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Propagation and Attenuation Characteristics of the Crustal Phase *Lg*

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Abstract—The *Lg* wave consists of the superposition of *S* waves supercritically reflected, and thus trapped, in the crust. This mode of propagation explains the strong amplitude of this phase and the large distance range in which it is observed. The numerical simulation leads to successful comparison between observed seismograms in stable continental areas and synthetics computed for simple standard crustal models. In regions with strong lateral variations, the influence of large-scale heterogeneities on the *Lg* amplitude is not yet clearly established in terms of the geometrical characteristics of the crustal structure.

The analysis of the decay of amplitude of *Lg* with epicentral distance allows the evaluation of the quality factor of *S* waves in the crust. The results obtained show the same trends as coda *Q*: a clear correlation with the tectonic activity of the region considered, both for the value of *Q* at 1 Hz and for its frequency dependence, suggesting that scattering plays a prominent part among the processes that cause the attenuation.

The coda of *Lg* is made up of scattered *S* waves. The study of the spatial attenuation of the coda indicated that a large part of the arrivals that compose the coda propagate as *Lg*. The relative amplitude of the coda is larger at sites located on sediments because, in these conditions, a part of *Lg* energy can be converted locally into lower order surface modes.

Key words: Regional seismograms, attenuation, coda, crustal structure, surface waves.

Interpretation

PRESS and EWING (1952) proposed the symbol *Lg* for the short-period continental phase they first identified on paths across North America. They interpreted this phase as a surface shear wave which propagates in a superficial waveguide consisting of the "granitic layer". They found this interpretation consistent with their observation that *Lg* waves disappear along paths across zones of oceanic structure where the granitic layer is absent. LEHMANN (1953) presented further observations and confirmed the extinction of *Lg* waves when part of the path is oceanic. Nevertheless, she pointed out that the *Lg* wave was associated with large arrivals even for earthquakes occurring in the lower crust, which led her to abandon

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the initial interpretation of *Lg* as a wave guided strictly in the upper crust. BÅTH (1954) presented detailed observations of *Lg* along Eurasiatic paths and introduced the concept of channel wave to explain the strong amplitude of *Lg* and the large distance range in which these waves are observed. Following GUTENBERG (1951), Båth assumed the presence of a low-velocity layer within the crust, at a depth of about 15 km, in which most of the earthquakes would occur. *Lg* would then be the continental counterpart of the *T* wave propagating in the oceans. GUTENBERG (1955) developed a complete scheme of interpretation of regional phases by the study of the propagation of channel waves in a lithosphere model with low-velocity zones.

In spite of the success of the concept of channel waves, OLIVER and EWING (1957, 1958) showed that the higher-mode surface-wave theory explains the presence and the arrival time of *Lg* waves without requiring the existence of a low-velocity layer in the crust. Identical conclusions were presented by KOVACH and ANDERSON (1964). Because of the similar behavior of the dispersion of higher modes of Love and Rayleigh waves for short periods, observed *Lg* characteristics must be roughly the same for the three components of motion. KNOPOFF *et al.* (1973) presented synthetic seismograms at low frequency computed from higher-mode Love waves at large epicentral distance. The synthetics display the main characteristics of both arrival time and amplitude of actual *Lg* waves at low frequencies. Similar results were obtained by PANZA and CALCAGNILE (1975) for the vertical and radial components of motion by using a summation of higher Rayleigh modes. Even in the low-frequency domain concerned by these studies, the results obtained did not account for energy travelling at group velocity less than 3.1 km/s although *Lg* is observed to have large amplitude in the group velocity window of 3.5–2.8 km/s (RUZAIKIN *et al.*, 1977; CARA *et al.*, 1981).

Recently, advances in numerical modeling techniques have made possible realistic simulation of elastic wave propagation in vertically heterogeneous crustal structures in a wide frequency range (e.g., KIND, 1978; BOUCHON, 1981; WANG, 1981). BOUCHON (1982) presented numerical modeling of short-period seismograms at regional distance, computed by the discrete wavenumber method (BOUCHON,

Table 1

Crustal model

Layer Thickness (km)	*P*-Wave Velocity (km/sec)	*S*-Wave Velocity (km/sec)	Density (gm/cm^3)
2	4.5	2.6	2.6
16	6.0	3.5	2.8
6	6.3	3.65	2.9
6	6.7	3.9	3.1
	8.2	4.7	3.3

1981). Figure 1 presents a comparison of synthesized and recorded vertical seismograms for a strike-slip event of magnitude 4.1 which occurred at a depth of 15 km near Saint Pourcain sur Sioule (central France) on February 11, 1978. The actual seismograms are digitally recorded on short-period instruments whose natural period is 1 s. The synthetics have been corrected to incorporate the instrumental response. The crustal model consists of a stack of flat layers whose physical properties are inferred from deep seismic sounding (Table 1). The very good agreement between records and synthetics shows that the mechanism of *Lg* propagation in a stable area can be completely described by a flat-layered structure. In particular, the duration of the theoretical *Lg* wavetrain represents roughly the duration for which the observed seismograms have large amplitude, i.e., in the group velocity window 3.5–2.8 km/s.

Bouchon interpreted the *Lg* wave as a superposition of *S* waves that are multiply reflected within the crust and incident on the Moho at an angle more grazing than the critical angle. The entire energy of these waves remains trapped in

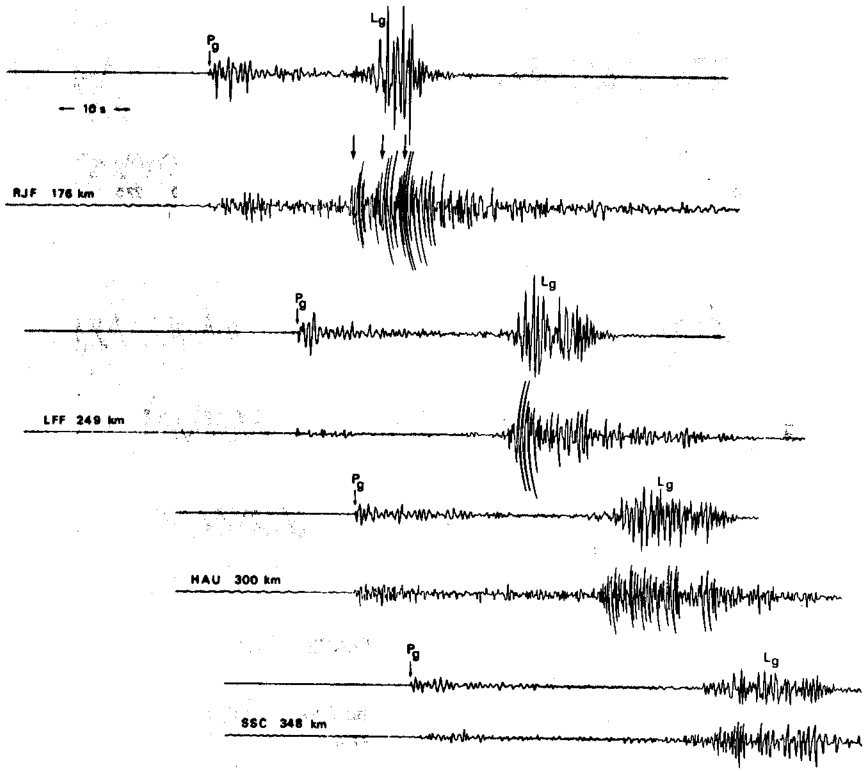

Figure 1
Comparison of synthesized (upper trace) and recorded vertical seismograms at four stations. The epicentral distances are indicated on observed traces (from BOUCHON, 1982).

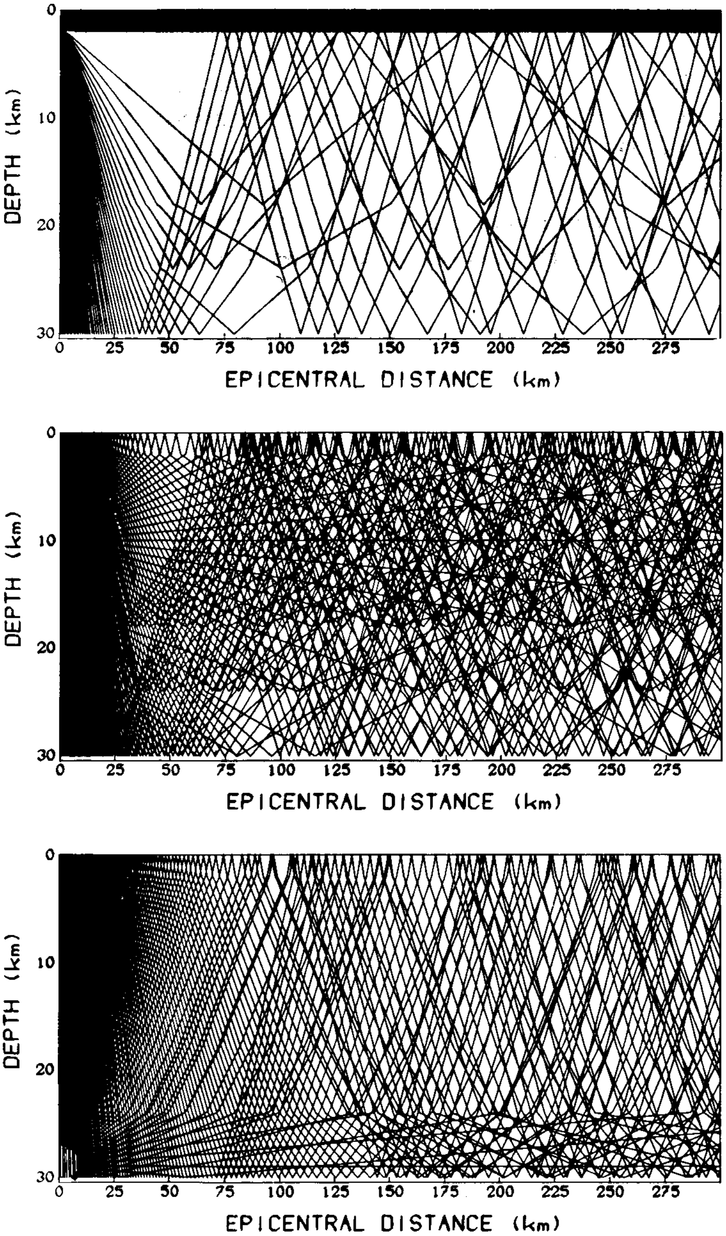

Figure 2
Supercritically reflected rays radiated by source at 0-, 10- and 29-km depth for the crust model given in Table 1 (from CAMPILLO *et al.*, 1985).

the crust. OLSON *et al.* (1983) gave a similar interpretation. For the shortest epicentral distance presented in Figure 1 (station RJF), the three large-amplitude wave groups which dominate the *Lg* record (indicated by arrows) can be associated with individual paths of *S* waves emitted downward and reflected on the Moho or emitted upward, reflected at the free surface and then on the Moho and for the last group, emitted downward, reflected a first time on the Moho, then at the free surface and a second time on the Moho. Of course, this simple scheme is widely complicated by multiples in the different layers, conversion and source directivity. Another source of the apparent complexity of the *Lg* wavetrain comes from the supercritical reflections on the different discontinuities of the crust. This can be illustrated by a ray diagram showing the different rays supercritically reflected within the crust (Figure 2) and thus which contribute to building up the crustal-guided wave. These simple calculations show that the sampling of the crust by *Lg* waves (the density of rays) is very dependent on the source depth. For a source in the middle crust the ray diagram indicates that numerous reflections from internal crustal discontinuities may interfere with the reflections on the Moho. From further numerical simulations (CAMPILLO *et al.*, 1985), we have shown that the later part of the computed *Lg* phase is the part of the wavetrain where the arrivals are the most representative of the propagation in the bottom crustal layer.

HERRMANN and KIJKO (1983) used higher-mode surface-wave theory to compute synthetic seismograms for a 4-layer crustal model. They showed that the simulation can predict some empirical vertical component *Lg* relations, such as spatial attenuation, if anelastic attenuation is included in the computation. Synthetic seismograms are able to account for the observed mean ratio of maximum horizontal to vertical motion and its large standard deviation (CAMPILLO *et al.*, 1984).

The agreement between the early part of *Lg* of actual records and synthetic seismograms, reported in these papers, indicates that the major characteristics of *Lg* are well understood in terms of elastic wave propagation in a stack of flat homogeneous layers, corresponding to the crustal structure inferred from other geophysical investigations. As is commonly observed for body waves, the primary phase is followed by a coda which cannot be generally modelled in a deterministic manner. The nature of *Lg* is clear enough, however, to use this phase for quantitative studies of the internal properties of attenuation of the crust or source parameter evaluation.

Spatial Attenuation of Lg-waves

Different problems must be considered to discuss the spatial attenuation of *Lg*. A first distinction concerns the time window studied in the *Lg* wavetrain. In the previous section, we pointed out that the main features of *Lg* in the group velocity window 3.5–2.8 km/s can be theoretically predicted for standard flat-layered crustal models, at least in stable continental areas. The effects of lateral variations of the structure will be discussed later. For group velocities smaller than 2.8 km/s the interpretation remains rather doubtful, but the mechanism of formation of these arrivals is probably essentially similar to the one of coda waves observed for local earthquakes. Because *Lg* propagation is the most efficient mode of seismic energy transport on large distance in the frequency range 0.5–10 Hz, a coda of scattered *S* waves propagating as *Lg* can be built up all along the path.

Considering the early part of the *Lg* wavetrain as a primary phase, i.e., the time window defined by the group velocity window 3.5 to 2.8 km/s, in which generally the maximum amplitude of the phase occurs, we can compute the anelastic attenuation coefficient or the quality factor of *S* waves from the observed decay of amplitude with epicentral distance, if we are able to evaluate theoretically the geometrical attenuation of the phase. This can only be achieved under restrictive assumptions about the lithospheric model.

The simplest approach entails the evaluation of the geometrical spreading in a flat-layered crust, corresponding to the mean structure encountered along the path. If *Lg* is a purely guided wave, it will be weakly sensitive to intracrustal heterogeneities; but this approach must be limited to stable continental areas. As we saw in the previous section, *Lg* can be viewed as a superposition of surface waves; therefore, its geometrical spreading in the frequency domain is inversely proportional to the square root of distance, assuming a purely elastic propagation in a schematic model. Nevertheless, the total energy of such a phase is very difficult to measure because of the noise, of the presence of scattered waves and of the surface waves due to the mode conversions which may occur near the station (DER *et al.*, 1984; CAMPILLO, 1987). One may be concerned that the apparent anelastic attenuation obtained from a measurement of total spectral energy is significantly different from the one evaluated for primary *S* waves.

Another approach involves using the amplitude in the time domain or the spectral density per time unit in a window chosen to contain the early energetic part of the wavetrain. In this case, we may expect to measure a crustal mean value of the quality factor of *S* waves that can be compared with results obtained from other techniques. Because the spectrum of the *Lg* wavetrain in the elastic response of a flat-layered crustal model to an impulsive double-couple source was found by HERRMANN and KIJKO (1983) to be remarkably flat over the frequency range 0.1–10 Hz (Figure 3), the geometrical spreading in the time domain must be used to correct spectral density per time unit. Because the signal duration increases with epicentral distance, the geometrical attenuation of *Lg* is greater in the time domain than in the frequency domain. Following the higher-mode surface-wave representation, NUTTLI (1973) proposed a geometrical decay of amplitude in the time domain with epicentral distance of the form:

$$A = A_0 r^{-(5/6)}$$

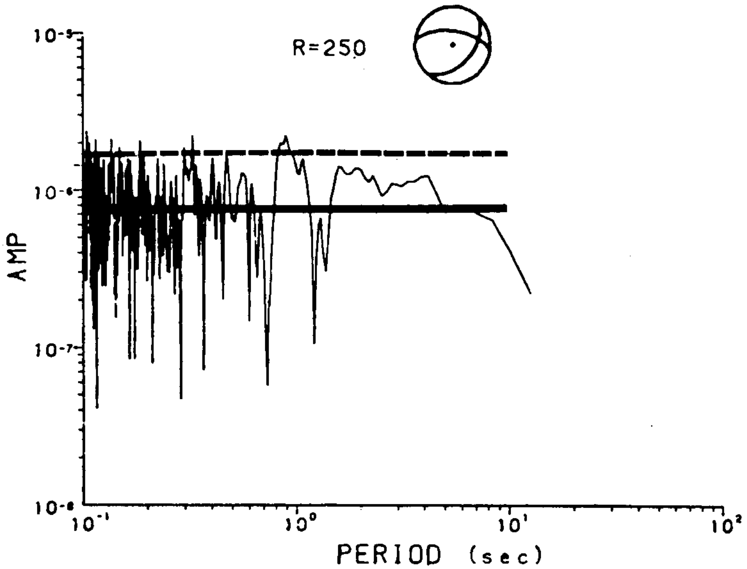

Figure 3
Fourier amplitude spectra of the vertical component *Lg* wave for a receiver at a distance of 250 km. The focal mechanism is shown in the inset. The two heavy lines correspond to the logarithmic mean and to the logarithmic mean plus one standard deviation (from HERRMANN and KIJKO, 1983).

corresponding to the theortical decay of a dispersive Airy phase. A very similar expression of the geometrical spreading was obtained from synthetic seismograms in the form:

$$A = A_0 r^{-0.83}$$

by CAMPILLO *et al.* (1984) who showed that the decay rate is independent both of the source mechanism and the source depth for an event within the crust (Figure 4). It is important to note the large scatter of the theoretical values of the amplitudes of *Lg* with respect to a suitable functional expression of the decay with epicentral distance. This is due to the successive appearance and disappearance of interferences between the different supercritically reflected waves.

Nevertheless, a stack of flat homogeneous layers is an oversimplified model of the earth crust and upper mantle. Even in a stable, purely continental area, one can expect to encounter lateral variations of structure along a path of several hundreds of kilometers. How the *Lg* amplitude can be affected by the passing through different geological units is not clearly established in terms of the geometrical characteristics of the lateral variations or in terms of the local internal properties of the crust. In their pioneering paper, PRESS and EWING (1952) remark: "roots of mountains systems and others thickness variations of limited lateral extent would not prevent transmission of *Lg*" and effectively *Lg* wavetrains are mostly observed to propagate across continental areas.

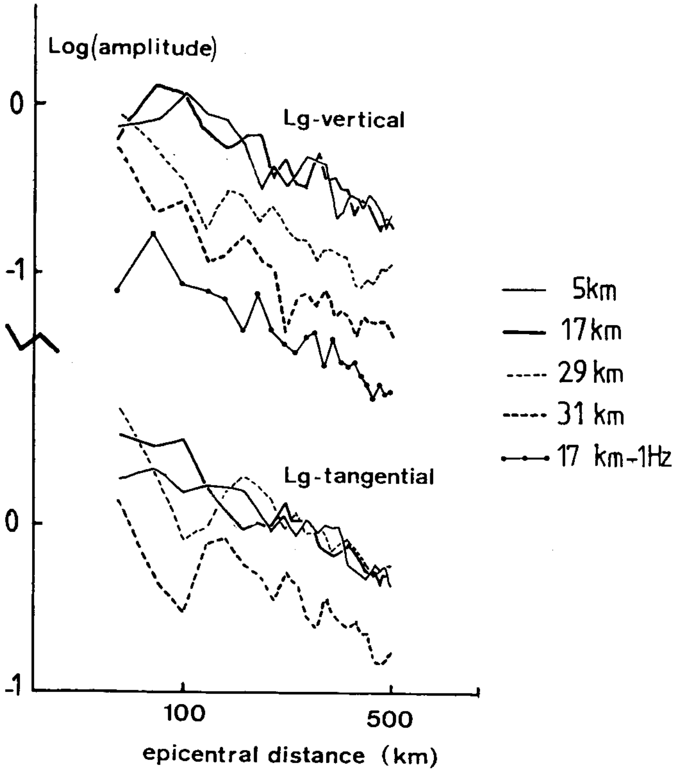

Figure 4
Geometrical attenuation curves for a vertical strike-slip source at different depths. In the last case the instrumental response of a 1 Hz seismometer is included (from CAMPILLO *et al.*, 1984).

In contrast, several experimental results show the existence of zones where *Lg* propagates very poorly, where strong weakening and even complete extinction were reported. Examples of vanishing were brought out for paths crossing part of the Tibetan plateau (RUZAIKIN *et al.*, 1977; NI and BARAZANGI, 1983) without a definitive explanation of this extinction: presence of a very low-*Q* zone within the crust or propagation effects related to abrupt changes in the structure. CHINN *et al.* (1980) observed that the efficiency of the *Lg* propagation is better for paths parallel to the structural trend of the Andes than for paths in the perpendicular direction. This point indicates that the weakening may be due primarily to changes in the crustal (or subcrustal) structure rather than to the properties of the materials of the crust in this region. Geographic variations of the quality of the *Lg* propagation across the Turkish and Iranian plateaus were reported by KADINSKY-CADE *et al.* (1981). GREGERSEN (1984) presented evidence of anomalies of propagation of *Lg* in northern Europe. KENNETT and MYKKELTVEIT (1984) and KENNETT *et al.* (1985) showed that the anomalous zone corresponds to the central North Sea graben.

In most cases, the quantitative identification of the respective parts played by large-scale structural heterogeneities and by the properties of the materials constituting the crust, remains almost impossible, because of the lack of information both on the structure of the lower crust and on the theoretical behavior of *Lg* propagating through a laterally heterogeneous medium. Particularly, the absence of certainty about the deep structure of recent mountain ranges makes premature further attempts of discrimination between the different causes of dissipation of energy in these regions. Nevertheless, in more stable areas, where a mean flat-layered model perturbed by local heterogeneities can be considered as a reasonable view of the actual structure, numerical tests can be performed. Simulations were made to evaluate the effect on *Lg* propagation of simple cases of lateral heterogeneities which can be encountered, such as variations of the Moho depth or changes in the thickness of the sedimentary layers. To date, these studies are limited to the two-dimensional case.

In the case of *SH* wave propagation, KENNETT (1984) presented a technique based on the representation of the wavetrain as a superposition of the model contributions for the flat-layered reference structure. The effect of a lateral variation is thus characterized by the matrices of reflection and transmission coefficients of the different modes at a given frequency. These computations showed the occurrence of a large amount of mode conversions when *Lg* propagates through a 10-km wide zone with an uplift of the Moho 6 km high. KENNETT (1984) proposed that the strong mode conversions which occur in the presence of lateral variations of the Moho depth may be responsible for the anomalies of *Lg* propagation reported across the North Sea. CARA *et al.* (1981) performed multimode analysis of *Lg* in southern California and showed the existence and the importance of converted modes in their records.

CAMPILLO (1987) computed synthetic *SH* seismograms in a layered model, including a smooth variation of the depth of the Moho over a 40-km extent and in the form of an uplift 10 km high. The technique used for the computation relies on the discrete wavenumber decomposition of wavefields and on the representation of the diffracting boundaries by source distributions (CAMPILLO and BOUCHON, 1985). The results show that the presence of the bump of the Moho affects widely the *Lg* waveshapes while the amplitude of the wavetrain is not significantly changed with respect to the case of the flat-layered crust. This suggests that mode conversions occur mainly between modes associated with guided-wave propagation in the crust, for such a smooth model of deep heterogeneity. This result is in good agreement with the fact that *Lg* wavetrains are mostly observed to propagate across stable continental areas where, probably, the thicknesses of the crust and the sedimentary cover undergo lateral variations.

In conclusion, the extensive use of the *Lg* phase to measure the quality factor of *S* waves is limited by our poor knowledge of its geometrical attenuation in a heterogeneous crust and thus is restricted to relatively stable areas. The actual limitation, however, is not yet established in terms of the geometrical characteristics of the crustal structure, although there are several examples of the capability of modeling *Lg* as *S* waves in flat-layered models (e.g., BOUCHON, 1982; OLSEN *et al.*, 1983; KIM, 1987).

Evaluation of the Quality Factor from Lg

For stable continental areas, *Lg* wave analysis represents a simple and robust technique to evaluate the quality factor of *S* waves. Because *Lg* propagates only in the crust and samples roughly the entire crust, the quality factor deduced from the attenuation of this phase can be regarded as the mean value of *Q*s in the crust, after some precautions are taken with respect to the source depth and the time window considered in the wavetrain.

A first approach entails evaluating *Q*s from the decay of *Lg* amplitude with epicentral distance. The decay can be measured either in the time domain or in the frequency domain. The measurement is made viewing *Lg* as a primary wave i.e., without consideration of the existence of scattered waves. Such studies were carried out for different regions of the world like North America (NUTTLI, 1973, 1982; STREET, 1976; JONES *et al.*, 1977; BOLLINGER, 1979; MITCHELL, 1980; DWYER *et al.*, 1983; ESPINOSA, 1984; HASEGAWA, 1985; CHAVEZ and PRIESTLEY, 1986), western Europe (NICOLAS *et al.*, 1982; CAMPILLO *et al.*, 1985) or Asia (NUTTLI, 1980, 1981). The results show good agreement with those obtained by coda analysis or by the spectral-ratio technique.

We have verified through numerical simulation that the quality factor measured from *Lg* is a mean value of *Q*s in the crust, in the case of earthquakes in the middle crust (CAMPILLO *et al.*, 1985). We considered *Lg* records of 18 earthquakes in or around France at 22 short-period vertical seismic stations. The dependence of *Lg* amplitude on the focal mechanism is neglected. Because *Lg* is made up of a superposition of rays which sample a wide range of take-off angles, the azimuthal dependence is weaker for this phase than for direct waves, even for a deterministic model of propagation. We computed by least-squares fits source excitations, a frequency-dependent quality factor and station responses that account for local structure effect. Figure 5 is a comparison between computed and observed decay of spectral energy with distance for different frequencies. The theoretical curves are normalized so that the mean values (in the least-squares sense) of the observed and synthetic spectra are the same. The good agreement observed at the different frequencies between predicted and actual values shows that *Lg* attenuation corresponds to crustal shear-wave attenuation.

The depth dependence of *Q*s is difficult to infer from *Lg*. Because we found, by numerical simulation, that the later part of the *Lg* wavetrain is the most sensitive to

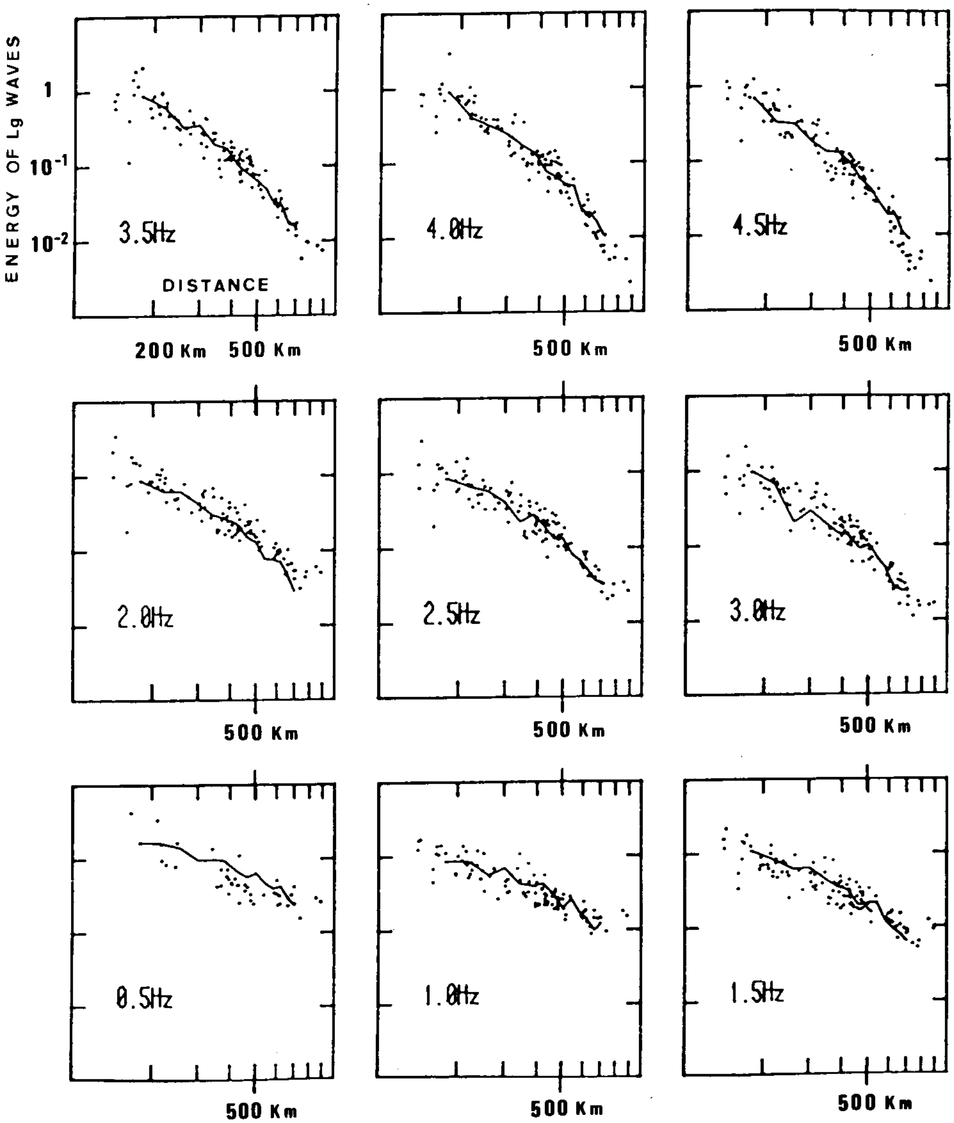

Figure 5

Decay of spectral density of energy of *Lg* waves with distance for different frequencies. Comparison between numerical simulation (solid lines) and observed points (from CAMPILLO *et al.*, 1985).

the attenuation properties of the lower crust, we performed the same data analysis in different time windows. The results show that the apparent quality factor weakly increases with decreasing group velocity (for example at 1 Hz, Qs is 290 in the window 3.6–3.1 km/s and reaches 350 in the window 2.6–2.3 km/s) and thus could indicate that Qs would be greater in the lower crust. The frequency dependence inferred for the different time windows is similar:

$$Q = Q_0 f^{-0.50}.$$

Comparing the results obtained in the Great Basin from earthquake and explosions, viewed as a deep and a shallow source, CHAVEZ and PRIESTLEY (1986) proposed that the quality factor presents a greater frequency dependence in the upper crust than at depth in the region they studied.

As pointed out by ESPINOSA (1984) and HERRAIZ and ESPINOSA (1986) for the United States, the regional values of the apparent quality factor show a definitive correlation with the geological characteristics of the areas considered. Figure 6 summarizes different functional dependences of Qs on the frequency as inferred from Lg. The results obtained in the different regions from Lg are in reasonable agreement with the regionalization proposed by SINGH and HERRMANN (1983) for coda Q, considering both the value of the quality factor at 1 Hz and its frequency dependence. We attempted to compute the spatial distribution of Qs from a homogeneous data set in order to avoid regional effects due to the structure close to the source. The results that we obtained for France (CAMPILLO, 1987) confirm that tectonic regions are associated with low values of Qs at 1 Hz and with strong frequency dependence of the quality factor while, in contrast, stable areas have high

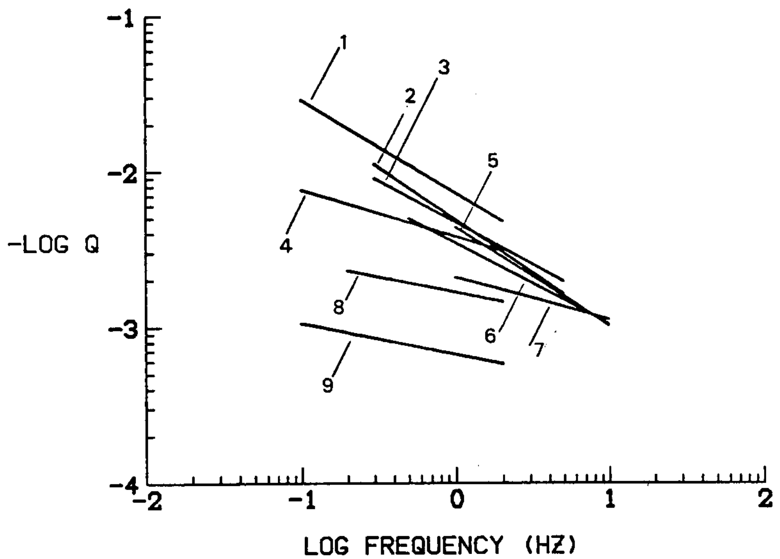

Figure 6

Qs as a function of frequency from different studies. The curves 1 to 4 concern the western U.S. The curves 5, 6 and 7 were established for Massif Central, central France and Paris basin, respectively. The curve 8 is for the eastern U.S. and 9 for the central U.S. (1: NUTTLI, 1986; 2–3: CHAVEZ and PRIESTLEY, 1986; 4: PESECKIS and POMEROY, 1984; 5–7 CAMPILLO, 1987, 6: CAMPILLO *et al.*, 1985; 8: MITCHELL, 1981; DWYER *et al.*, 1983).

values of Qs and a weak dependence on frequency. At a frequency of 10 Hz, the regional trends disappear, showing that the peak of absorption due to scattering around 1 Hz (AKI, 1982) is the most important phenomenon for attenuation of S waves in the crust in the short-period domain.

PESECKIS and POMEROY (1984) and NUTTLI (1986) used the technique of coda analysis developed by HERRMANN (1980) to evaluate Q from *Lg* coda. These authors proposed that the value of Q measured from the coda of regional records (several hundred kilometers of epicentral distance) is accurate to correct the seismograms from propagation effects and, thus, to compute the source excitation. Under this statement is the assumption that the coda of *Lg* is generated continuously, and nearly homogeneously, in a region along the path between source and receiver; so Q-coda could be identified to a mean value of Qs along the path. The values obtained by PESECKIS and POMEROY (1984) and NUTTLI (1986) are significantly different from those obtained in the same region (the Great Basin) by CHAVEZ and PRIESTLEY (1986), who used the spatial decay of spectral amplitude. This discrepancy may be explained by the origin of the coda of *Lg*.

Origin of Lg Coda and Site Effects

DER *et al.* (1984) presented analysis of *Lg* records at different seismic arrays that indicated the presence of a significant amount of omnidirectional scattered waves which appears early in the wavetrain. The interpretation of this coda is difficult because of the very striking dependence of its amplitude on the local geological characteristics of the station (BAKER *et al.*, 1981). From the array analysis made for *Lg* for different positions of the time window in the wavetrain by DER *et al.* (1984), it appears that the spatial coherency of the phase falls when the amplitude is still one half or one-third of its maximum. Nevertheless the waves of the coda present a certain directivity; if the arrivals are not limited to the precise direction of the source, the F-K spectrum analysis shows that the azimuthal distribution is far from being random (DER *et al.*, 1984). This suggests that the coda is made up of arrivals scattered mainly forward with respect to the source-receiver direction.

Two types of mechanisms of generation of the coda can be invoked to explain these properties. First, the scattering of the S waves which compose the *Lg* wavetrain produces S waves some of which can be also trapped in the crust and thus can propagate very efficiently. The travel paths of the guided scattered waves present at a given time in the *Lg* coda are contained in a prolate ellipse, because of the large epicentral distance considered. So, as *Lg* can propagate over long distances, the resulting phenomenon presents an apparent directivity and the coda seems to be gradually made up by multipathing in a region around the primary source-receiver path. Secondly, the presence of rapid variations of the shallow crustal structure, brings about local mode conversions that can produce late arrivals with large amplitudes. This second mechanism is associated with a type of propagation that does not allow the transport of energy over large distances because of the strong variability of the superficial structure and of the large attenuation in the shallow materials. In contrast, the first mechanism invoked, scattering of *S* waves resulting in an apparently forward scattering of *Lg*, is based on the most efficient mode of propagation over large distances for the frequency range considered here, i.e., the mode of propagation of *Lg*.

The hypothesis that a prominent part is played by random scattering of *Lg* in forward directions is supported by the stability of the measurements of the apparent attenuation deduced from amplitude decay with distance in different group velocity windows, including the coda (CAMPILLO *et al.*, 1985). In this study, we assumed the same mode of propagation for the different group velocity windows, i.e., the same geometrical spreading, for example. If the *Lg* coda was built up by homogeneously backscattered *S* waves, as is commonly accepted for the coda of local earthquakes, the trade-off between constant lapse time and constant group velocity would change with epicentral distance and thus would introduce a bias in the evaluation of apparent *Q* that would have led to an underestimation. On the contrary, our results show a weak increase in the value of the apparent quality factor with decreasing group velocity, which confirms that most of the coda consists of waves propagating as *Lg*.

Nevertheless, striking site effects on the level of the *Lg* coda have been reported (BAKER *et al.*, 1981). Figure 7 presents short-period vertical records of the same earthquake at two close stations, one located on granite (OB2NV) and the other in a deep alluvial valley (YF4NV). The unfiltered seismograms illustrate the sensitivity of the *Lg* amplitude, and of the coda level, on the type of surficial material below the station. The properties of the seismic response of alluvial valleys (BARD and BOUCHON, 1980a,b) produce the higher amplitudes and longer coda observed at YF4NV, as pointed out by BAKER *et al.* (1981). The band-pass filtered seismograms are also shown on the figure. One can notice that the coda is stronger in the frequency bands below 1.5 Hz than for the higher frequencies on the records at the two stations.

In the previous example, the coda amplification is due to a shallow structure associated with low-velocity material. The numerical simulation of *SH*-wave propagation shows that mode conversions can occur in the presence of lateral variations of the thickness of the sedimentary layer and thus produce large amplitude secondary arrivals (WANG and HERRMANN, 1984; CAMPILLO, 1987). These conversions give rise to large surface amplitude but they do not affect significantly the energy of the phase propagating at depth. Such a phenomenon can therefore be considered as a local effect.

In order to evaluate the importance of coda generation near the receiver, we have measured the spectral amplitudes of *Lg* in the group velocity windows

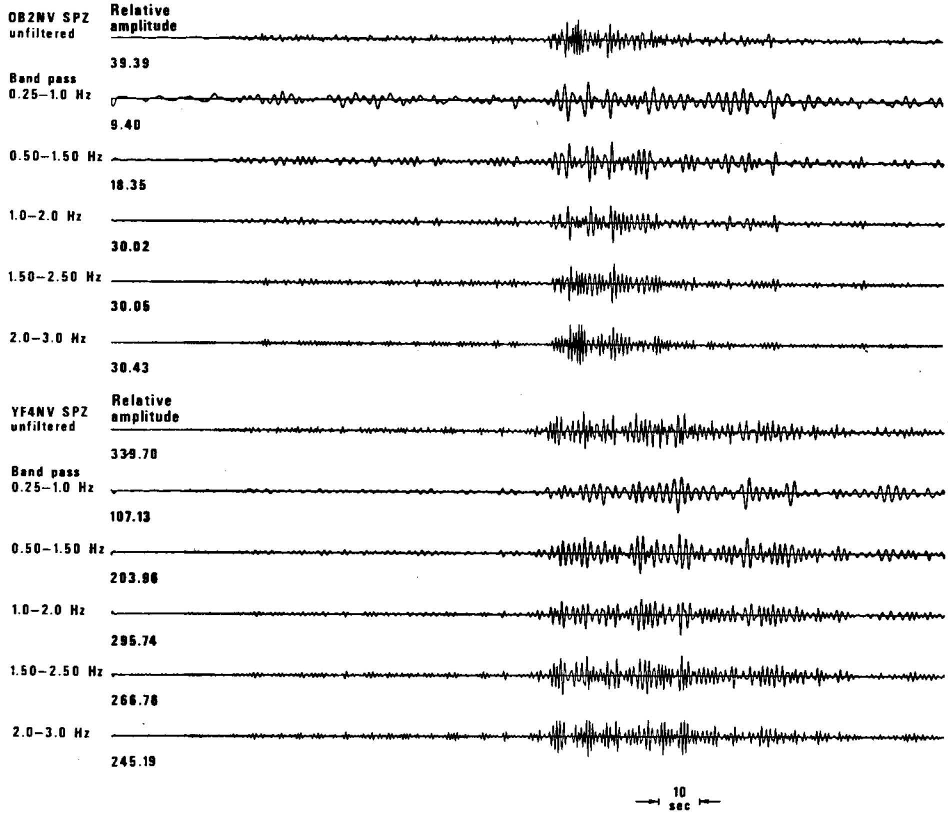

Figure 7
Comparison of band-pass filtered vertical seismograms at the two close stations OB2NV and YF4NV located respectively on granite and in a deep alluvial valley (from BAKER *et al.*, 1981).

3.6–3.1 km/s (*Lg*1) and 2.6–2.3 km/s (*Lg*3) for a series of earthquakes recorded at different stations of the French seismic network. We have selected the data for which the signal-to-noise ratio is greater than 4. We have computed the ratios between the amplitudes of the two windows for records corresponding to epicentral distances ranging between 360 and 440 km. Figure 8 presents the mean values of the ratio *Lg*1 over *Lg*3, as a function of frequency, for the entire set of stations, for the stations located on granite and for the stations located on a sedimentary cover. The results indicate in all cases that the coda is more efficiently generated at low frequency and that the ratio increases only slightly with frequency above 4 Hz. The stations located on sediments present a relative amplitude of the coda systematically higher than those located on granite. Nevertheless, the shapes of the curves are very similar between 1 and 5 Hz. This can be explained because the amplification due to sediments, usually frequency-dependent, is averaged over a large set of incidence angles, azimuths and finally over different sites.

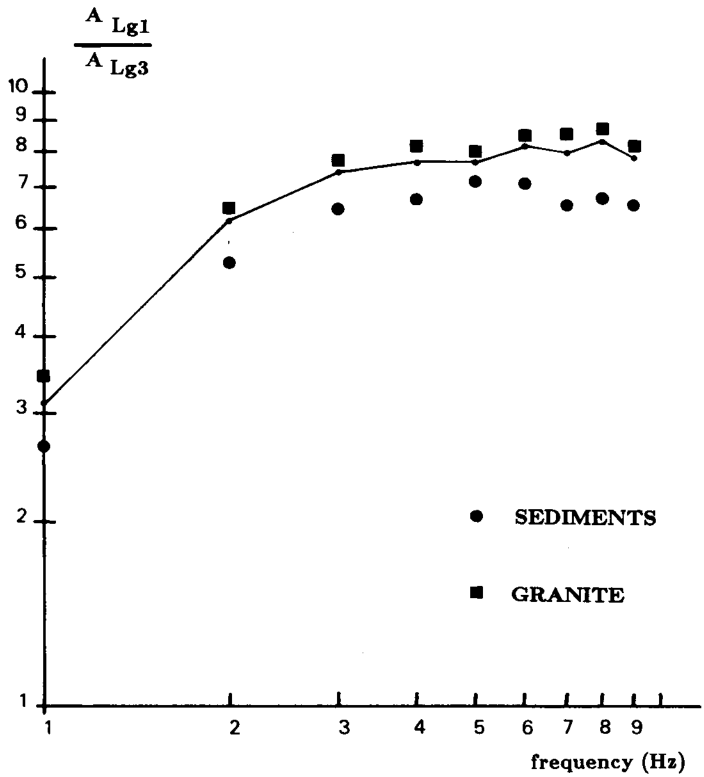

Figure 8
Mean amplitude ratio between $Lg1$ and $Lg3$ as a function of frequency at a mean epicentral distance of 400 km (continuous line) for the stations of the French network. The mean ratio obtained for the stations located on granite and sediments are also shown.

Both mechanisms, invoked previously to be the cause of the *Lg* coda, seem to contribute to it. The observed frequency dependence of the mean amplitude ratio between primary waves and coda for *Lg* is a characteristic of the process of generation of the coda. A main feature is the strong decrease of coda amplitude with increasing frequency between 1 and 3 Hz. This is due to the presence of direct Rayleigh waves (NUTTLI, 1986). In contrast, the behavior at higher frequencies can be explained if we asume that the *Lg* coda is basically built up by scattered *S* waves. From single-scattering theory and accepted values of the scales of heterogeneities in the crust, the backscattering coefficient should be almost independent of frequency above 3 Hz (WU and AKI, 1985). Therefore, the ratio between the amplitude of the primary phase and the coda is simply increasing with frequency because of the apparent attenuation of the medium. The observed increase is in agreement with the measured values of the quality factor in central France.

Conclusion

Lg is a seismic phase that can propagate to large epicentral distances in continental areas and which comprises *S* waves trapped within the crust. In regions where a stack of flat layers is a reasonable model of the crust and upper mantle, this mode of propagation is clearly established by the comparison between observed and theoretical properties of this phase. Further theoretical studies are necessary to allow a quantitative interpretation of *Lg* propagation across regions with strong lateral variations of structure.

Lg amplitude analysis is a powerful tool for the measurement of the apparent attenuation of *S* waves owing to the high efficiency of the elastic propagation of this phase over long distances and because its domain of propagation is limited to the crust. Array analysis shows that the coda of *Lg* comprises scattered waves. The relative parts played by remote scattering, local backscattering and site effects remain widely unknown. These relative importances are probably very dependent on the amplitude and scale of heterogeneity of the crust in the region considered. For example, the spatial attenuation in a fixed group-velocity window indicates that a main part of the arrivals in the coda propagates as *Lg* waves, in the particular case of a set of records in France. The relative amplitude of the coda is enhanced at sites located on sediments, with respect to sites on granite, because of the local conversion of *Lg* into lower order surface modes.

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