eastern United States [*Mitchell and Herrmann*, 1979].

sional wave Q is twice as large as Q_μ. Previous studies, such as that of *Mitchell* [1980], indicate that inversion results are affected only slightly by moderate changes in the ratio of compressional wave to shear wave Q.

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 data are available [*Mitchell*, 1980]. Figure 8 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 is included, significant amplitudes are 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 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 20–30 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] combined inversions of (6) and a 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_l and inverts fundamental-mode attenuation coefficient data to obtain Q_μ^{-1}. If the resulting model is acceptable (i.e., it predicts attenuation coefficients which agree with observed values and does not fluctuate wildly), Q_{Lg} and its frequency dependence at 1 Hz are determined and compared 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 Q_0 and η. This process can be repeated for various depth distributions of ζ. Because of the large uncertainties in attenuation measurements, as well as the nonuniqueness inherent in Q inversions, there will be many models which satisfy fundamental-mode and Lg data. As found by Mitchell and Xie in the Basin and Range, however, some features of the models are independent of the depth distribution assumed for ζ.

Modern inversion theory [*Backus and Gilbert*, 1968] not only provides a direct determination of Q_μ with depth but also gives uncertainties in the model and estimates of the resolvability of features in the model. Figure 9 shows a model obtained by inverting fundamental-mode Rayleigh wave attenuation data, obtained in the Basin and Range, assuming $\zeta = 0.0$ (the frequency-independent case). Low Q_μ values (≈ 50) characterize the upper crust, high values (≈ 1000) occur at midcrustal depths, and values decrease gradu-

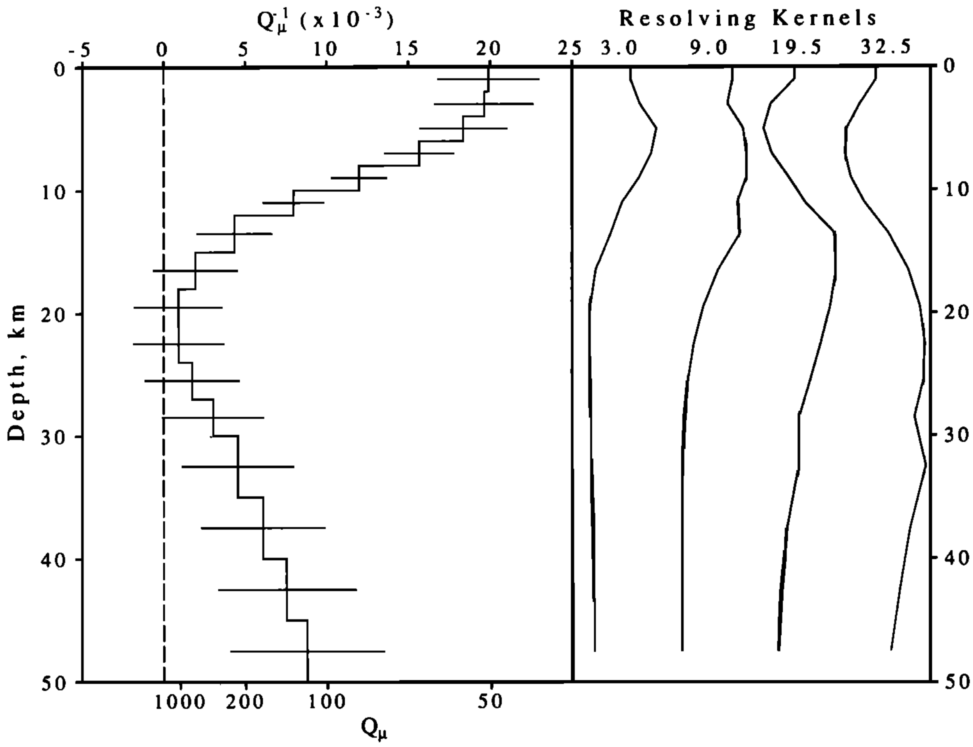

Figure 9. (left) Q_μ^{-1} model for a path across the Basin and Range province obtained from an inversion in which it was assumed that Q_μ is independent of frequency (ζ = 0) at all depths. The model adequately satisfies both fundamental-mode attenuation coefficient data and observed values for Q_{Lg} (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 [from *Mitchell and Xie*, 1994].

ally at greater depths. The resolving kernels in the right side of Figure 9 indicate that resolution is best in the upper crust (10 km or less) and degrades with increasing depth (20 km or more at 32.5 km depth). The value obtained for Q_μ at any depth is an average of the true model over the width of the corresponding resolving kernel. Clearly, only gross features of the model can be resolved, and any changes that occur over small depth ranges have no real meaning. The standard deviations in Figure 9 are small enough that we can attribute significance to the major features of the model. A preset parameter allows us to trade resolution off 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 that fluctuate wildly from layer to layer. The resolution kernels and standard deviation values for the model in Figure 9 are fairly typical and can be taken to be roughly representative of the models shown in subsequent figures.

Upper Mantle Models

As mentioned earlier, the first Q_μ models of the mantle assumed that the anelastic properties of Earth were spherically symmetric. Since most proposed attenuation mechanisms are thermally activated and since temperatures in the upper mantle vary significantly [*Pollack et al.*, 1993], it is likely that there will be regional variations of Q_μ. The Rayleigh wave attenuation coefficient values for the regions of the world shown in Figures 5, 6, and 7 also suggest that Q_μ must vary regionally in both the crust and upper mantle.

Models that have been obtained from the inversion of surface wave attenuation coefficients indicate variations of upper mantle Q_μ over broad continental regions. Such models include those for eastern (ENA) and western (WNA) North America [*Solomon*, 1972; *Lee and Solomon*, 1978; *Chen*, 1985; *Hwang and Mitchell*, 1987], as well as eastern (ESA) and western (WSA) South America, the Indian shield (IS), and the Himalaya (HI) [*Hwang and Mitchell*, 1987]. In addition, regional variations of Q_μ have been found to occur in the upper mantle within the western Cordillera of North America [*Al-Khatib and Mitchell*, 1991]. Figure 10 summarizes models obtained in the last 10 years for both stable and tectonically active continental regions. The broader-scale model results (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 a

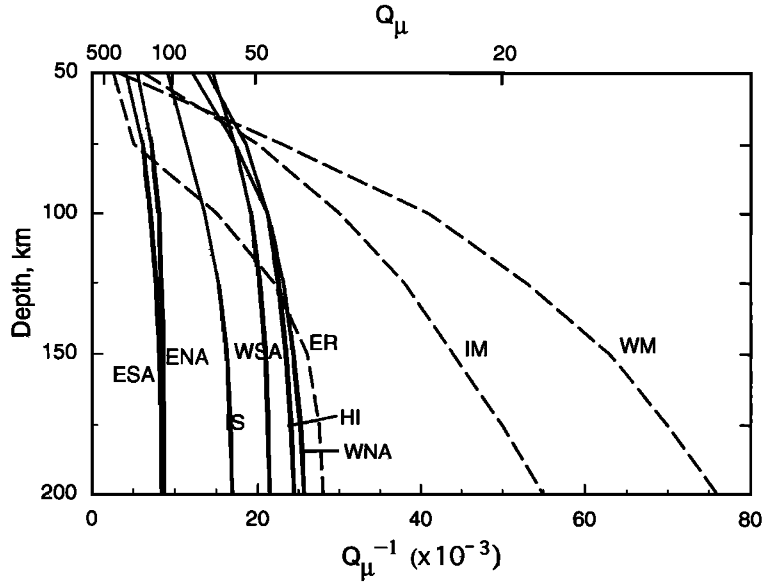

Figure 10. Upper mantle models of Q_μ^{-1} determined for various stable and tectonically active regions. ESA (eastern South America), ENA (eastern North America), IS (Indian shield), WSA (western South America), HI (the Himalaya), and WNA (western North America) models were determined from the inversions of two-station measurements of Rayleigh wave attenuation coefficients on three continents [*Hwang and Mitchell*, 1986]. 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].

higher Q_μ at all depths than do models for tectonically active regions (WSA, HI, and WNA). *Hwang and Mitchell* [1987] found Q_μ at depths between 100 and 200 km in stable regions to be 125–150 and found those in tectonically active regions to be 40–70.

Al-Khatib and Mitchell [1991] divided the western United States into three regions that they called the eastern Rockies (ER), largely confined to the Rocky Mountains, the Intermountain region (IR), stretching from the western margin of the Rocky Mountains westward through the Basin and Range, and the Western Margin (WM), consisting mostly of California and parts of western Oregon. With this finer regionalization they found Q_μ in the uppermost mantle of all three regions to be lower than that obtained for tectonically active regions by *Hwang and Mitchell* [1987]. At depths greater than about 125 km, Q_μ for the eastern Rockies is similar to that in the tectonically active regions, but lower values occur in the Intermountain region (20–40) and the Western Margin region (15–50). This comparison indicates that Q_μ in the upper mantle varies widely from region to region and that models derived from data acquired for long paths over broad regions can lead to models that differ from models derived from data obtained over smaller areas within that broad region. The gross consistency of the patterns of Q_μ distribution indicates that models obtained from surface wave attenuation reflect real changes of intrinsic attenuation in the upper mantle. As indicated in an earlier section, however, amplitude data are subject to systematic bias, which often cannot be accounted for properly. Conclusions about the distribution of Q_μ in Earth should therefore be based only on 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 and t^* (travel time/Q) values estimated from teleseismic body waves. Studies at both long periods [*Solomon and Toksöz*, 1970] and short periods [*Der et al.*, 1982] require Q in the upper mantle beneath the western United States to be lower than that beneath the central and eastern United States.

Currently available methods and data preclude the resolution of 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 et al.* [1982] have summarized available information on that topic. Frequency dependence of Q, at seismic frequencies, is also observed in laboratory measurements on forsterite and dunite [*Berckhemer et al.*, 1982; *Jackson et al.*, 1992].

Crustal Models

Studies of crustal Q_μ require data at shorter periods than those used for mantle studies. Short-period 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 regions where both criteria can be met is still limited.

The displacement eigenfunctions in Figure 8 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. Therefore attenuation data from *Lg* waves at 1 s and fundamental-mode surface waves at longer periods can be combined to study the frequency dependence of Q_μ in the crust. Short-period stations of the WWSSN network (with a peak re-

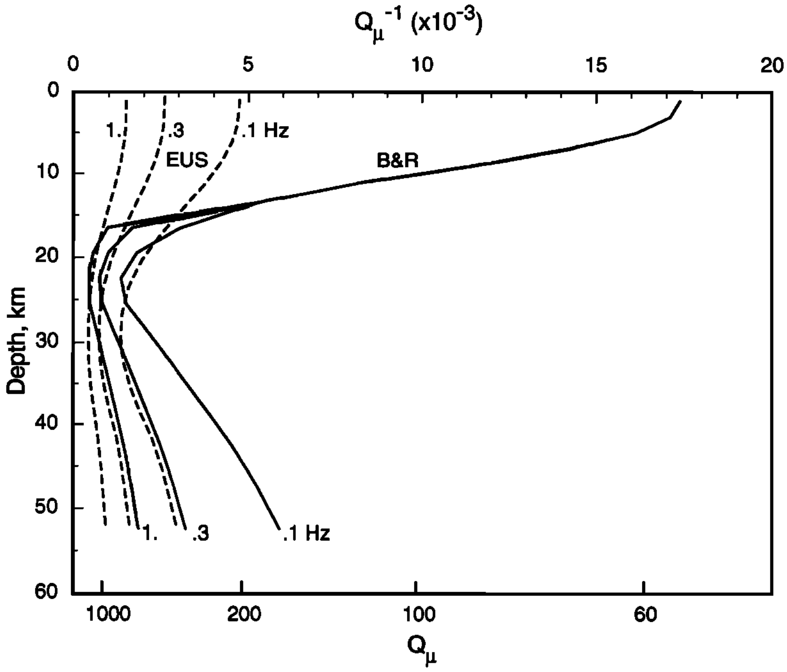

Figure 11. Frequency-dependent crustal models of Q_μ for the eastern United States (EUS) [*Cong and Mitchell*, 1988] and the Basin and Range province (B&R) [*Mitchell and Xie*, 1994]. The EUS model assumes a uniform frequency dependence ($\zeta = 0.5$) at all depths and the B&R model assumes that Q is independent of frequency to depths of 15 km and that frequency dependence at greater depths is 0.5.

sponse near 1 s) began operating in the early 1960s, and many recordings of the *Lg* phase have been amassed since then. There is therefore a large database for studies of 1-s *Lg* waves.

Regions where both fundamental-mode and *Lg* waves have been used to obtain crustal models of Q_μ are still few in number. 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 studies in both regions can be considered to be relatively free of focusing effects produced by lateral variations in structure. 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 examination of surface wave–particle motion for the data used [*Mitchell et al.*, 1993].

Figure 11 shows models obtained for both regions. The eastern United States (EUS) model requires Q_μ to vary with frequency, at least in the range from about 0.3 to 1.0 Hz [*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 11 is plotted for the case where $\zeta = 0.5$ at all depths. This number is not well constrained and could vary with depth as well as frequency.

Crustal Q_μ models for the Basin and Range are also nonunique. The B&R model in Figure 11 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 a family of models which satisfactorily explained fundamental-mode Rayleigh wave and *Lg* attenuation data (Q_{Lg} and its frequency dependence at 1 s). Common features of all of those models are low Q_μ values in the upper crust, a rapid increase in Q_μ at midcrustal depths, and a weak average frequency dependence throughout the crust. If it is assumed that ζ does not vary 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. Frequency-independent models had previously been derived for the Basin and Range [*Cheng and Mitchell*, 1981; *Patton and Taylor*, 1984; *Lin*, 1989] that were not constrained by Q_{Lg} information.

A rapid increase in Q_μ at midcrustal depths was first 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 is a common feature of models derived from fundamental-mode surface wave attenuation when sufficient data at short periods are available. *Mitchell* [1991] found that a sharp increase in Q_μ at midcrustal depths can explain the strong frequency dependence of Q_{Lg} and Q^c_{Lg} observed at 1 Hz in the Basin and Range province. 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 with distance from the source [*Al-Shukri et al.*, 1988], also suggest an increase of Q with depth within the crust.

Overviews of Q^c_{Lg} studies indicate that Q_0 is typically high and its frequency dependence is typically weak in stable regions, whereas Q_0 is typically low and its frequency dependence is typically high in tectonically active regions [*Nuttli*, 1988]. Because of this correlation with tectonic history and clear contrast

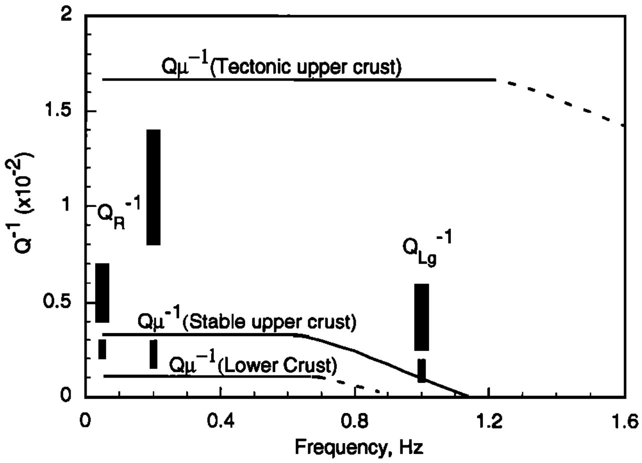

Figure 12. Thin solid and dashed lines denote values of Q_μ^{-1} derived for the lower continental crust and for the upper crust of stable [*Mitchell*, 1980] and tectonically active [*Mitchell and Xie*, 1994] regions. Dashed portions of the lines indicate conjectured frequency ranges where Q_μ^{-1} may become frequency dependent. The vertical bars toward the left of the figure denote observed ranges of values for Rayleigh wave internal friction Q_R^{-1} at 0.05 and 0.2 Hz (20-s and 5-s periods), and those toward the right (at 1 Hz) 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 obtained in stable regions.

between tectonic regimes, it is likely that the types of model obtained for North America, which explain *Lg* coda results there, are applicable to many of the world's continental regions. A rapid increase in Q_μ at midcrustal depths is likely to be a prevalent feature of the world's continents.

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

Aki [1980a] speculated that Q^{-1} for the lithosphere might peak at about 0.5 Hz. Available crustal models of Q_μ^{-1} make it possible to further investigate that possibility. Figure 12 presents Q_μ^{-1} distributions for the lower crust and for the upper crust of stable and tectonically active regions over the frequency range from 0.05 to greater than 1 Hz (continuous curves) and compares them with Q^{-1} for fundamental-mode Rayleigh waves and *Lg*. The Rayleigh wave and *Lg* values are predicted by models where upper crustal Q_μ^{-1} for both stable and tectonically active regions overlie lower crustal Q_μ^{-1} values of 0.001 (Q_μ = 1000). For the 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 (Q_μ = 300 or Q_μ^{-1} = 0.33 × 10^{-3}) and has a value of about 0.001 (Q_μ = 1000) 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 at least as high as 1.2 Hz. Studies of shear wave Q at higher frequencies [*Hough and Anderson*, 1988; *Leary and Abercrombie*, 1994] in California suggest that Q_μ may become frequency dependent at some point, as indicated by the dashed line in Figure 12. The vertical bars in the figure indicate ranges for Q_R^{-1} (at 0.05 and 0.2 Hz) and for Q_{Lg}^{-1} (at 1 Hz) for stable (narrow bars) and tectonically active (wide bars) regions. The low Q_μ^{-1} in the lower crust has 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 that a peak occurs in Q^{-1}, at least for tectonically active regions. Such a peak is, however, not necessary if we use Q_μ^{-1} and remember that those values vary greatly from region to region.

CORRELATION OF UPPER CRUSTAL Q_μ WITH Q_{Lg}^c

Available Q_{Lg}^c and upper crustal Q_μ information obtained at 1 Hz indicates that both are high in stable cratons and both are low in regions of current and recent tectonic activity. For example, in stable portions of eastern North America, Q_μ at 1 Hz is about 1000 [*Mitchell*, 1980] and Q_{Lg} is between 800 and 1200 [*Singh and Herrmann*, 1983], whereas both Q_μ [*Cheng and Mitchell*, 1981; *Mitchell and Xie*, 1994] and Q_{Lg}^c [*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 and India [*Hwang and Mitchell*, 1987] correlate with Q_{Lg}^c in those continents [*Raoof and Nuttli*, 1985; *John*, 1983]. The Q_μ values in those regions are inferred from the attenuation of fundamental-mode Rayleigh waves over the period range from about 5 to 30–100 s. Since wavelengths of fundamental-mode Rayleigh waves at those periods are between about 10 and 300 km and those of *Lg* coda waves are only a few kilometers, the similarity of the patterns of Q_{Lg}^c and Q_μ variation suggests that both waves are predominantly affected by intrinsic anelastic properties of 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 short-period body waves. Body waves have been used to study several small-scale features such as fault zones and geothermal areas [e.g., *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, they are mentioned here because they provide constraints that 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 consists of regions of low Q surrounded by higher-Q material is discussed in a later section.

An interesting possibility which has been proposed on the basis of local Q studies is that temporal variations in Q can be observed that occur over timescales 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 widespread support, they are consistent with the conclusions expressed in this review, that Q_μ can change significantly with time. A model presented in a later section attributes lateral Q_μ variations to differences in the density of cracks and volume of fluids present in different portions of the upper crust. If those parameters can change over geologic 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 Q_μ

Figures 10 and 11 indicate that Q_μ of different models vary by as much as a factor of 10 in the upper mantle at depths reaching 200 km and vary by factors between 3 and 12 in the upper crust, depending on frequency. The low values, both in the upper crust and the upper mantle, occur in regions characterized by high heat flow. It is thus pertinent to ask whether variations in temperature can explain regional differences in Q_μ. 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 at upper mantle depths. Increasing temperatures are thought to produce progressive unpinning of dislocations and, consequently, lower Q_μ values there.

Gueguen et al. [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 $E \sim 440$ kJ mol^{-1} for single crystals of olivine. *Berckhemer et al.* [1982] found values for E of the order of 500–700 kJ mol^{-1} for polycrystalline mantle rocks. If temperature is the only factor which causes differences in Q from region to region and if the relation used by *Gueguen et al.* [1989] is appropriate for the upper mantle, we can write an expression for the ratio of Q at a particular depth in two regions as

$$\frac{Q_{\mu_2}}{Q_{\mu_1}} = \omega^{\zeta_2-\zeta_1} \exp\left[\frac{E}{R}\left(\frac{\zeta_2 T_1 - \zeta_1 T_2}{T_1 T_2}\right)\right]. \tag{7}$$

This equation can be solved for E if it is assumed to be the same in both regions for a particular depth of interest and if Q_μ, ζ, and T are known for those regions. Laboratory measurements indicate values for ζ for mantle rock in the range 0.15–0.3 for ultramafic rocks; consequently, determinations of E using (7) should be made for that range of possible ζ values. We computed E using the continental temperature values of *Pollack et al.* [1993] at a depth of 150 km, where temperatures are in the high-temperature background regime. Using temperatures appropriate for the central United States (1050°C) and for the Basin and Range province (1540°C) at that depth, there are several combinations of ζ values which predict E 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 in the range 500–700 kJ mol^{-1} will result if ζ values are in the range 0.2–0.3. This range has been estimated as being appropriate for 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 0.4–0.6 predict E values in the proper range. This frequency dependence is higher than that suggested by laboratory measurements for upper mantle rock but cannot be excluded for the lower crust and upper mantle in the Basin and Range on the basis of surface wave attenuation [*Mitchell and Xie*, 1994]. Although these calculations do not rule out factors other than temperature as the cause for variations in upper mantle Q, they do show that temperature alone is a plausible explanation for Q_μ variations in the portion of the upper mantle in the high-temperature background regime.

The same calculation was performed for the upper crust at a depth of 10 km. The value of ζ was taken to be 0.5 in the eastern United States and 0.0 in the Basin and Range [*Mitchell*, 1991]. Q_μ was taken to be 800 and 70, respectively, for those regions (Figure 11). Again, using the temperature profiles of *Pollack et al.*

[1993], the activation energy is calculated to be about 10 kJ mol^{-1}, a value which is 2% or less of that for upper mantle rock. It seems too low to be realistic if temperature is the only factor which contributes to the difference in Q_μ between those regions. There is, however, no information on E for granitic rock at crustal temperatures and pressures, so such low values of E cannot be regarded as impossible. If we conclude that the value of 10 kJ mol^{-1} is unlikely, then we must appeal to another mechanism to explain regional Q differences in the upper crust. The motion of fluids in cracks is a mechanism for seismic attenuation in the crust that is consistent with theoretical predictions [*O'Connell and Budiansky*, 1977; *Winkler and Nur*, 1979], laboratory results [*Gordon and Davis*, 1968; *Durham and Bonner*, 1994], and deep drill hole observations [*Borevsky et al.*, 1984; *Morrow and Byerlee*, 1992].

A MODEL FOR REGIONAL Q_μ VARIATIONS

It has long been known from studies of fundamental-mode surface waves [*Mitchell*, 1975; *Cheng and Mitchell*, 1981], as well as from regional phases [*Aki*, 1980b], that seismic attenuation is related to the degree of tectonic activity in continents, tectonically active regions being characterized by lower Q than stable regions. Results discussed earlier in this paper further suggest that Q_μ in the upper crust of any region is directly proportional to the time which has elapsed since the last major tectonic activity there. This section presents a model which will explain these regional and temporal variations of Q_μ. The model is an extension of previously proposed models which explained regional variations of Q_μ in terms of variations in the density of fluid-filled cracks in the upper crust [*Mitchell*, 1975, 1980; *Mitchell and Xie*, 1994].

As mentioned earlier, a model in which the motion of fluids in a network of cracks reduces Q_μ is consistent with laboratory measurements, theoretical predictions, and observations in deep wells. Although limited resolution prevents a precise statement about the depth that fluids in the crust penetrate, our models indicate that that depth is 10–15 km. Higher Q_μ values at greater depths imply that fluids are absent, or that they are much less abundant, at those depths that 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.

Studies of both fundamental-mode surface wave attenuation and Q^c_{Lg} indicate that Q_μ in the upper crust is lowest in regions that have most recently undergone tectonic deformation. This suggests that faults, cracks, and interstitial fluids, which enhance attenuation, are abundant in those regions at the time of, and following, periods of crustal deformation. The fluids may be present because of metamorphic reactions that were produced by elevated temperatures from frictional heating or internal deformation [*Etheridge et al.*, 1984; *Newton*, 1989]. After the cessation of deformation, fluids will be slowly lost either by migration to the surface [*Fyfe et al.*, 1978] or by retrograde metamorphism, cracks will close, and Q_μ will increase. Figure 13 schematically illustrates this process. Results from the continent of Africa (Plate 1) suggest that reductions in observed Q^c_{Lg} represent the effect of tectonic activity that occurred as long ago as the early Paleozoic era.

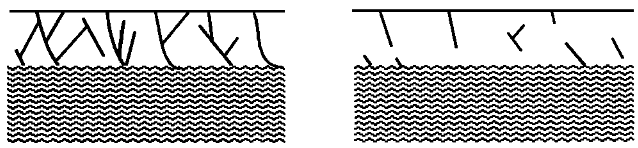

Figure 13. Schematic of a fluid-filled crack model which illustrates how Q_μ in the upper crust may be laterally variable and why it may increase with time following periods of tectonic activity. (left) Crust which has recently undergone tectonic deformation. (right) Crust after sufficient time has passed since deformation so that fluids are absorbed or lost and many of the cracks have closed. Q_μ will be lowest in regions of intense fracturing and enhanced fluid content. The wavy lines represent the plastic lower crust.

Surface wave results alone cannot say much about the distribution of the fractures and cracks which are the likely cause of the reductions in Q_μ proposed in this review. Both direct determinations in deep wells [*Borevsky et al.*, 1984; *Morrow and Byerlee*, 1992] and laboratory measurements [*Durham and Bonner*, 1994] of crustal permeability, however, indicate that fluid motion is not uniform in the crust. Rather, fluid circulation is restricted to fracture zones and confined aquifers. 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 nonuniformly fractured region, because of their long wavelengths, will be affected by fluid movement over a broad region and thus yield low Q_μ, whereas local body wave Q will be low where fractures are present and high where they are absent.

Midcrustal Q_μ is much higher than Q_μ in the upper crust and varies regionally to a much smaller degree. At depths greater than about 25 km, however, Q_μ decreases 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 indicated that differences in temperature could explain regional differences in Q_μ at depths in the mantle where temperature is sufficiently high. Increasing temperatures serve to unpin dislocations, a process which decreases Q and produces a transition to viscoelastic behavior [*Karato and Spetzler*, 1990]. Partial melt could also decrease Q_μ [*Shankland et al.*, 1981], but experimental results at seismic frequencies show that the onset of decreasing Q_μ occurs well before temperature reaches solidus levels [*Jackson et al.*, 1992].

SUMMARY AND CONCLUSIONS

Measurements of surface wave attenuation coefficients and Q^c_{Lg} values have many pitfalls, but modern processing methods allow us to look at their regional variations in some detail. By using these processes and by carefully screening data to avoid systematic errors, we can study not only the anelastic structure of the crust but its tectonic evolution as well.

Q_μ and its frequency dependence in the upper crust of continents vary with geographic region, with observed regional differences of Q_μ exceeding an order of magnitude. The regional differences are related to the tectonic evolution of the continental crust; regions of recent deformation exhibit the lowest Q_μ, and old stable cratons exhibit the highest Q_μ. These differences can cause wide variability in the appearance of broadband seismograms and in the attenuation of fundamental-mode surface waves, L_g, and $L_g$coda.

Regional variations of Q_μ in the upper crust are most easily explained by the movement of fluids in cracks that are abundant during and following periods of deformation and that decrease in number with time following the deformation. Fluid volumes continually decrease, by loss to the surface or by retrograde metamorphism, and seismic wave dissipation is progressively reduced with time. Tomographic maps of Q^c_{Lg} suggest that we can resolve the effect of attenuation caused by tectonic activity that 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 results 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 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 at midcrustal depths may be a widespread feature of continental regions.

In the upper mantle it is possible to explain regional variations in Q_μ by differences in temperature 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 the internal friction of shear waves, 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 know little about possible regional variations of frequency dependence there. In addition, our knowledge of Q_μ in the lithosphere of oceanic regions, which comprise the greatest portion of Earth's surface, is rudimentary. Almost nothing is known of the nature of the frequency dependence of Q_μ in oceanic regions. These and other questions related to crustal and upper mantle anelasticity should be fruitful research areas in years to come.

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