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VII.3 Variability of explosion Lg spectra with near-source structure and focal depth

Introduction

Lg amplitudes are used to estimate magnitudes of seismic events, and in particular of nuclear explosions (see Ringdal and Hokland, 1987). Modelling of Lg waves in the context of seismological verification research has mainly addressed the study of the potential of Lg waves for discrimination between explosions and earthquakes, and has assumed laterally homogeneous structures between the source and recording station (e.g., Nakanishi, 1981; Lilwall, 1988). We study here explosive sources and attempt to assess how small variations in focal depth or in structure around the source region affect recorded Lg-spectra. We use the Eastern Kazakh area as source region and NORSAR as the recording site. Following Levshin (1985), who pointed out the importance in surface wave amplitude modelling of using adequate structures at the source and receiver sites, we allow for different crustal structures in the Eastern Kazakh and the NORSAR region, and assume a smooth, lateral variation in between.

Crustal models

Four different crustal structures of the Eastern Kazakh area are proposed by Priestley et al (1988), one of them originating from the compilation of Russian literature by Leith (1987; written communication to Priestley et al), and the three others by inversion of broad-band teleseismic P-waveforms recorded in the area. We select two of these models: the model compiled by Leith, which presents a regular increase of velocity with depth (hereafter called model 1), and one of the three inverted models (the BAY-Japan model, called model 2) which all differ from model 1 mainly by a low-velocity zone at depths around 5 to 10 km, and higher velocities in the lower crust. In addition to the effect of an overall crustal modification, we study the effect of a local change in the model properties around the focal depth of the explosion. For that purpose, we use models where the interfaces at 1 km depth in models 1 and 2 are transformed into velocity gradients over depth intervals of 1 km (models 1' and 2'). These gradients may model either a real gradient in velocities, or a splitting of the single interface into a few interfaces with smaller velocity contrasts. At the receiver end, we use a model of the crust under the NORSAR array derived by Gundem (1984). Models 1 and 2 and the crustal model under NORSAR are presented in Fig. VII.3.1.

We assume that the structures vary smoothly around the source and receiver sites. More quantitatively speaking, the models described above should <u>not</u> exhibit variations that are large enough to affect significantly the modal eigenfunctions of the Lg waves in at least 10 wavelengths (40 km here) around the source and station sites. Our own experience is that a 1 km change in Moho depth, and even smaller changes in sediment thicknesses, are large variations in this sense. The lateral variation of the crust is well known around NORSAR, and variations of the Moho depth of a few kilometers have been found in the area (Berteussen, 1977), showing that we are certainly at the limit of our smoothness assumptions at the receiver site. In Eastern Kazakh, the lateral crustal variations are not expected to be strong (Priestley et al, 1988).

The propagation path between Eastern Kazakh and NORSAR lies in an old continental shield and does not cross any significant tectonic feature. In the absence of more detailed information, the more realistic assumption which can be made on the model along the propagation path is that the structure varies smoothly from the source to the receiver region.

Attentuation is introduced in the models with quality factors at 1 Hz increasing from 80 at the top of the crust to 300 at 1 km depth and to 1000 at the Moho depth. The frequency dependence of the quality factors is taken as $\sqrt{f}$, following the observations of Campillo et al (1985) in Central France.

Modelling procedure

Since we have assumed that the model is smooth at the Eastern Kazakh source site, the excited Lg wavefield can be decomposed there on the local Lg modal eigenfunctions, and the modal excitation $e_j(h,\omega)$ at the receiver site. The wavefield produced in Eastern Kazakh and recorded at NORSAR can thus be written, separating excitation, propagation and reception terms (see Levshin (1985) for the details):

$$\underline{u} = \Sigma_{i,j}\ \underline{u}_i(0,\omega)\ a_{ij}\ e_j(h,\omega) \qquad (1)$$

where $\underline{u}_i(0,\omega)$ is the surface site response of mode i, $e_j(h,\omega)$ is the excitation of mode j for a source at depth h, and a_{ij} is the propagation matrix involving velocity, attenuation, geometrical spreading, and possibly mode conversions along the propagation path.

If the structure is smoothly varying along the propagation path, it has been shown that there is no mode conversion (Woodhouse, 1974), and expression (1) reduces to:

$$\underline{u} = \Sigma_i\ \underline{u}_i(0,\omega)\ a_i\ e_i(h,\omega) \qquad (2)$$

Each mode propagates along a ray, and the propagation term a_i depends on the ray pattern for mode i. Since we do not have enough detailed information on the crust between Eastern Kazakh and NORSAR to make ray-tracing worthwhile, we calculate the phase and attenuation of each mode with the assumption that it has followed the great circle between source and receiver, and that its slowness on that path is a symmetric function of distance (all smooth symmetric functions lead to the same result). We use the geometrical spreading factor $1/\sqrt{r}$, where r is the source-station distance.

In order to avoid differences in calculated spectra which would originate from slightly different propagation path characteristics and not from differences in the source area itself, the model along the propagation path is fixed, whichever model is used for the source area, and is taken as varying from model 1 to the NORSAR model.

Synthetic Lg spectra are calculated for explosions in the four models described previously in this report. We focus our attention on focal depths around 1 km, and we take unit displacement steps as source functions. The spectra are calculated in a period interval between 0.8 and 1.25 s, corresponding to Lg wave periods observed at NORSAR.

The phase and group velocity dispersion curves of the Lg modes are displayed in Figs. VII.3.2 and VII.3.3 for models 1 and 2, and in Fig. VII.3.4 for the NORSAR model. It can be seen that due to the Moho being shallower at NORSAR than in the Eastern Kazakh area, fewer modes are of Lg type at NORSAR than in models 1 and 2. We thus have to define which modes at which frequencies will be considered as part of the Lg wavetrain at NORSAR. We adopt very simple criteria which are similar to the criteria used in data analysis: all arrivals arriving at NORSAR in a time window corresponding to group velocities (which are functions of the local group velocities along the whole path) between 3.1 and 3.62 km/s are considered as Lg waves. With that definition, some of the higher modes which are excited as Lg modes in Eastern Kazkh are excluded from contributing to the seismogram at NORSAR because they convert to Sn modes early during the propagation.

An example of a synthetic displacement spectrum is shown in Fig. VII.3.5. We notice the good agreement of its general characteristics with those of a data spectrum (Fig. VII.3.6). The slopes, which are strongly dependent on the attenuation in the structure, are similar; the peak-and-trough patterns, which originate from the multimodal character of the Lg waves, have similar amplitude and periodicity in both figures.

In order to facilitate the comparison between different spectra, we smooth them by a procedure equivalent to adding spectra from many recording stations distributed in a 60 km distance window around a main station site. The spectrum of Fig. VII.3.5 is shown after smoothing in Fig. VII.3.7.

Synthetic Lg spectra at NORSAR from Eastern Kazakh explosion sources

The synthetic displacement spectra of Lg waves recorded at NORSAR and originating from explosions at 3 different focal depths in models 1, 1', 2 and 2' are displayed in Figs. VII.3.8 and VII.3.9. We notice in Figs. VII.3.8a and VII.3.9a that the spectra are totally insensitive to a source depth variation from 0.7 to 0.9 km in models with a sharp interface at 1 km depth. Conversely, moving the source from the upper layer to the layer underneath the interface (source at 1.1 km depth) reduces the spectral amplitude by a factor corresponding to about 0.6 magnitude units. When the sources are located in velocity gradient zones (Fig. VII.3.8b and VII.3.9b), the decrease of spectral amplitude with focal depth is more regular but still significant.

Now comparing the spectra for identical source depths but different crustal structures, we notice a general increase in amplitude by a factor of 1.5 to 2, or 0.2 magnitude unit difference, between sources located in models 1 or 1' (Fig. VII.3.8) and sources located in models 2 or 2' (Fig. VII.3.9). There is an overall factor of 10 in the middle frequency range between amplitudes of the lowest and highest spectra of Figs. VII.3.8 and VII.3.9. The detailed crustal model at the source and the source depth appear to be important factors in the Lg spectra amplitude. On the other hand, we do not observe clear and significant differences in the slopes of the different spectra. This modelling does not predict any change in the dominant frequency of the data (the presented spectra multiplied by the instrumental response) for explosions at different depths or in slightly different crustal environments.

Spectra in a laterally homogeneous structure

To appreciate the influence of the NORSAR crustal structure on the spectral shapes, we show in Fig. VII.3.10 spectra calculated with another structure than NORSAR's at the recording site. We assume lateral homogeneity of the crustal structure along the whole path, which means that we calculate spectra for a station which would be situated at 4200 km distance from the Eastern Kazakh site, but in a crustal structure identical to model 1. The discrepancy between the spectral amplitudes for different source depths and source area models is reduced to a factor of 3 between the extreme cases, in comparison with the factor of 10 computed with the NORSAR model at the receiver end. The difference in structures between the Eastern Kazakh and NORSAR crust enhances the amplitude differences in Lg spectra.

Conclusion

We have shown that Lg spectra modelled in smoothly varying structures from Eastern Kazakh to NORSAR vary in amplitude when small realistic variations are introduced in the focal depth or in the crustal structure of the source site. Large amplitude variations, equivalent to up 0.6 magnitude unit difference, can be expected when the source focal depth crosses a layer interface with strong velocity contrasts. The equivalent of 0.2 magnitude unit variations may occur when the crustal structure is modified. On the other hand, no significant variation of spectrum slope or spectral content is observed with such source environment modifications.

These results can be compared with the characteristics of NORSAR recordings of Lg-waves from Eastern Kazakh explosions. The main difference between spectra from different explosions is an amplitude shift, but no large variation in the dominant frequency is observed (Kværna & Ringdal, 1988). The magnitude histogram presents distinct maxima separated by less than 0.1 magnitude units, and which do not correlate with explosion locations within the test site area (Ringdal and Hokland, 1987).

The stability of the spectral content of the Lg wavetrain is a feature common to our modelling exercise and the observed data. On the other hand, the observed spectral amplitudes are more stable than one would expect from the modelling results. The distinct character of the observed magnitude peaks, which are uncorrelated with explosion epicenter location, seems to indicate that the 20 x 20 km site area is small and homogenous enough for the epicentral location not to influence the Lg magnitude of the explosions. The different peaks can be interpreted as corresponding to sources of difference sizes fired in similar conditions, but our modelling results show that they could also originate from identical explosions fired in different geological layers or at different depths within a velocity gradient area.

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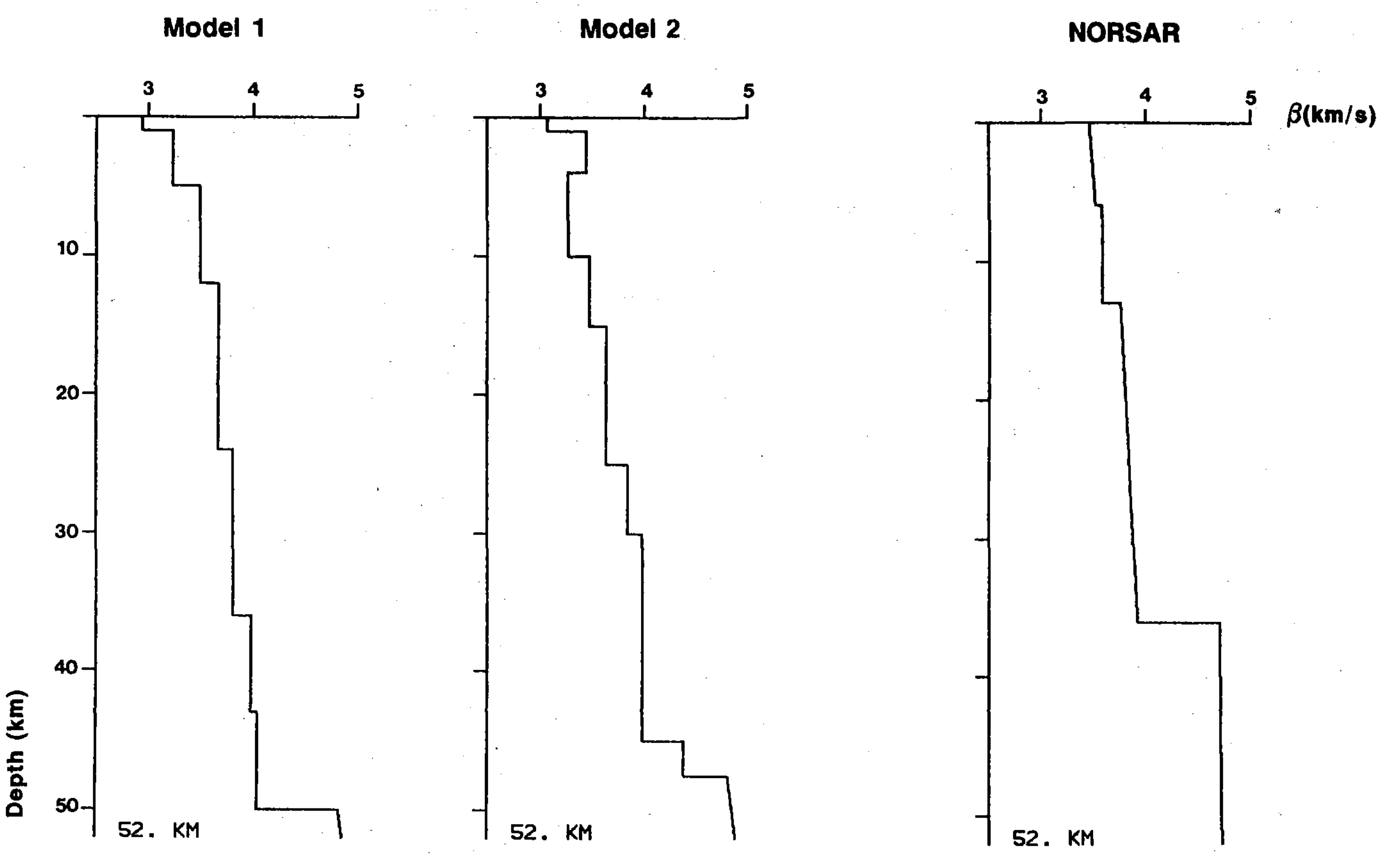
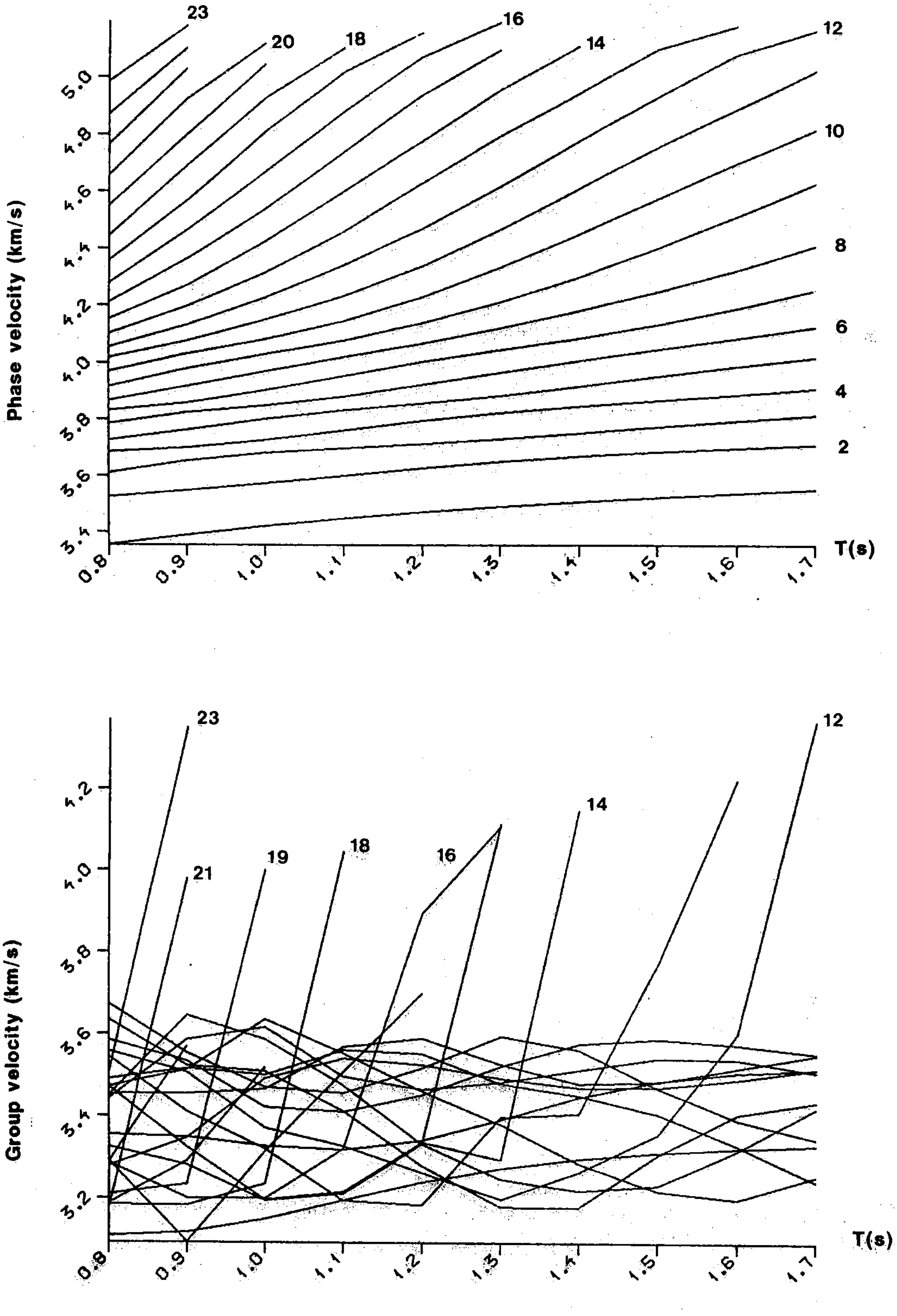


Fig. VII.3.1. S-wave velocity as a function of depth in the Eastern Kazakh models 1 and 2 and in the NORSAR model.



<u>Fig. VII.3.2.</u> Phase and group velocity dispersion curves for the Lg modes in model 1. Some of the curves are labelled with mode number.

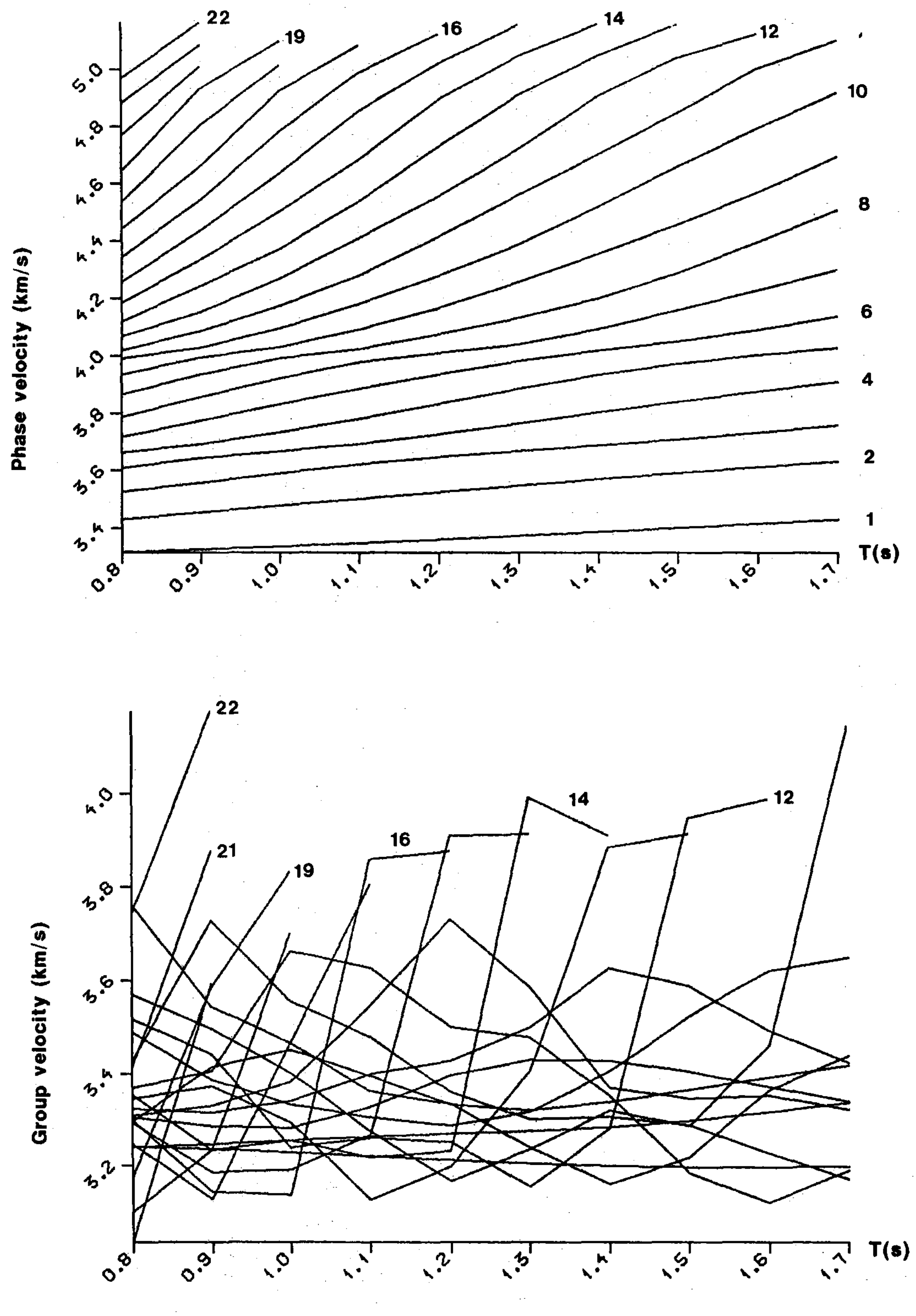


Fig. VII.3.3. The same as Fig. VII.3.2 for model 2.

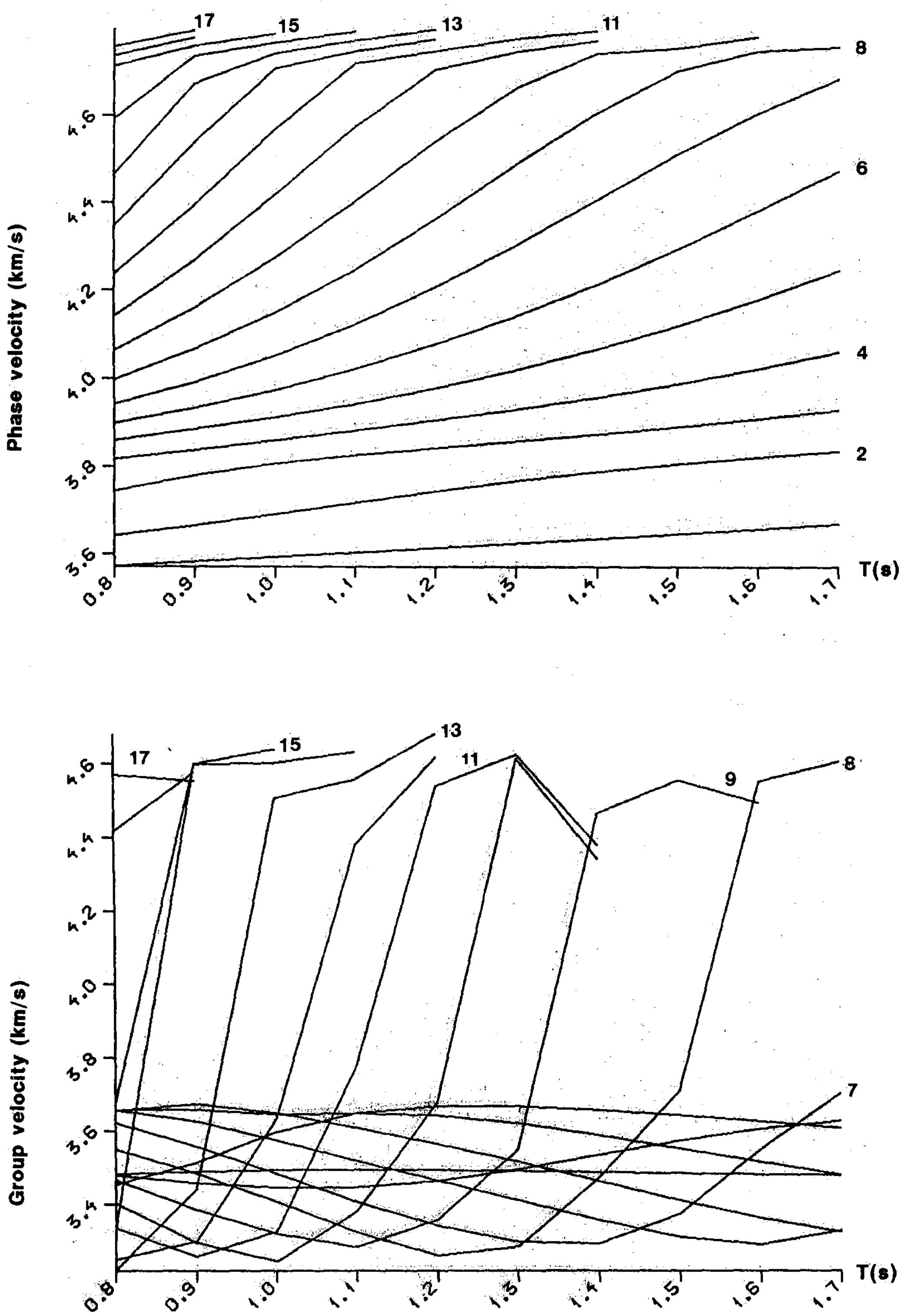


Fig. VII.3.4. The same as Fig. VII.3.2 for the NORSAR model.

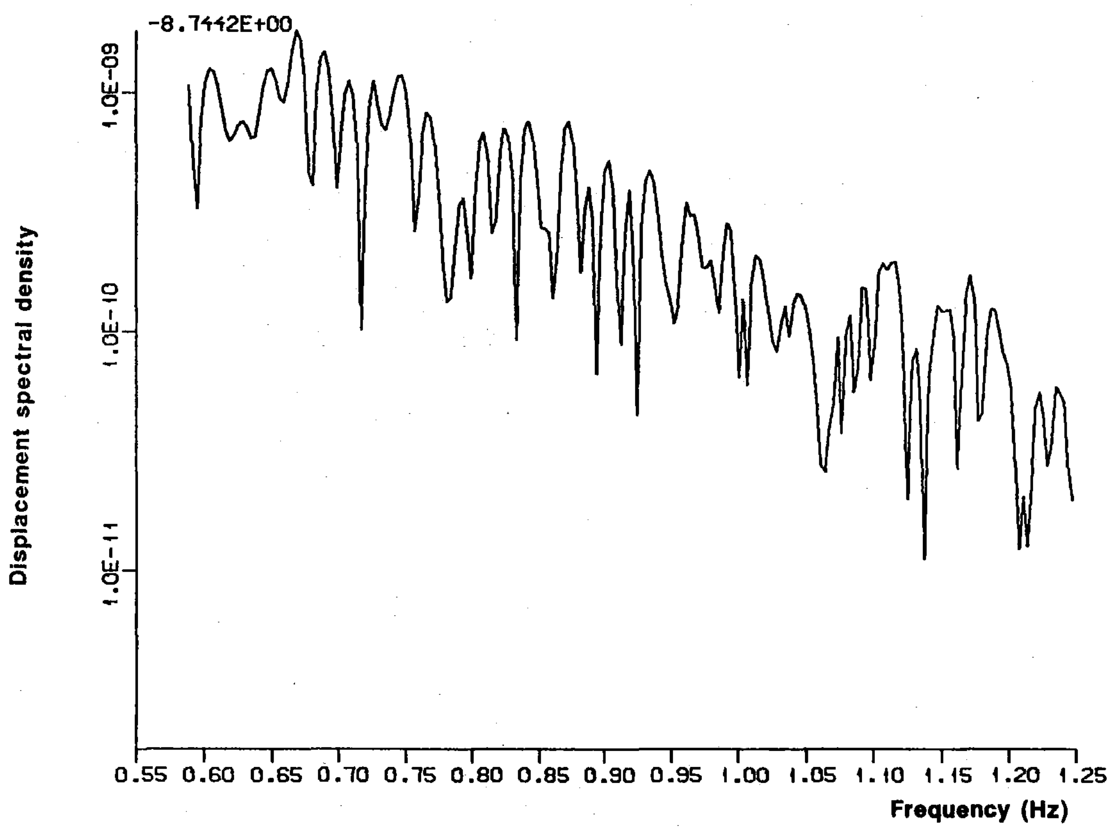


Fig. VII.3.5. Synthetic spectrum for an explosion at 0.9 km focal depth in model 1 and recorded at NORSAR.

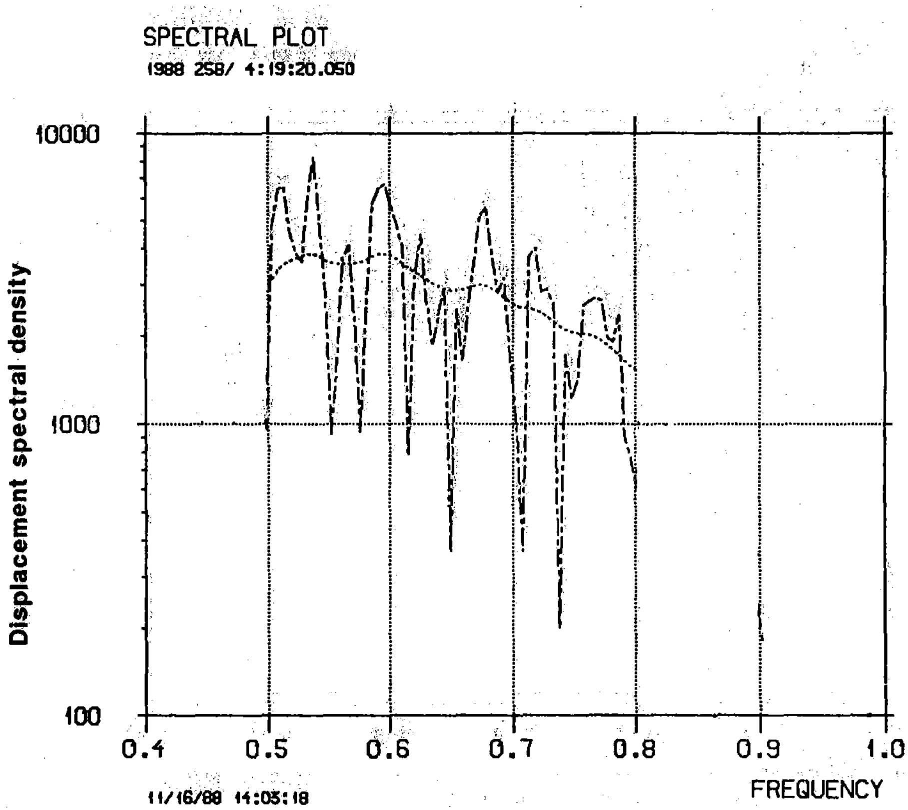


Fig. VII.3.6. Spectrum of the Lg wave recorded at NORESS shown in the frequency interval with best SNR (dashed-dotted line), and associated smoothed spectrum (dotted line).

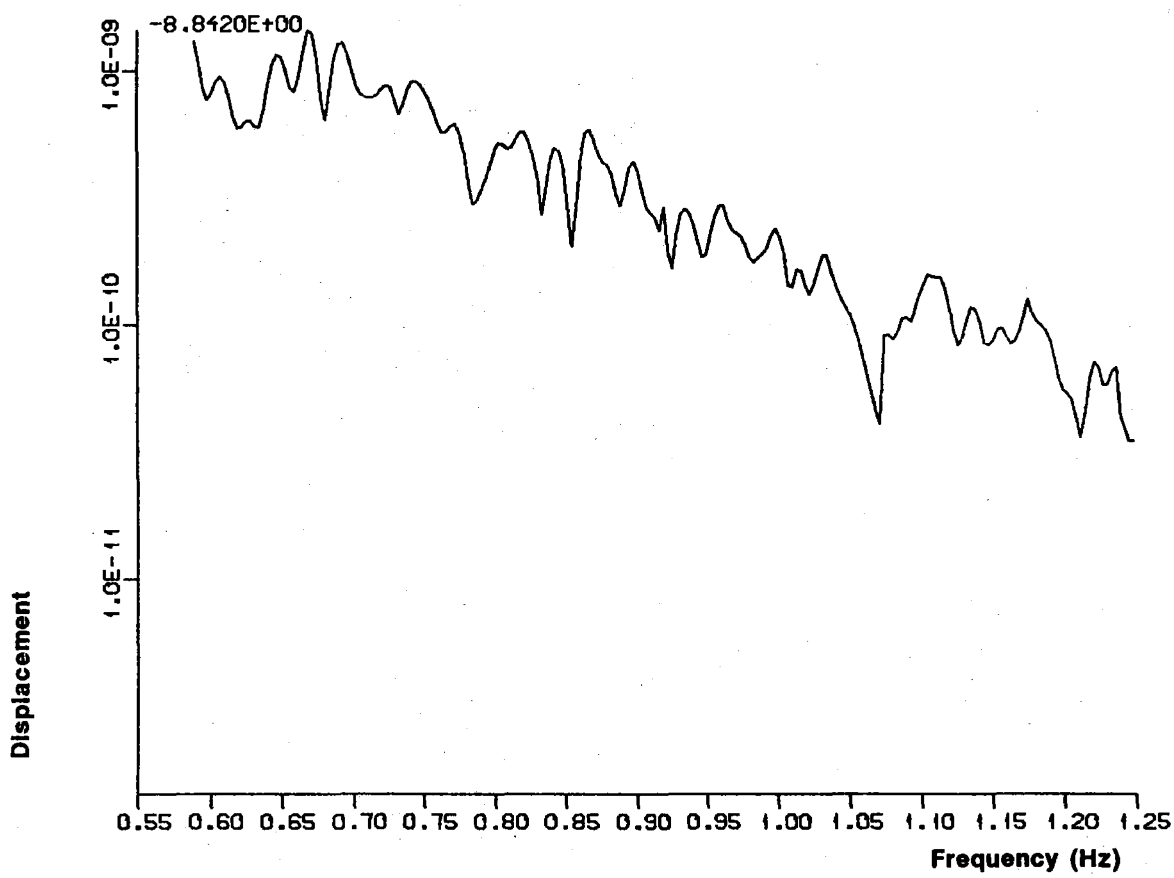
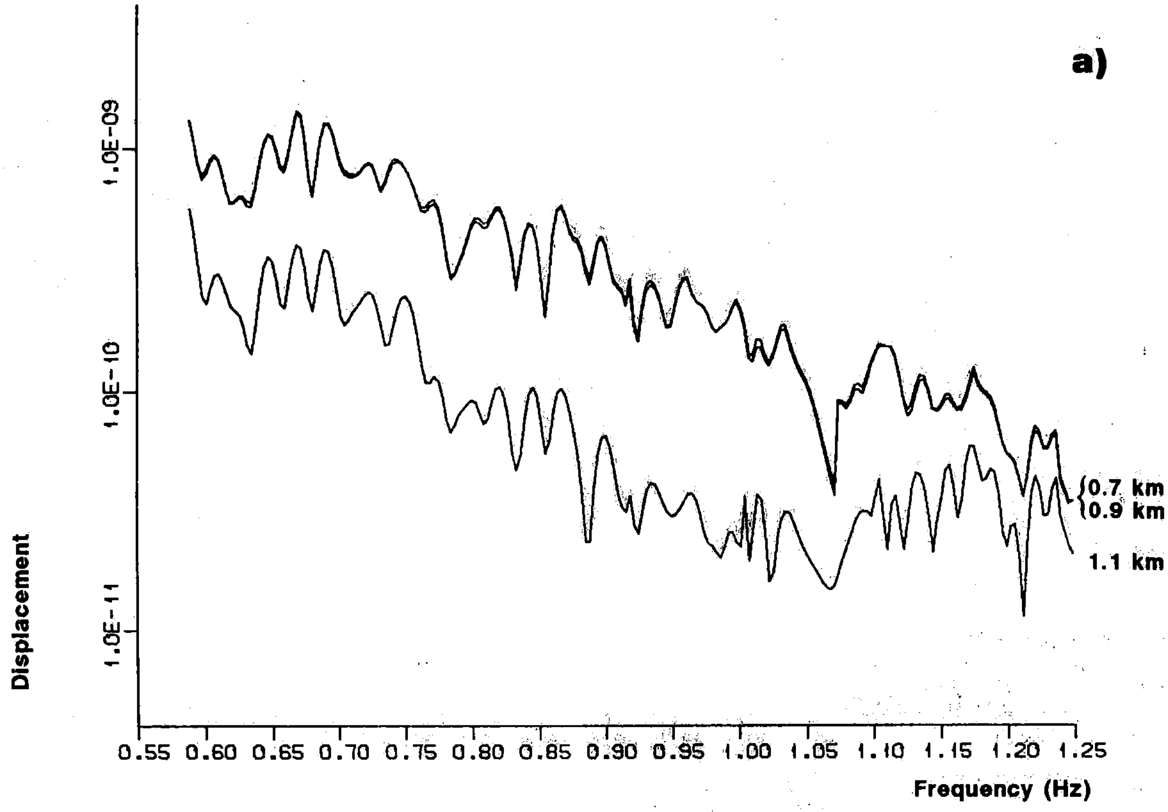


Fig. VII.3.7. Synthetic spectrum of Fig. VII.3.5 after smoothing by addition of spectra from stations within a 60 km window around the central station.



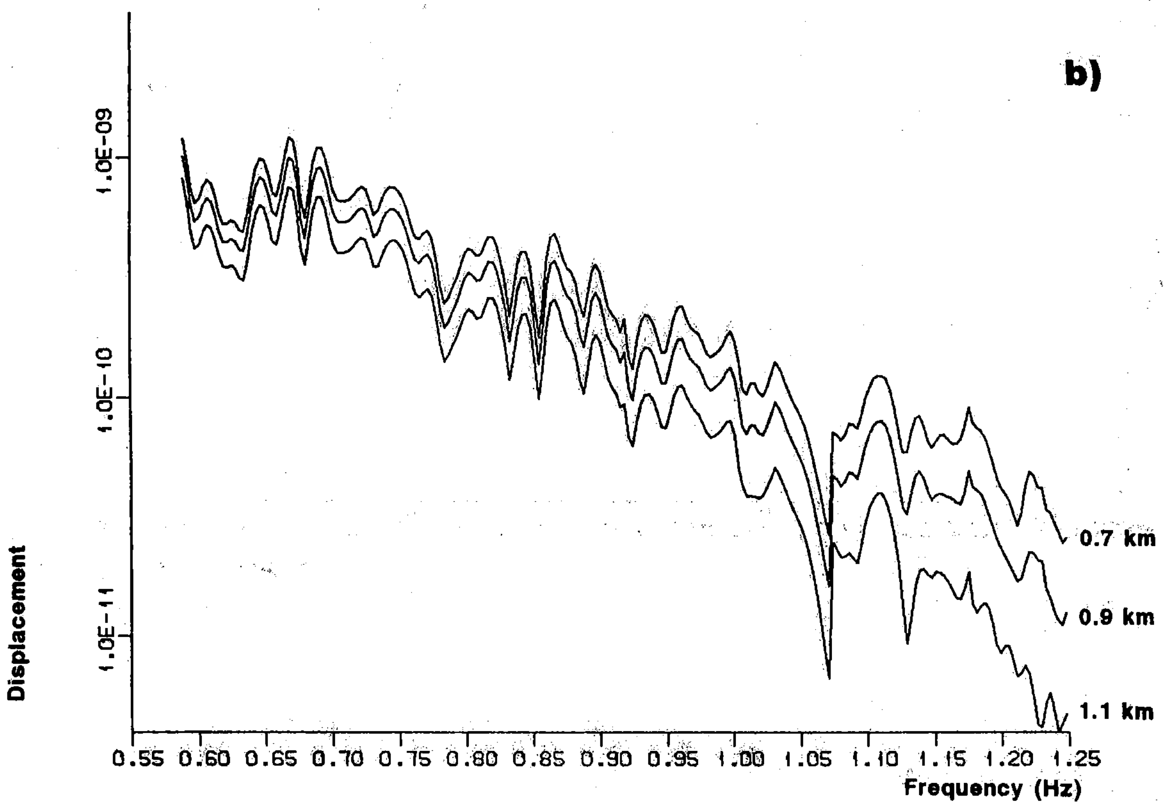
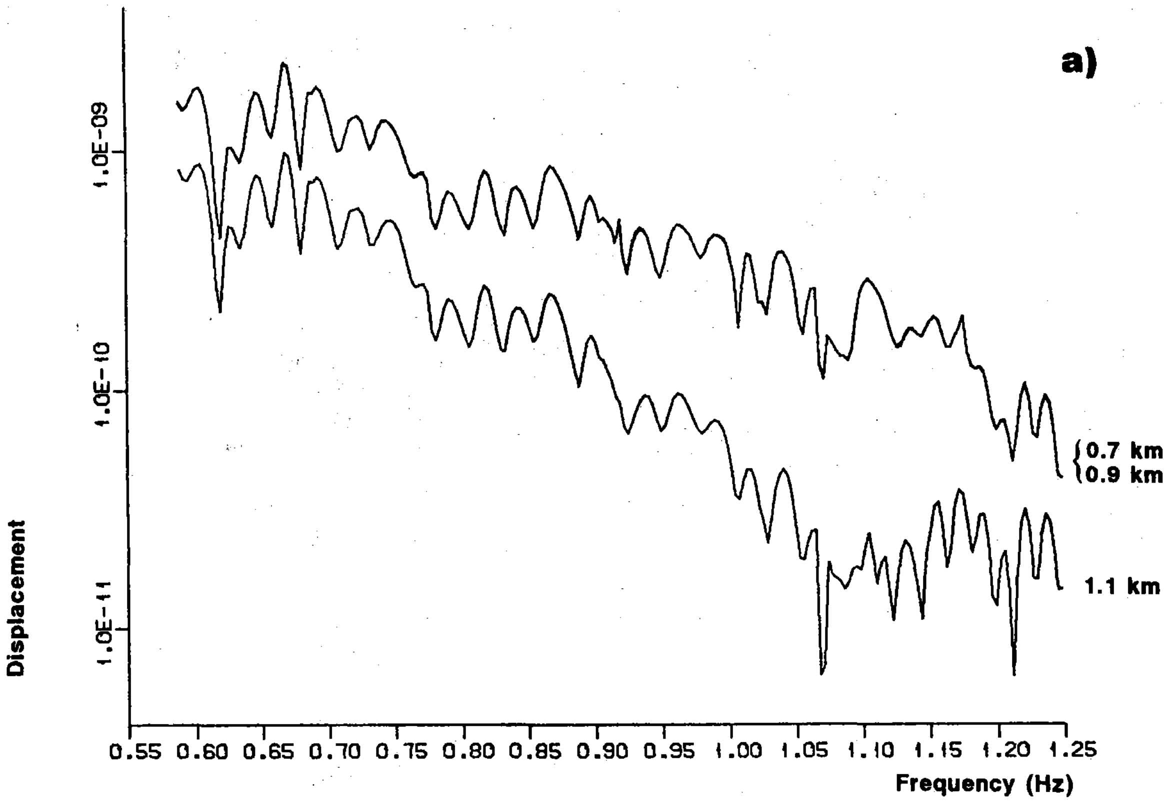
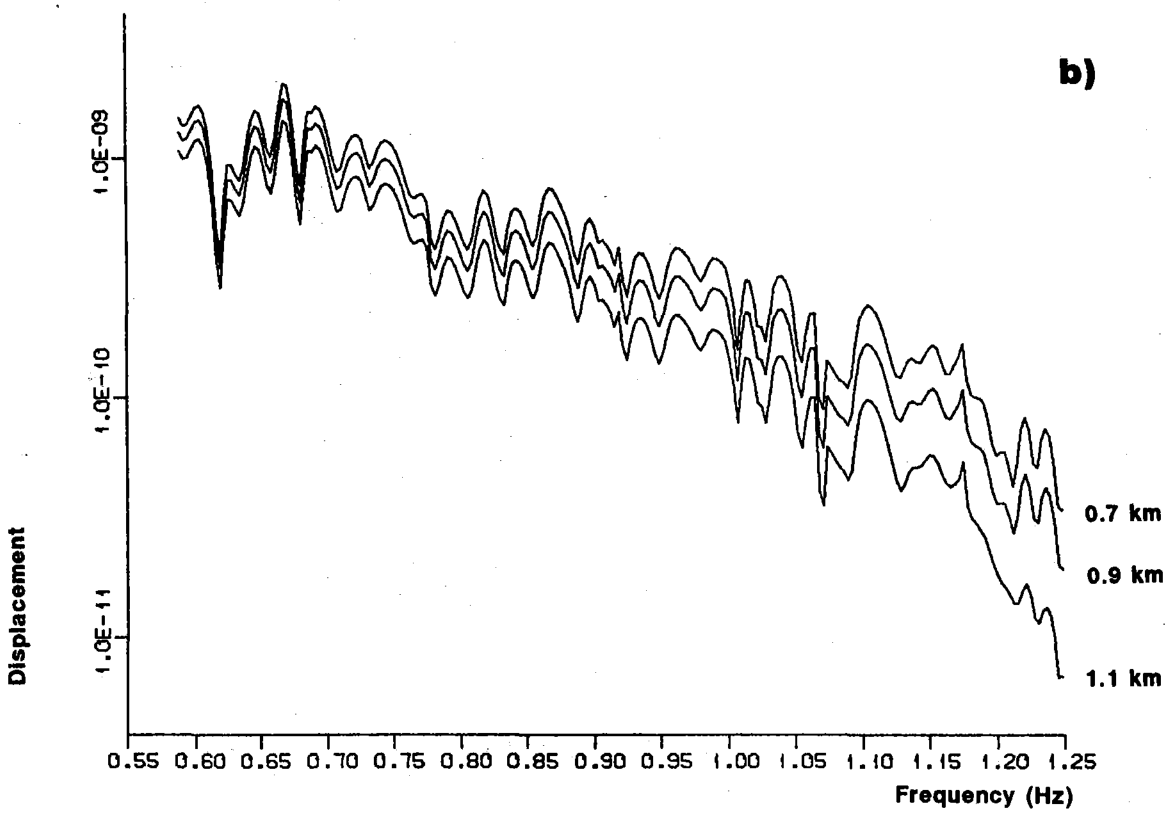


Fig. VII.3.8. Smoothed synthetic Lg spectra at NORSAR from explosions at 0.7, 0.90 and 1.1 km focal depth in model 1 (plot a) and in model 1′ (plot b).





<u>Fig. VII.3.9.</u> The same as Fig. VII.3.8 for model 2 (plot a) and model 2′ (plot b).

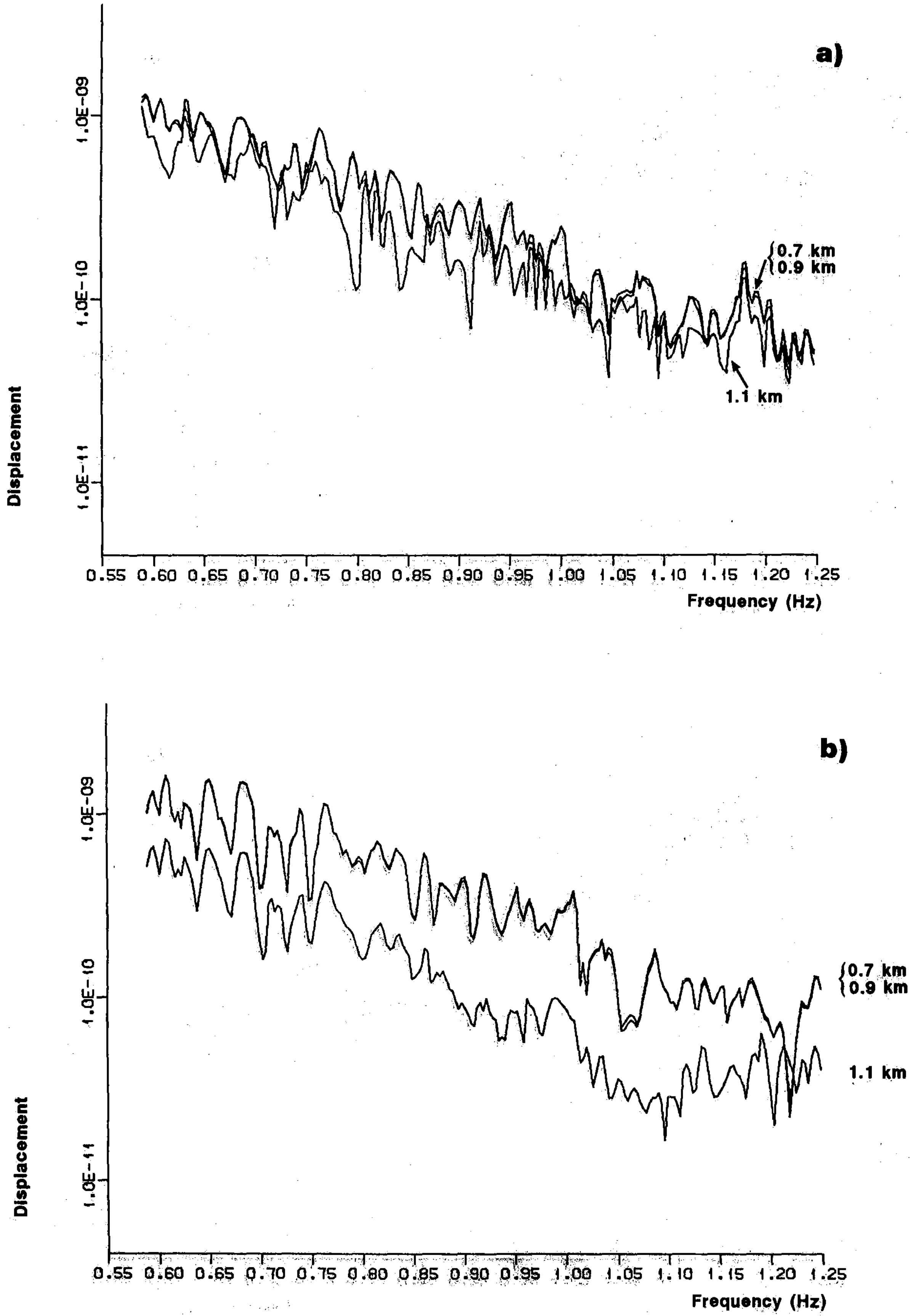


Fig. VII.3.10. Smoothed synthetic Lg spectra from explosions at 0.7, 0.9 and 1.1 km in model 1 (plot a) and model 2 (plot b), assuming complete propagation and recording in model 1.