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Influence of the Lower Crustal Structure on the Early Coda of Regional Seismograms

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We use synthetic seismograms for a flat-layered Earth model to study the influence of the widely observed strong reflectivity of the lower crust on the propagation of seismic regional phases. The lower crustal layering induces a clear increase in the duration of the Pg and Lg phases on both the vertical and radial components beyond the group velocity of their terminations in a model with a homogeneous lower crust. The transverse component is weakly affected. The amplitude and the spectral content of the early Lg coda are strongly dependent on the distribution of layer thicknesses. The predominance of the low-frequency components (below 2 Hz) observed in the Lg coda of real data is correctly accounted for with 1-km-thick layers in the lower crust. From particle motions at various depths and the analysis of the dispersion curves of normal modes, we hypothesize that the late arrivals result from the interaction of plate-type disturbances of the high-impedance layers. These waves leak energy in the mantle at a low rate. This mode of generation of the coda waves explains the puzzling observation that the spreading coefficient of Lg seems to be applicable to the early coda of regional seismograms. The study of the influence of the lamellae on the average Q factor of the continental crust shows that the lower crust layering does not account for the observed low values of the apparent Qs. Another origin is needed for this strong attenuation, for example scattering by randomly distributed heterogeneities.

INTRODUCTION

Since the introduction of the coda method by *Aki and Chouet* [1975] to measure apparent attenuation in the lithosphere, numerous measurements and applications have been carried out including the study of spatial [*Singh and Herrmann*, 1983] and temporal variations of the quality factor [*Aki*, 1985]. The background of the method is the theory of single scattering in an elastic space with isotropically distributed inhomogeneities [see *Sato*, 1990]. In this paper we shall also discuss the effect of crustal heterogeneities. However, the framework of our modeling is quite different since we consider the crust as a stack of flat layers. Although this representation is oversimplified, it allows us to model easily the complete phenomenon of wave propagation, including multiple scattering. Furthermore, the fine structure of the crust in western Europe presents a distribution of heterogeneity which, in some aspects, is closer to a horizontal layering than to an isotropic distribution, especially in the lower crust.

We shall investigate the influence of the thin layering of the lower crust revealed by deep reflection profiles on the propagation of regional phases. In particular, we shall study the effect of the lamination on the duration of the seismograms and the consequences of this structure on the measurements of apparent attenuation using regional records in France. The influence of a series of thin layers with variable thickness on wave propagation has already been extensively investigated [e.g., *Menke*, 1983; *Mallick and Frazer*, 1990]. However, the contexts were different from the study of regional phases presented in this paper.

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SOME ATTENUATION CHARACTERISTICS OF REGIONAL PHASES

The regional phases Pg and Lg are one of the most characteristic features of the short-period seismograms recorded in the continental domain. These waves are observed from distances as close as 150 km up to several thousand kilometers of epicentral distances. They disappear along oceanic paths [*Press and Ewing*, 1952] and at some very strong structural boundaries [e.g., *Ruzaikin et al.*, 1977; *Kennett et al.*, 1985]. The interpretation of these arrivals as P and S waves multiply reflected within the crust is supported by the success of the numerical modeling of these phases (see *Campillo* [1990] for a review of recent results). These models were based upon simple crustal structures as deduced from deep seismic soundings. For example, *Bouchon* [1982], *Kim* [1987], and *Bertil et al.* [1989] have presented realistic seismograms obtained with models consisting of a few flat homogeneous layers. However, these computations fail to explain the long duration of observed seismograms. Figure 1 presents examples of records of vertical motion at stations of the Laboratoire de Détection Géophysique (LDG) from an earthquake in the region of Lacq (southwestern France). The paths are purely continental and propagate through the Hercynian province of Massif Central. They are compared to synthetic seismograms computed for a simple model consisting of four flat homogeneous layers. All computations of synthetic seismograms presented in this paper are performed using the discrete wavenumber decomposition [*Bouchon*, 1981] associated with the reflection-transmission matrix technique of *Kennett* [1983]. The source consists of a point strike-slip dislocation at 10 km depth. The receivers lie in an azimuth of 30° with respect to the fault plane. Seismograms are computed in the frequency range 0–10 Hz. The source function corresponds to an omega-square model with corner frequency at 1 Hz, and the seismograms are convolved with the LDG station response. The comparison between synthetics and records shows that the model succeeds in predicting the

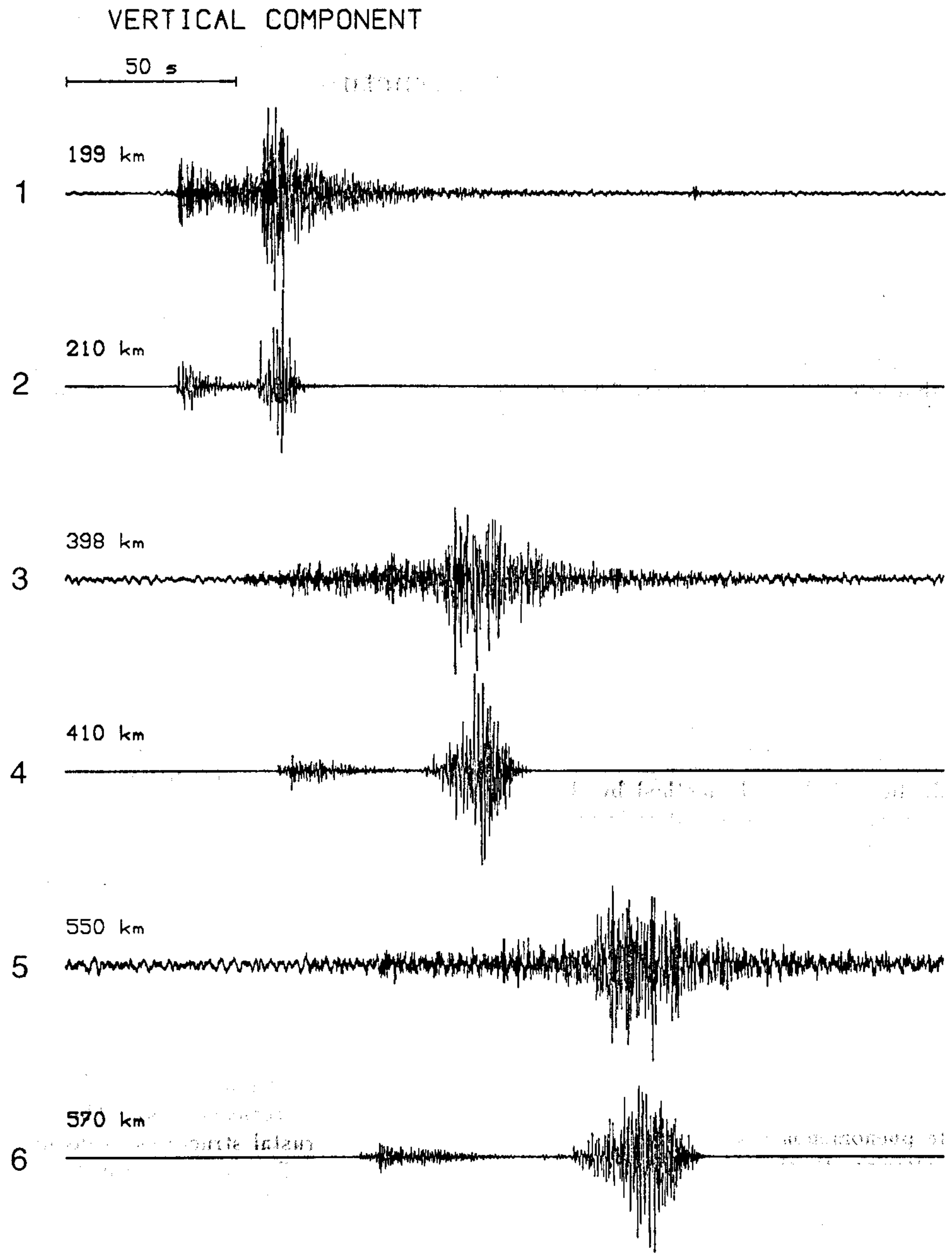


Fig. 1. Comparison between real records of the *Pg* and *Lg* phases (traces 1, 3, and 5) of an earthquake in southwestern France and synthetics (traces 2, 4, and 6) obtained at similar epicentral distances with a four-layer model.

amplitudes and shapes of the energetic parts of both *Pg* and *Lg*. It fails, however, to fit the long progressive temporal decay of the amplitude of these phases as observed on actual seismograms.

Another striking observation is the decay of spectral energy of the *Lg* phase with epicentral distance. *Campillo et al.* [1985] have shown that the introduction of a frequency dependent quality factor for the whole crust allows to simulate the decay of spectral energy with distance. In their study of the attenuation of *Lg* in different group velocity windows, they also found that the computed apparent quality factors of the crust were almost identical for the windows 3.6–3.1 km/s, 3.1–2.6 km/s, and 2.6–2.3 km/s.

However, the process of anelastic attenuation cannot account for the long duration of the coda. This suggests that scattering could be the cause of both frequency dependent attenuation and coda of the seismograms. This last interpretation is consistent with the frequency dependence of Qp and Qs and with the ratio Qs over Qp observed in France by *Campillo and Plantet* [1991]. An intriguing result obtained in this last study is the correlation of a zone of strong attenuation with a zone of strong heterogeneity of the lower crust as revealed by a wide-angle profile. This suggests that the intense scattering occurring in the heterogenous lower crust might be the cause of the attenuation anomaly.

HETEROGENEITY OF THE CRUST AS DISPLAYED BY DEEP REFLECTION PROFILES

The heterogeneous character of the crust is clearly shown by deep seismic reflection experiments. One of the major results of vertical seismic reflection profiling in western Europe [*Bois et al.*, 1988; *Matthews*, 1986; *Fuchs et al.*, 1987] is the recognition of the existence of a bright subhorizontal seismic layering in the lower part of the continental crust. On the other hand, the upper crust is nearly transparent except for the shallow sedimentary layers. The nature and the origin of these lower crustal reflectors remain controversial [*Warner*, 1990*a*]. However, the lateral coherency of the reflections suggests that the reflecting structures are mostly horizontal and very elongated in shape. Deep seismic reflection profiling thus provides the most reliable information about small-scale crustal heterogeneities since the reflectivity is a direct observation of the fluctuations of the properties of the Earth that induce the scattering of seismic waves. In this paper we study the implications of the layered lower crust for the characteristics of guided wave propagation. Because they consist of multiply reflected waves, *Pg* and *Lg* must be affected by heterogeneities in the same way as the waves used in reflection seismics, although the frequency ranges considered for these two domains are different. Moreover, the large amplitude of impedance contrasts that are probably associated with the layering, the large thickness of the heterogeneous zone, and the lateral extent of the region affected in western Europe support the idea that the influence of the layering on *Pg* and *Lg* phases has to be considered. Some characteristics of the *P* wave velocity structure of the lower crust are known as a result of the detailed analysis of reflection records. Particularly, one can evaluate the main features of acceptable distributions of velocity with depth such as the mean thickness of the layers or the mean velocity contrast between them. Such constraints have been obtained by modeling of wide-angle reflection data recorded in the Black Forest [*Sandmeier and Wenzel*, 1986] and in northern France [*Paul and Nicollin*, 1989], as well as from normal incidence data from the continental shelf off southwest Britain [*Warner*, 1990*b*; *Paul and Hobbs*, 1991]. These studies all lead to the same average thickness of 100 m for the reflective layering of the lower crust. They also show that the impedance contrasts required to explain the brightness of the reflections are as large as 20%, leading to velocity contrasts of at least 10% between high- and low-velocity layers. However, the frequency range considered for deep seismic reflection is 8–30 Hz, while for the study of short-period regional phases the domain of interest is 1–10 Hz. As a consequence, the waves can be sensitive to different scales of layer thicknesses in the two domains. Therefore the effect of the mean thickness must be investigated by a series of numerical experiments.

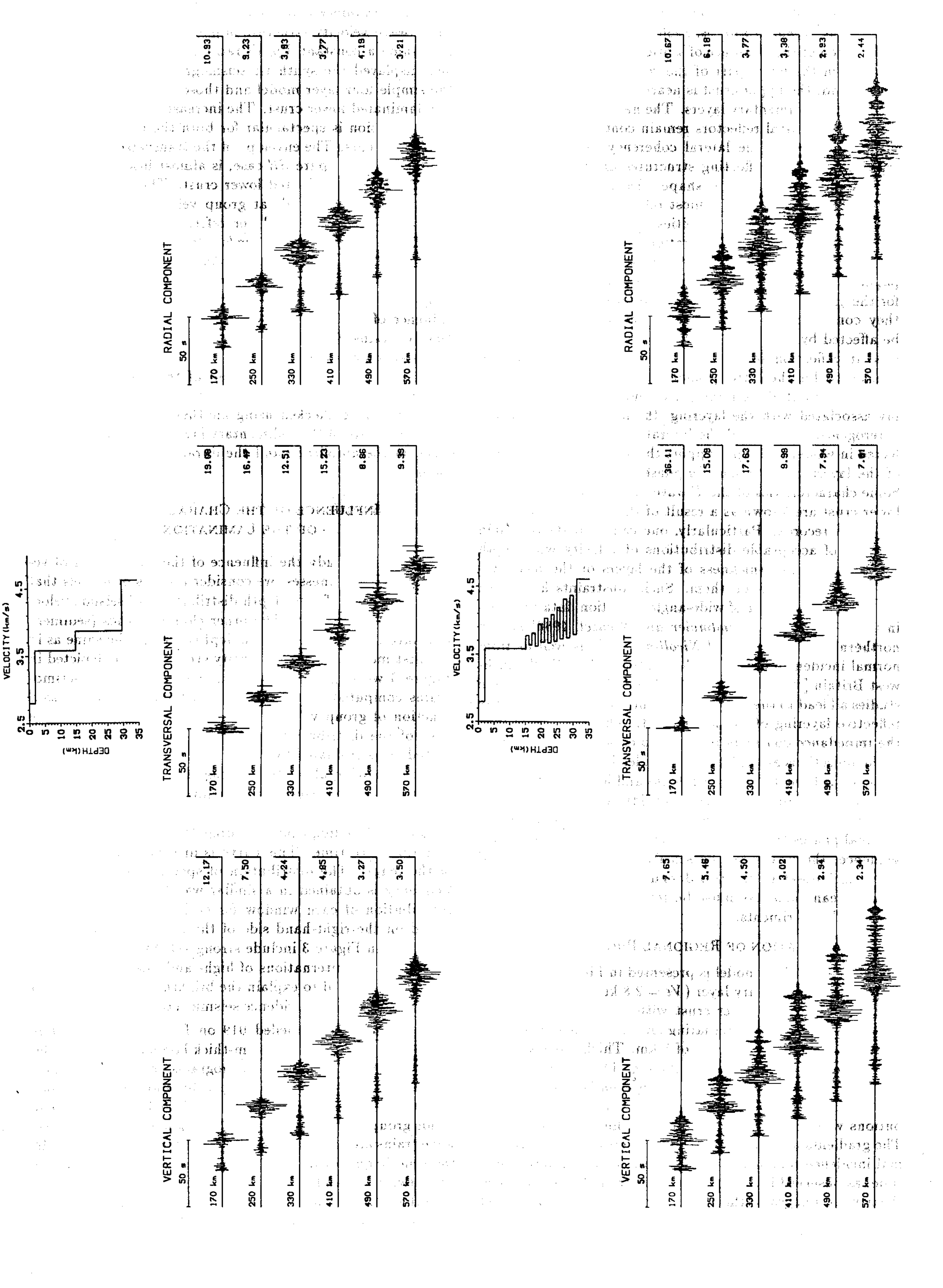
Simulation of Regional Propagation

The first velocity model is presented in Figure 2. Below a 2-km-thick sedimentary layer ($Vs = 2.8$ km/s) and a 13 km-thick homogeneous upper crust with $Vs = 3.58$ km/s, the lower crust is made of alternating low- and high-velocity layers with an average thickness of 1 km. Thicknesses are chosen randomly in the range 900–1000 m. The *P* wave velocity law is inspired from the model given by *Sandmeier and Wenzel* [1986], with alternations between two Gaussian distributions with average values increasing linearly with depth. The gradients are chosen so that the energy reflected at normal incidence remains nearly constant with two-way travel time, as observed in seismic data. At present, there are few shear wave images of the reflective lower crust [*Lüschen et al.*, 1990; *Sandmeier and Wenzel*, 1990]. As a consequence, the *S* wave velocity structure is poorly known, and we had to assume a constant Vp/Vs ratio of 1.73. In Figure 2 are also displayed the synthetic seismograms corresponding to the simple four-layer model and those obtained in presence of a laminated lower crust. The increase in duration due to the lamination is spectacular for both the radial and vertical components. The envelope of the transverse component, corresponding to pure *SH* case, is almost insensitive to the presence of the laminated lower crust. The coda of *Lg* exhibits significant arrivals at group velocities as low as 2.3 km/s, while for the four-layer reference model the lowest velocity observed was 2.9 km/s. Indeed, the envelope of the late part of the *Lg* wave train shows some unrealistic features very different from the regular decay of actual seismograms. Nevertheless, it is important to note the drastic influence of the lamination on the duration of both *Pg* and *Lg* wave trains. One may remark that this result contradicts the assertion of *Cormier et al.* [1991] that the *Lg* wave train is almost insensitive to the detail of the crustal structure and that its coda is controlled by the total thickness of the crust. We have checked using another model that the velocity structure of the sedimentary layer has a much smaller effect on the amplitude and the duration of the *Pg* and *Lg* codas.

Influence of the Characteristics of the Laminations

In order to study the influence of the distribution of velocities and thicknesses, we consider a series of models that correspond to different depth distributions of seismic velocity in the lower crust. The other characteristics (sedimentary layer, upper crust, Moho depth) remain the same as in the first model. Different velocity structures are depicted in Figure 3 with the corresponding synthetic vertical seismograms computed at a distance of 370 km and plotted as a function of group velocity. In each case, a time-frequency map of the distribution of spectral amplitude is also presented. It is evaluated by computing Fourier spectra in time windows moving along the seismograms. The windows are 8.125 s wide. In each time window, the total amount of spectral amplitude is evaluated by summing up the contribution of each frequency, yielding the curve of amplitude variation with time. This curve is displayed along the top of the maps. The distribution of spectral amplitude with frequency is obtained in a similar way by summing up the contribution of each window for each frequency. It is depicted on the right-hand side of the maps. All the models presented in Figure 3 include strong velocity contrasts with the systematic alternations of high- and low-velocity layers which are required to explain the brightness of lower crustal reflectors in vertical incidence seismic data [*Warner*, 1990*b*].

The first model (labelled 019 on Figure 3 because it is made up of 19 layers) has 1-km-thick laminations and is the one used for computing the seismograms presented in Figure 2. The time-frequency map (Figure 3*a*) allows one to visualize the specific distribution of energy of the late arrivals, that is, for group velocity smaller than 2.9 km/s. Well-organized wave trains are present in individualized frequency bands. In the time-frequency domain they have a remarkable continuity between 3.2 and 2.3 km/s. For model 019 the dominant arrival lies in the band 1.5–2 Hz and exhibits a slight in-


RADIAL COMPONENT
TRANSVERSAL COMPONENT
VERTICAL COMPONENT
50 s
170 km
250 km
330 km
410 km
490 km
570 km
VELOCITY (km/s)
DEPTH (km)

verse dispersion. Well-defined characteristics like these are much more difficult to see in the coda of *Pg*. Let us consider now models with a thinner layering in the lower crust. In model 033 (Figure 3*b*), the layers have a mean thickness of 500 m. Again, the increase in duration with respect to the reference model is very clear. The energy map shows that the main contribution is within the frequency range 2–4 Hz. Another arrival can be seen at a frequency of about 8 Hz. This indicates that the frequency content of the late arrivals is inversely proportional to the mean thickness of the layering. It is confirmed by the results obtained with model 121 (Figure 3*c*) where the average thickness of the layers is 100 m. In this case, the energy map shows that there is no generation of late arrivals for frequencies lower than 8 Hz. The three models that we have just discussed present unimodal distributions of layer thicknesses. The influence of more realistic distributions must also be tested.

In model 049 the lower crust is made up of layers with a bimodal distribution of thicknesses in the ranges 900–1100 m and 90–120 m. The velocity structure and the results of the computations are shown in Figure 3*d*. The patterns seen on the time-frequency plot are more complex and diffuse than in the previous examples. The arrival at about 2 Hz which was noticed for model 019 is present. The effect of mixing two scales of thicknesses is to spread the energy in the time-frequency domain. With model 031, the layer thicknesses are randomly distributed between 100 m and 1000 m. We present the results obtained in Figure 3*e*. The coda is well developed and reaches group velocities as low as 2.2 km/s, as in the case of the models with narrower distributions of thicknesses. Indeed, in this case the energy is widespread in the time-frequency domain. On the seismograms, the decay of amplitude with time after the onset of *Lg* is more regular than in the previous examples, resulting in an envelope that seems more realistic. Model 031 corresponds to a particular draw of random thicknesses between 100 and 1000 m. In order to check the importance of the details of the distribution, we set up a new model (model 035) with the same characteristics and we recalculated the synthetic seismograms. The results are presented in Figure 3*f*. Indeed, the wave fields obtained for models 031 and 035 present some differences, but the overall characteristics are the same, and in particular the envelopes of the coda are similar. On the other hand, the details of the energy maps are different, showing that the spectral amplitude of the late arrivals changes rapidly from one draw to another.

We have performed these different simulations in models with roughly the same statistical characteristics of layering in order to check what differences could be observed between records obtained for different paths in the same region. This indicates that the mean results obtained for a large number of paths will show strong maxima of excitation of the early coda only for frequencies corresponding to a range of thicknesses predominantly represented in the crust beneath most of the path.

We present in Figure 4 an example of energy map for a seismogram recorded in central France at approximately the same epicentral distance as the synthetics of Figure 3. It illustrates the predominance of the low-frequency component in the later part of the *Lg* wave train for group velocities smaller than 3.1 km/s. We computed the mean ratio of spectral amplitudes in the group velocity windows 3.6–3.1 and 2.6–2.3 km/s for a series of paths beneath central France and showed that the mean ratio decreases for frequencies smaller than 2 Hz [*Campillo*, 1990]. This spectral behavior is independent of the shallow geologic conditions beneath the stations. It indicates that, on the average, the early coda is mainly excited in a frequency range corresponding to layer thicknesses around 1 km. One must remember that we have no information for frequencies lower than 1 Hz.

Nature of the Late Arrivals

We have shown that the presence of a laminated lower crust implies a duration of the regional seismograms much longer than the one predicted by a simple model having a homogeneous lower crust. It appears that the spectral characteristics of this early coda are closely related to the statistical properties of the stratification. We have not addressed yet the problem of the identification of the waves that form this slow wave train. To this aim we will consider the simplest case among those presented in the previous sections: model 019. We have seen that a specific mode of propagation exists for frequencies around 2 Hz with a group velocity that reaches 2.3 km/s. We have computed the particle motion on the surface and at depth for different group velocity windows at 530 km of epicentral distance. The amplitude spectra of the seismograms are strongly dominated by the low-frequency component around 1.5 Hz, as shown in Figure 3*a*. In Figure 5 (left) we present the particle motion computed in the group velocity window 3.45–3.40 km/s, which corresponds to the maximum amplitude of *Lg*. At the surface, the particle motions are complex and show both radial and transverse components, indicating that *Lg* is made up of *SV* and *SH* waves propagating over a wide range of incidence angles. The same observation can be made at depths of 14 and 28 km. Beneath the Moho, at 34 km depth, the radial component of the motion disappears as expected for vanishing *SV* and *SH* waves. If we consider now a time window in the wave train that appears only in presence of the laminated lower crust, the polarization is completely different. The particle motions computed in a window corresponding to group velocities between 2.70 and 2.65 km/s are displayed on Figure 5 (right). At the surface and in the upper crust the motion in the incidence plane is complex and does not indicate any type of linear polarization. As we observed previously, the transverse (*SH*) component is equal to zero. In the lower crust (28 km depth) the polarization becomes almost linear and vertical. When the Moho is reached, the direction of polarization changes suddenly and corresponds to a *SV* wave propagating downward in the lower half space with an angle of about 66°. These polarization patterns indicate that the later arrivals that occur in presence of thin layering in the lower crust have a very distinct nature from the primary *Lg* (defined between 3.5 and 3.0 km/s). Be-

Fig. 2. Synthetic seismograms obtained between 170 and 570 km of epicentral distance for the four-layer model with a homogeneous lower crust (top) and for a 19-layer model with layering at kilometric scale in the lower crust (bottom).

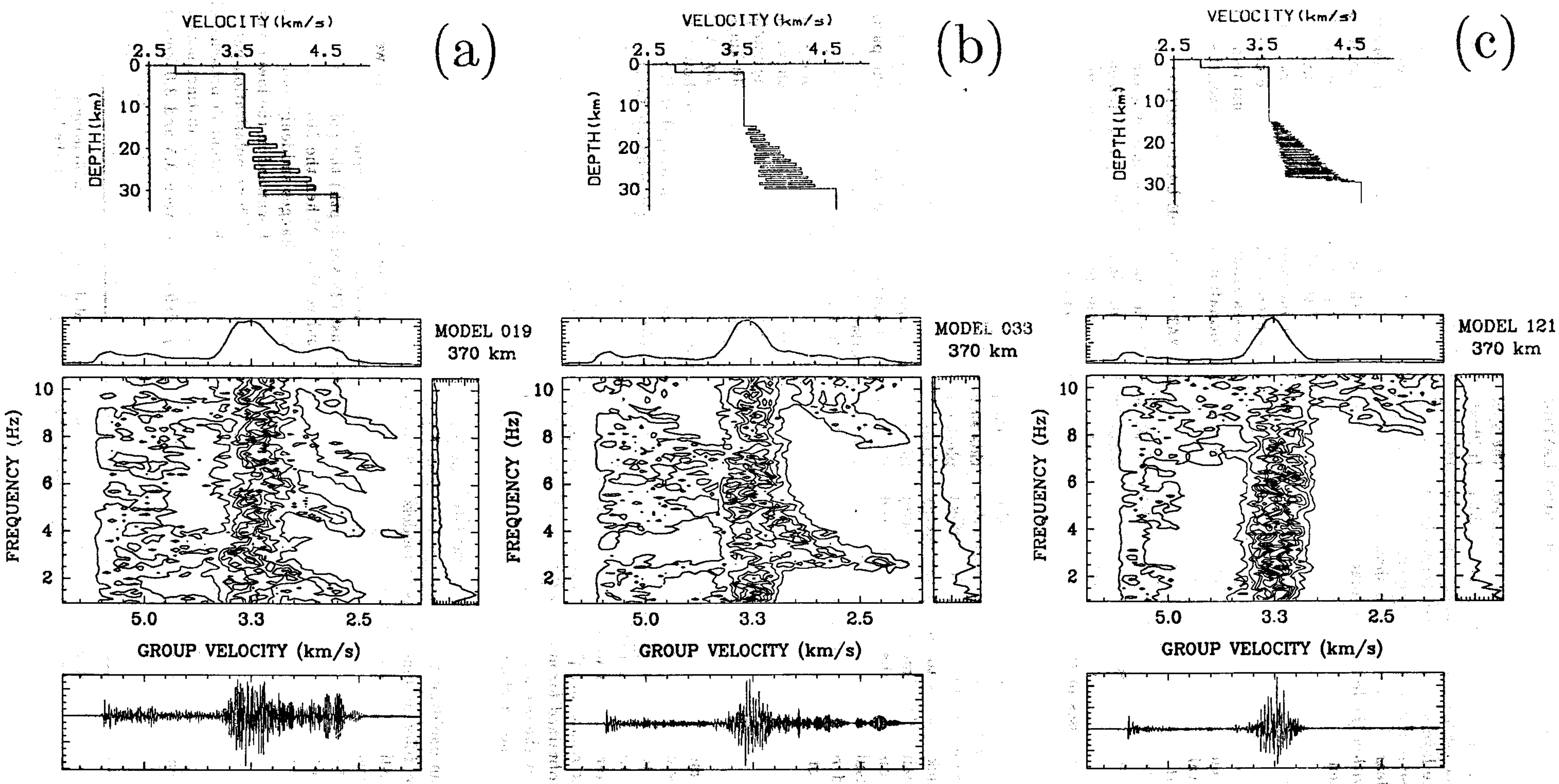


Fig. 3. Time-frequency maps of spectral amplitude for synthetic seismograms obtained at 370-km epicentral distance with different lower crustal velocity structures. The velocity models are presented above the corresponding maps. The horizontal axis is scaled linearly with respect to the inverse of the group velocity. The seismograms are plotted at the same scale. The curve plotted along the top axis of each map is obtained by summing up the contributions of all the frequencies. The curve presented to the right-hand side of each map is obtained by adding the contributions of all the time windows. It is used as a normalization curve to compensate for the attenuation of high frequencies before plotting the contour levels on the maps.

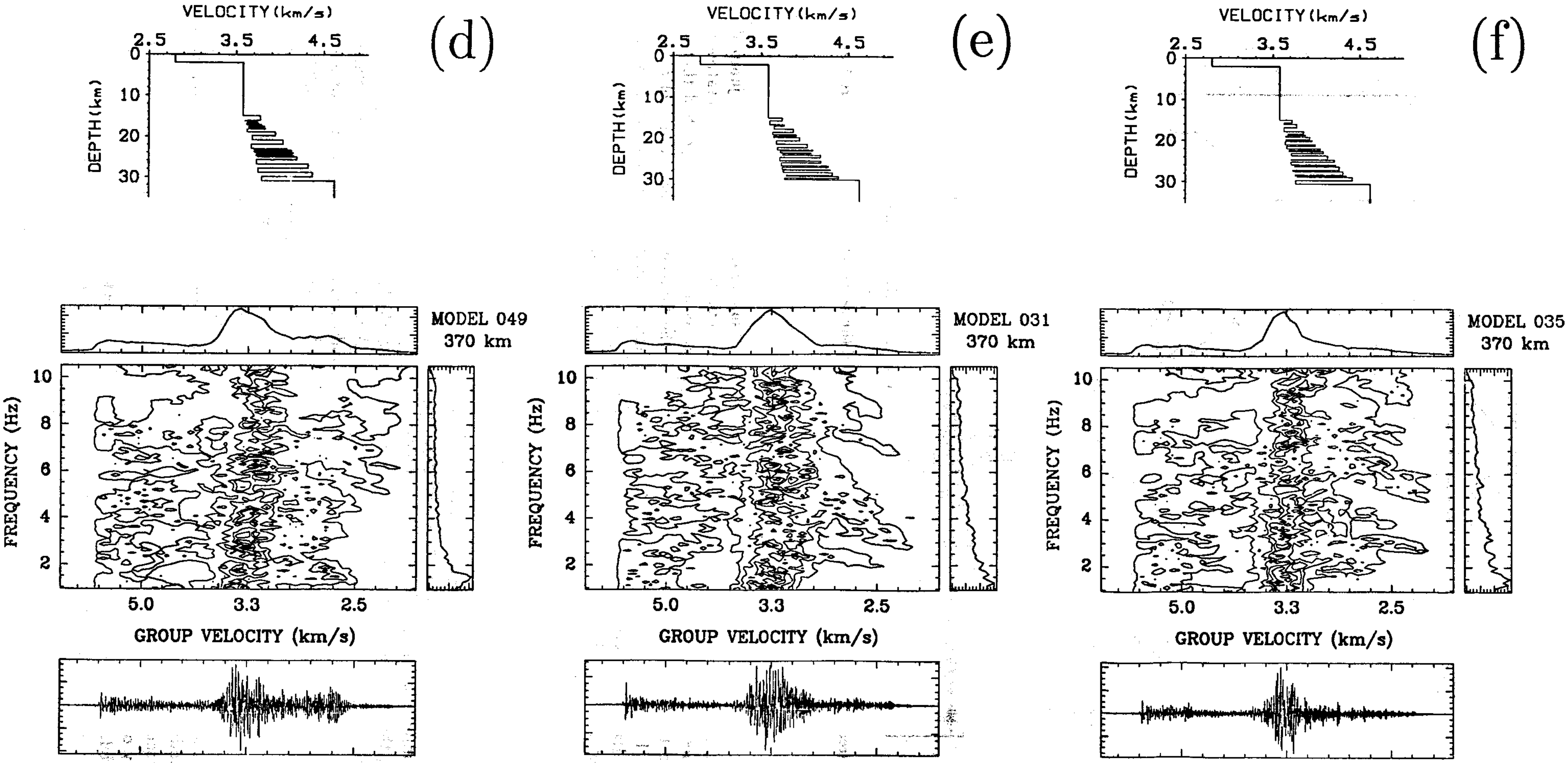


Fig. 3. (continued)

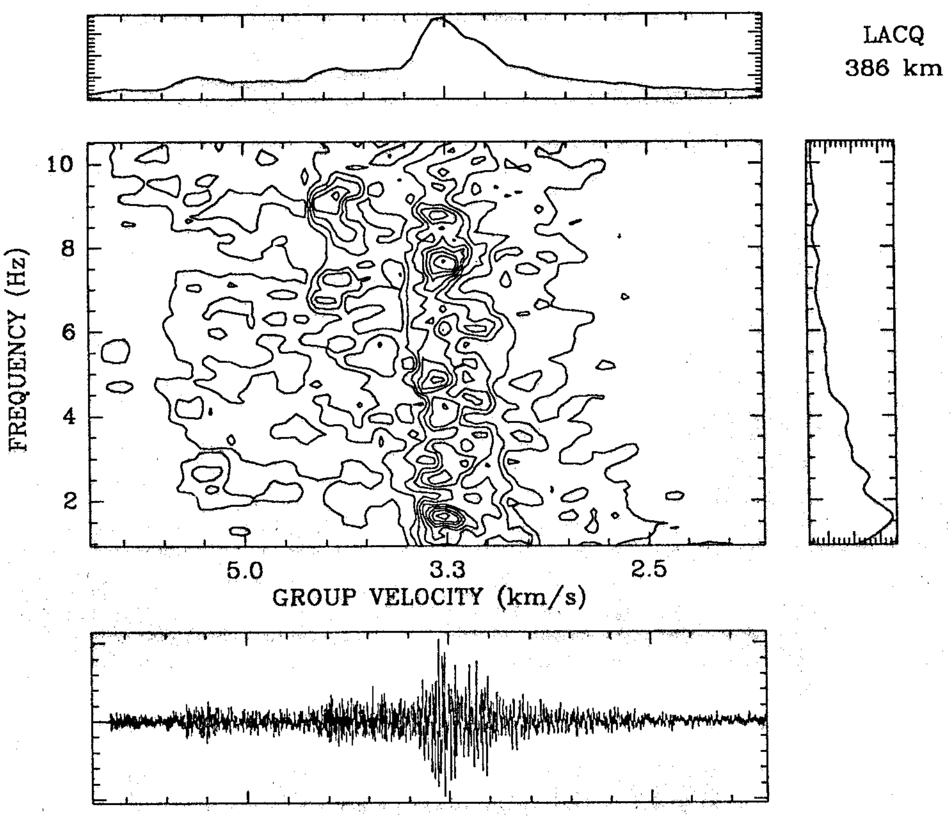


Fig. 4. Time-frequency map of spectral amplitude for a record of the same earthquake as in Figure 1 at 386-km epicentral distance.

fore discussing further the nature of these waves, one must notice that the vertical polarization observed in the lower crust for group velocities between 2.70 and 2.65 km/s is a phenomenon limited to the mode of propagation observed at frequencies between 1 and 2 Hz. When considering frequencies higher than 4 Hz and the same group velocity, the polarization of the motion in the lower crust is observed to be no longer vertical but almost circular. On the other hand, the linear polarization of the waves propagating in the mantle is observed to have the same angle regardless of the frequency considered. This last point suggests a leakage in the mantle of part of the energy of these "modes". To check this conclusion, we use the framework of normal mode theory. The dispersion curves of normal modes are computed using the computer programs developed by *Herrmann* [1985] following the approach of *Haskell* [1953]. We consider first the four-layer reference model with a homogeneous half-space mantle. The dispersion curves of all the higher modes of rank greater than 6 are plotted in Figure 6*a*. The lower group velocity of the crustal modes is about 3.1 km/s. Considering now model 019 that includes a laminated lower crust, the dispersion curves presented in Figure 6*b* indicate a lower limit of group velocity at about 2.9 km/s. Nevertheless, group velocities lower than 3 km/s are reached by a small number of modes. It may be noticed that the contribution at 2.9 km/s occurs at a frequency of 1.8 Hz that is the dominant frequency of the late arrivals observed in our synthetics for this model. However, our numerical simulations have shown that energy can travel at velocities much lower than those computed for normal modes.

It is a usual practice to introduce a jump of wave velocities in the mantle to get modes representative of upper mantle propagation as *Sn*. We have verified that such a modification of the four-layer reference model does not change the lower limit of the group velocity as shown in Figure 6*c*. On the contrary, when model 019 with a jump of velocity in the mantle is considered, the dispersion curves reach velocities smaller than 2.7 km/s (Figure 6*d*). The locations of the minima of the group velocity that correspond to Airy phases give an image very close to the energy map obtained for the same model as presented in Figure 3*a*. In particular, the distribution of the minima indicates a slight inverse dispersion of the resulting energy wave train. These modes appear only in presence of the laminated lower crust together with the jump of velocity. This indicates that they are associated with the lamellae and that they are not normal crustal modes. After the observation of the polarization, this is another indication of the importance of leaky modes in the early coda. Their characteristics suggest that these modes result from the interaction of the plate type vibrations of the high-impedance layers. We have checked that this effect does not exist for Love wave higher modes. This is in agreement with our previous observation that there is no coda produced by the lamination for the transverse component.

Apparent Attenuation

With respect to the evaluation of *Qs* in the crust from actual seismograms, the important point is to evaluate the rate of amplitude decay of these modes. A perfectly elas-

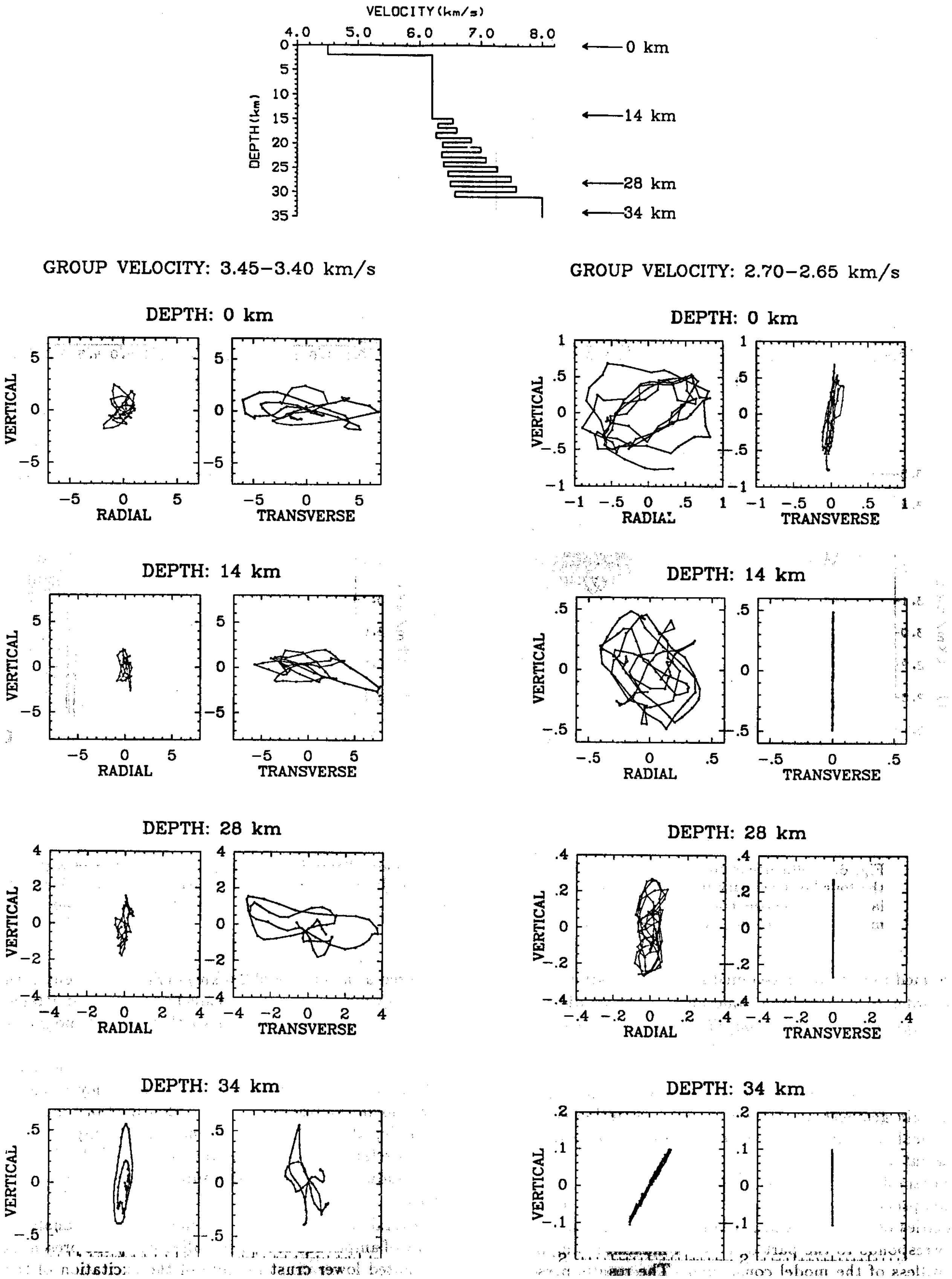


Fig. 5. Particle motion obtained from synthetic seismograms computed at 530 km distance and at four different depths: 0, 14, 28, and 34 km. The velocity model includes 1-km-thick lower crustal layers. On the left-hand side, particle motions correspond to the primary *Lg* (group velocity between 3.45 and 3.40 km/s). On the right-hand side, lower group velocities (2.70 to 2.65 km/s) are considered to study the characteristics of the late arrivals which appear when the lower cust is layered.

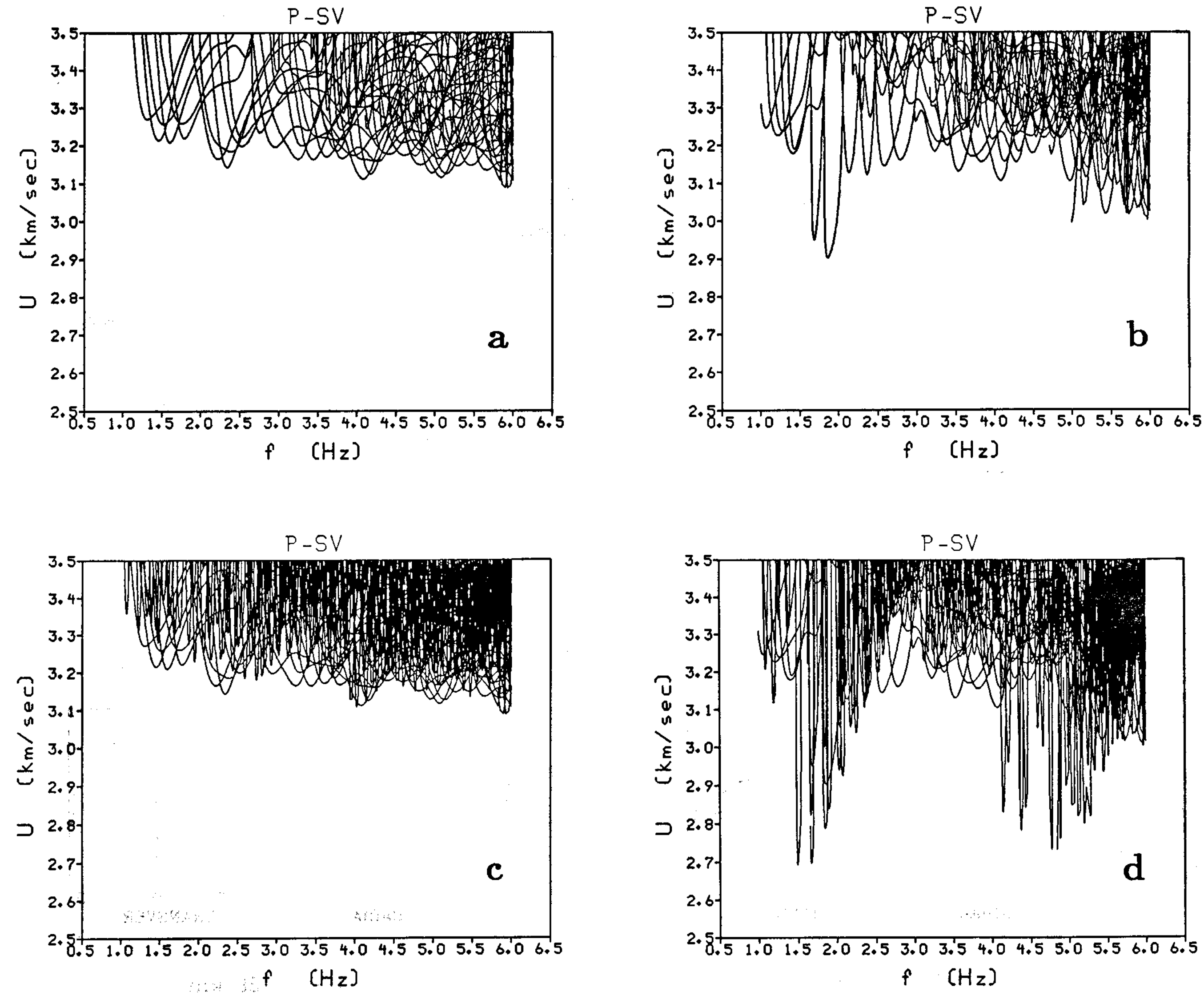


Fig. 6. Influence of the lower crust laminations on the normal modes dispersion curves. (*a*) Result obtained with the four-layer velocity model and a homogeneous half space mantle. (*b*) Laminated lower crust with 1 km-thick layers and homogeneous half space mantle. (*c*) Four-layer model with a velocity jump in the mantle. (*d*) 19-layer model with a velocity jump in the mantle.

tic medium has been assumed in our calculations. However, in order to perform the same analysis as we did with data [*Campillo et al.*, 1985; *Campillo and Plantet*, 1991], we studied the decay of spectral amplitude with epicentral distance for different group velocity with the synthetics computed for model 019. The seismograms were corrected for time domain geometrical spreading of *Lg*, as evaluated from numerical simulation in a reference model. We evaluated apparent Q values from linear regression of the logarithm of the amplitude with distance. Figure 7 shows the results of this processing for two time windows centered at group velocities of 3.3 km/s and 2.6 km/s. The first case (3.3 km/s) corresponds to the part of *Lg* with maximum amplitude regardless of the model considered. The results presented in Figure 7 (left) indicate that there is no measurable apparent attenuation of *Lg* due to the excitation of low-velocity modes in presence of lamination. The poor correlation is explained by the rapid variations of *Lg* amplitude due to interferences between multiply reflected S waves. Considering a window centered at a velocity of 2.6 km/s (Figure 7 (right)), the loss of energy of the late arrival leads to an apparent attenuation. The reader must keep in mind that the seismograms were corrected from the spreading of *Lg*. The values of Q that were computed are never smaller than 1000. The minimum values are not associated with the frequency ranges of the low-velocity arrivals which were identified in Figure 3*a*. For example, at 2 Hz, where the maximum amplitude occurs, the correlation coefficient is very small and the Q value very high, indicating that the spreading is very similar to the one of *Lg*.

Several conclusions can be drawn from this analysis. The decay of amplitude of *Lg* is not affected by the presence of the laminated lower crust in spite of the excitation of the slow waves. Surprisingly, the decay of the slow modes is close to the decay of *Lg*. This means that the leakage of energy into the mantle is weak with respect to the geometrical spreading of a guided wave. In the frequency bands 1.5–2 Hz or 4–5 Hz, where significant energy travels at small group velocities

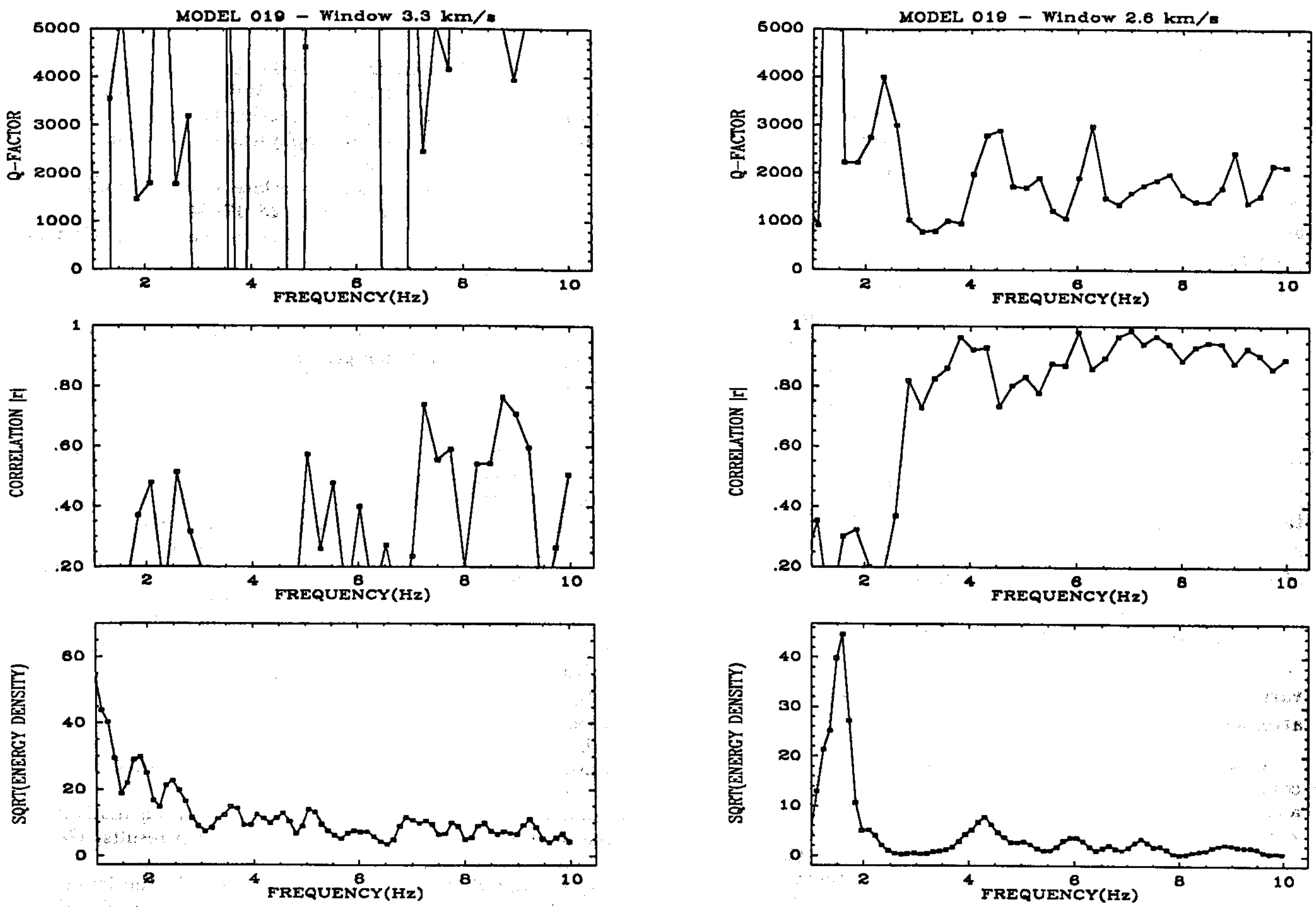


Fig. 7. Results of the measurement of the apparent Q factor for model 019 from the decay of spectral amplitude with distance. The density of spectral energy is computed in a 8.125-s-wide window centered on a group velocity of 3.3 km/s on the left-hand side and 2.6 km/s on the right-hand side. The variations of the Q factor with frequency are presented on top. The curves in the middle show the linear correlation coefficient for the linear regression used to measure Q. The bottom curves show the square root of the energy density for the seismogram at 170-km epicentral distance in the corresponding time window.

(2.5 km/s), the values of the equivalent Q factor representing the leakage of energy are much larger than the average values measured for the crust in France (about 400 at 1.5 Hz and 640 at 4 Hz, respectively). This mode of generation of the early coda is therefore consistent with the observation that the Q measurements made in different group velocity windows between 3.5 and 2.3 km/s lead to almost the same value even when the analysis assumes the same geometrical spreading regardless of the window considered [*Campillo et al.*, 1985].

However, this observation is no more universal than is the existence of a laminated lower crust. For example, the lower crust in shield areas often appears to be transparent to the high-frequency waves used in deep seismic reflection experiments [*Nelson*, 1991]. When reflections are recorded, like in the Baltic and Bothnian Echoes from the Lithosphere (BABEL) profile of the Gulf of Bothnia [*BABEL Working Group*, 1990], they still are less bright and laterally continuous than in Paleozoic and Mesozoic areas. On the other hand, the coda of *Lg* in shield areas is much less developed than for the seismograms recorded in France. This can be seen by comparing the data observed in Scandinavia and studied by *Kim* [1987] with those recorded in France and presented here. Our analysis is relevant for France and may be relevant for other regions where the reflective lower crust has been observed.

Conclusion

We have included present-day knowledge of the structure of the lower crust in the models used to simulate the propagation of regional phases. A striking implication is the fact that taking into account the lamination of the lower crust increases considerably the duration of the wave train which begins with the onset of *Lg*. Numerical simulations predict arrivals of energy with group velocity as low as 2.3 km/s. This is in good agreement with the envelopes of seismograms recorded in France.

The study of the polarization of these waves and the modal representation of the wave field suggest that the late arrivals consist of plate vibration-like disturbances which leak energy in the mantle at a very weak rate. On the other hand, these modes show a decay of amplitude with distance which is close enough to the geometrical spreading of *Lg* to allow that the apparent quality factor deduced from these waves is close to the one obtained from *Lg*. This model is therefore in agreement with the observation that Qs for the crust is almost constant when measured in different group velocity windows assuming the spreading of *Lg*. Another result of this study is that the spreading of *Lg* is not affected by the presence of the thin layering of the lower crust, in spite of the generation of the coda.

Indeed, the existence of a layered lower crust does not account for the observed values of the apparent quality factor, and we have to assume that the loss of energy of the wave train observed in real data is due to anelastic attenuation or to the dispersive effect of scattering over randomly distributed inhomogeneities. After taking into account the findings of the present study, we maintain our previous conclusion [*Campillo and Plantet*, 1991] that random scattering very likely prevails for the attenuation of short-period waves.

Our results illustrate the fact that the early coda of regional seismograms may be strongly affected by waves relevant from the multiple scattering theory like the slow arrivals discussed in this paper. We have verified that these waves exist also at distances smaller than the critical distance for the reflection on the Moho. Indeed, we can emphasize the recommendation of *Aki* [1982] that the analysis of coda using backscattering theory is valid only for very large lapse time (larger than twice the *S* wave travel time). According to our results, the early coda can contain wave trains which exhibit an inverse dispersion that mimics the variation of prominant frequency expected from the effect of attenuation.

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