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Phase Velocity of the Arabian Platform and the Surface Waves Attenuation Characteristics by Wave Form Modeling

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ABSTRACT. The crustal structure beneath the Arabian platform has been investigated using available seismic records produced at the Riyadh (RYD) Station. The phase velocities along 14 individual propagation paths were determined using the phase matched filter technique. The phase velocities were found to increase from 2.3 km/s at 4 s to 4.1 km/s at 42 s. Phase and group velocities were combined and used to invert for an S wave seismic crustal velocity structure. The inversion process produced a model with β (the shear velocity) between 1.48-2.41 km/s in the top 3 km which represents the sedimentary cover rocks. The upper crust β increases from 3.1 at 5 km to 3.67 km/s at 25 km depth. The Lower crust is about 20 km thick, and its β increases rapidly from 3.76 km to 4.43 km/s. The Moho discontinuity is at about 45 km depth with an average β of about 4.7 km/s. The resulting shear velocity model was then combined with the Q model representative of the region, and synthetic seismograms were computed. Comparison between the observed data and theoretical results showed that the shear wave velocity model of the Arabian platform obtained in this study produces excellent match between recorded and computed seismograms. This procedure is suggested as a tool that can be used to study the attenuation characteristics of the Arabian platform as well as the Arabian plate.

Introduction

The crustal structure of the Arabian plate has been the focus of attention of many scientists around the world for the past several decades. Several studies have demonstrated the continental nature of the crust of the Arabian plate (*e.g.*, Niazi (1968), Mooney *et al.* (1985), Badri (1991), Ghalib (1992), Mechie *et al.* (1986), Mokhtar *et al.* (1988) and others). Figure 1 shows some examples of the crustal structures obtained in these studies. More recently, Mokhtar and Al-Saeed (1994) deduced the shear velocity model of both the Arabian shield and the Arabian platform (Fig. 1). They used seismological data from RYD Station and computed the group velocities of Rayleigh and Love waves along three propagation paths from southern Red Sea, Gulf of Aden, and southern Iran to Riyadh RYD station in the middle of the Arabian plate. Their results indicate that the shear wave velocity in the Arabian platform can be represented by a two layer crust with a shear wave velocity of 3.38 km/s, and 4.0 km/s in the upper and lower layers, respectively. The average total thickness of the crust is 45 km beneath the platform. These results were obtained by inversion of the group velocities only along the southern Iran-RYD propagation path.

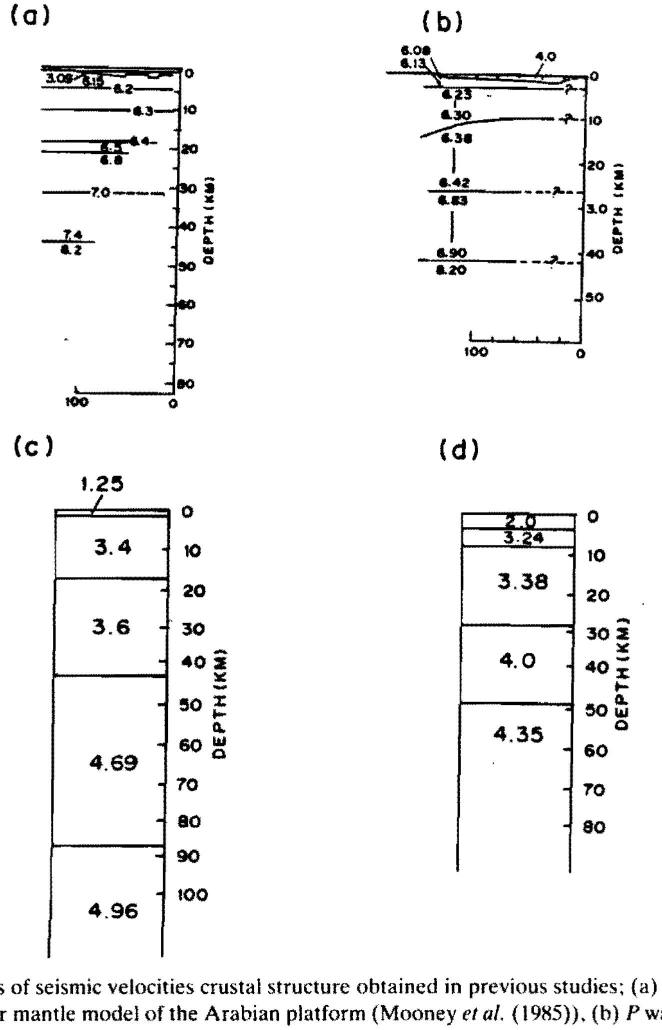

FIG. 1. Examples of seismic velocities crustal structure obtained in previous studies; (a) P wave crustal and upper mantle model of the Arabian platform (Mooney *et al.* (1985)), (b) P wave crustal and upper mantle model of the Arabian platform (Badri (1991)), (c) S wave crustal and upper mantle structure of the Arabian peninsula (Niazi (1968)), and (d) S wave crustal and upper mantle model of the Arabian platform (Mokhtar and Al-Saeed (1994)).

It is well known that inversion of phase velocity data alone produces a non unique solution because of the trade-off between velocities and layer thicknesses, and inversion of group velocities alone produces an even more non unique solution resulting from the derivative relationship between phase and group velocities (Pilant and Knopoff, 1970). Although phase and group velocities are not completely independent variables, they do provide slightly differing sensitivities to a given structure and can be used simultaneously as independent measurements to increase resolution (Wiggins, 1972).

In this study, data obtained from RYD Station are used to compute the phase velocity of the fundamental mode Rayleigh and Love waves. The phase velocities are used simultaneously with the group velocities of both phases determined by Mokhtar and Al-Saeed (1994) to invert for the crustal structure of the Arabian platform.

Synthetic seismograms are then computed for selected events in order to test the resulting inverted velocity model.

Phase Velocity Measurements

Sâto (1955 & 1956) first applied Fourier analysis to a seismogram to determine the phase velocity of seismic surface waves for a wave train which traveled a known distance from the epicenter. In order to avoid the difficulty associated with the source phase shift, data can be used from two stations. Assuming that both stations are at the same azimuth from the epicenter, then $C(\omega)$, the phase velocity of the medium between the two stations is given by

$$C(\omega) = \frac{\omega}{K(\omega)} = \frac{\omega(r_2 - r_1)}{\phi_2 - \phi_1 - \phi_{i2} + \phi_{i1} + 2l\pi} \tag{1}$$

Where r_1, and r_2 are the epicentral distances of the two stations from the source, ϕ_{i1}, and ϕ_{i2} are the instrument response of the two stations, ϕ_1, and ϕ_2 are the phase spectrum of the seismograms at the two stations. Equation (1) is simplified further if the same type of instruments are used at the two stations. l is an integer which is estimated from the knowledge of the possible range of phase velocity values. The usual technique of selecting l is to find a particular value that gives reasonable values of phase velocities at the longest periods, and to assume that the phase velocity curve changes smoothly with period. However, the uncertainty in this parameter usually leads to errors in the values of the phase velocities.

Phase-matched filters technique (PMF) can be utilized to obtain an estimate of the phase velocity of surface waves from a single station. The method is based on the use of a class of linear filters in which the Fourier phase of the filter is made equal to that of a given signal. PMF was proposed by Herrin and Goforth (1977) to clean up the amplitude spectrum of the primary signal by identifying and removing the multiple arrivals and allowing the recovery of the complex spectrum of the primary wave train along with its apparent group velocity dispersion curve. Herrin and Goforth (1977) considered the convolution and cross-correlation of a signal, $s(t)$, with a time function, $f(t)$, as follow

$$s(t) * f(t) \Longleftrightarrow S(\omega)\ F(\omega)\ e^{i[\sigma(\omega) + \phi(\omega)]} \tag{2}$$

and

$$s(t) \times f(t) \Longleftrightarrow S(\omega)\ F(\omega)\ e^{i[\sigma(\omega) - \phi(\omega)]} \tag{3}$$

where, $*$ indicates convolution,

$\times$ indicates cross-correlation, and

$|S(\omega)|$ indicates the amplitude spectrum of $S(\omega)$, and

$\Longleftrightarrow$ indicates the right hand side is the Fourier transform of the left hand side. The Fourier transform of the left hand side. The Fourier transforms of $f(t)$ and $s(t)$ are given by

$$F(\omega) = \int_{-\infty}^{\infty} f(t)\ e^{i\omega t}\ dt \tag{4}$$

$$S(\omega) = \int_{-\infty}^{\infty} s(t)\ e^{i\omega t}\ dt \tag{5}$$

Now suppose that we choose $f_p(t)$ such that the Fourier phase is the same as that of $s(t)$, the output of the cross-correlation operation will then have the Fourier transform $|S(\omega)||F_p(\omega)|$ which is an even function in the time domain and is known as the pseudo-auto correlation function. Goforth and Herrin (1979) described an iterative technique which can be used to find a phase match filter when the exact form of the signal is unknown such as the case of the seismic signal. In this technique, an initial estimate of the phase of the filter is obtained from the group delay which is a function of the group velocity. It is assumed that the seismic signals result from a dispersive process with a continuous dispersion curve. The group delay, $T_{gr}(\omega)$ associated with the signal is the epicentral distance divided by the group velocity at frequency ω minus the signal delay. From Papoulis (1962), it is noted that the group delay and Fourier phase of the signal are related as follows

$$t_{gr}(\omega) = \frac{d\,\theta(\omega)}{d\,\omega} \tag{6}$$

$$\theta(\omega_1) = \int_0^{\omega_l} \omega_l\ t_{gr}(\omega)\ d\,\omega \tag{7}$$

Thus a trial group velocity dispersion curve and an amplitude spectrum of the signal are used as an input. Using the epicentral distance to the signal source and Equation (7) above we compute the Fourier phase of the filter $f(t)$ and perform the correlation in the frequency domain. The result is then transformed to the time domain, windowed to reject correlation functions from interfering signals or multipath arrivals, then transformed again. The result will have the complex spectrum

$$|S(\omega)|\ |F(\omega)|\ e^{i[\sigma(\omega) - \phi(\omega)]} \tag{8}$$

The difference in phase is used to correct the group delay of the trial filter, and the process is repeated until the phase spectra of the filter and the desired signal in the band of interest are identical. PMF technique application to surface wave data from a single station has been described in detail by Russell (1987) and the Herrmann (1987). MATCH computer program (Herrmann (1987)) has been used in this study to estimate the phase velocities of the surface waves. The program works interactively and displays the computed phase velocities as well as the pseudo auto-correlation function and the time series of both the original trace and that of the phase matched filter.

Phase Velocity Results for the Arabian Platform

Since we have only RYD Station, we have to use the single station method to determine the phase velocities along each propagation path. Measuring the phase velocity from one station is rather complicated by the requirements of determining the unknown source phase. It is true that in using single station technique, one cannot eliminate the source phase which can be eliminated by the two station technique. However, it is known that this effect will lead to considerable errors if the distances were very short and the periods under consideration were long. According to Knopoff and Schwab (1968), the magnitude of the error in a phase velocity of 4.0 km/s at 50 seconds period deduced from a station say 1000 km away from the source is 0.033 km/s. The error in a group velocity of 3.0 km/s at the same period and for the same distance is 0.026 km/s. The data used from RYD Station, show that the longest period considered is about 42 seconds and the shortest distance is about 750 km. Thus, the errors due to the shift in the source phase on the measured velocities are much smaller than those due to epicentral mislocation and errors in origin times for this period and epicentral distances.

Fourteen long period seismogram records for the events of Table 1 were obtained from RYD Station and were used to determine the phase velocities of the fundamental mode surface waves within the Arabian platform along the paths shown in Fig. 2. Data processing and the different steps that were taken to correct the long period seismograms are explained in detail in Mokhtar and Al-Saeed (1994). The phase velocities of Rayleigh and Love waves along the studied propagation path are shown in Fig. 3 and 4, and are listed in Tables 2 and 3. These values were determined by taking the mean of the individual earthquake-station paths. The error bars represent the standard deviation from the mean.

The phase velocities of Rayleigh waves along the studied paths increase rapidly at the short periods from 2.3 km/s at 4 seconds to about 2.9 at 10 seconds. The increase in phase velocity beyond 10 seconds is more or less slower and the phase velocity reaches 4.1 km/s at 42 seconds. Similar results holds for the Love waves as shown in Fig. 4 and Table 3.

These results were combined with the group velocities results of the Arabian platform determined by Mokhtar *et al.* (1993), and both phase velocities of the fundamental mode and the group velocities of the fundamental and first higher modes

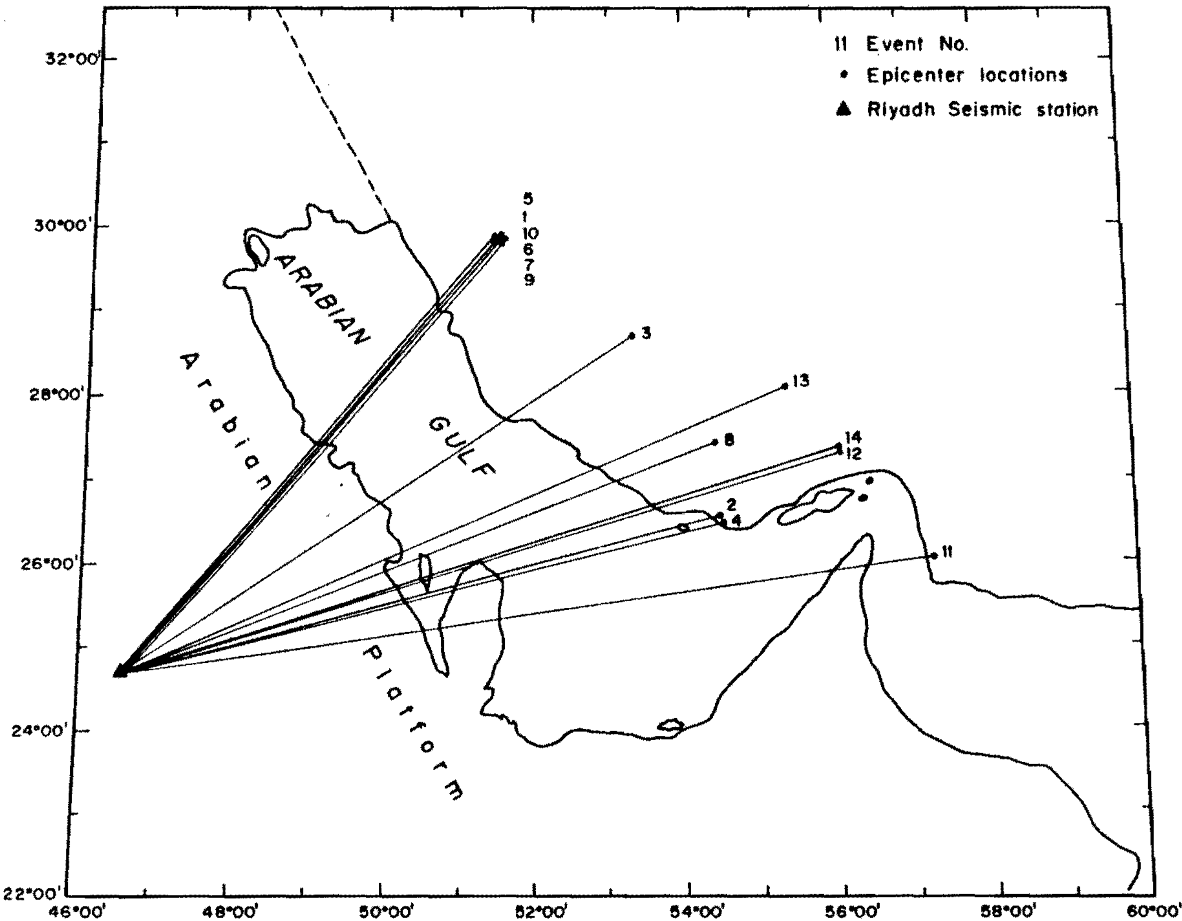

FIG. 2. Propagation paths of surface waves across the Arabian platform along which the phase velocities and surface waves attenuation characteristics are studied.

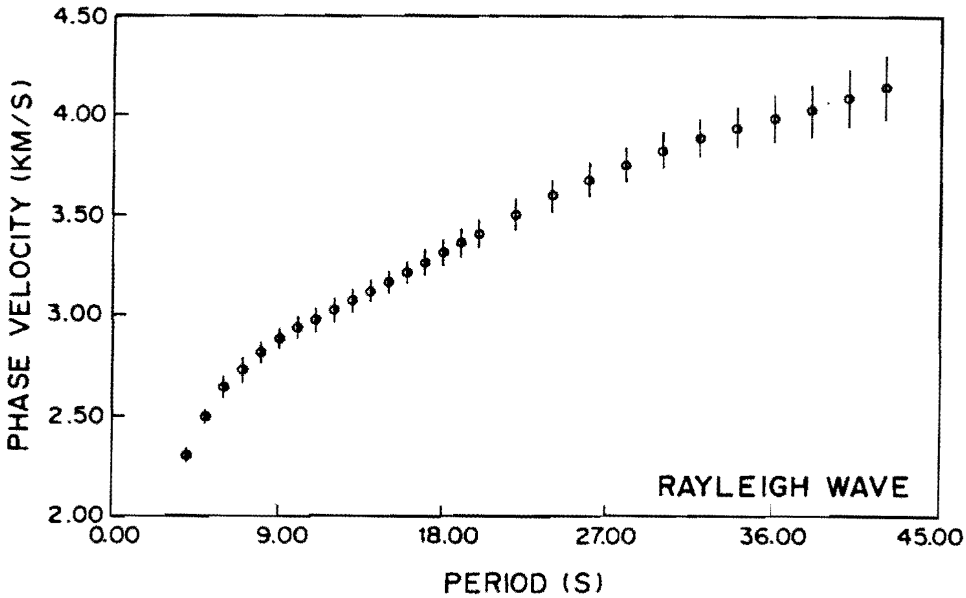

FIG. 3. Rayleigh wave phase velocity dispersion curve. The small circles represent the average values of the phase velocities along the different paths. The vertical lines represent the standard deviations.

TABLE 1. Parameters of earthquakes used in this study (Obtained from the Epicentral Data Report (EDR) published by the USGS).

Event no.	Date	Origin time (UTC)	Latitude (N°)	Longitude (E°)	Depth (km)	Magnitude	
						m_b	M_s
1	13 July 1986	00:48:45.55	29.986	51.520	10	4.9	4.6
2	09 Aug. 1986	06:37:24.04	26.748	54.969	33	5.1	4.4
3	01 Oct. 1986	03:57:52.95	28.815	53.311	10	4.5	–
4	06 Oct. 1986	02:21:46.42	26.670	54.580	60	4.9	4.5
5	18 Oct. 1986	08:48:30.78	29.989	51.454	54	4.9	3.9
6	20 Nov. 1986	10:09:07.86	29.869	51.585	16	4.8	–
7	20 Nov. 1986	20:08:01.60	29.983	51.643	31	5.2	4.1
8	14 Dec. 1986	09:09:18.09	27.526	54.363	33	4.8	–
9	20 Dec. 1986	23:47:08.91	29.985	51.623	25	5.4	5.1
10	11 Jan. 1987	12:31:26.06	29.969	51.788	9	4.8	4.1
11	18 Feb. 1987	20:46:57.43	26.091	57.361	33	4.4	–
12	29 Apr. 1987	01:45:22.63	27.437	56.109	8	5.9	5.3
13	12 May 1987	07:15:13.10	28.163	55.534	40	5.2	4.9
14	29 Apr. 1987	01:45:22.63	27.437	56.109	40	5.9	5.3

were used to obtain a seismic velocity model for the *S* wave. Details of the inversion procedure can be found in Mokhtar *et al.* (1993) and Mokhtar and Al-Saeed (1994). The shear velocity model obtained is similar to that obtained by Mokhtar and Al-Saeed (1994), and is shown in Fig. 5. The fit between the observed and theoretical dispersion of Love and Rayleigh waves are shown in Fig. 6 and 7.

TABLE 2. Rayleigh wave phase velocity results.

Mode	Period (s)	Phase velocity (km/s)	Standard deviation
Fundamental	4.00	2.298	0.037
	5.00	2.487	0.039
	6.00	2.634	0.060
	7.00	2.718	0.066
	8.00	2.805	0.056
	9.00	2.875	0.054
	10.00	2.930	0.056
	11.00	2.967	0.063
	12.00	3.014	0.063
	13.00	3.062	0.062
	14.00	3.108	0.059
	15.00	3.153	0.059
	16.00	3.200	0.061
	17.00	3.249	0.066
	18.00	3.299	0.070
	19.00	3.348	0.074
	20.00	3.396	0.076
	22.00	3.491	0.081
	24.00	3.583	0.085
	26.00	3.668	0.089
	28.00	3.745	0.091
	30.00	3.816	0.093
	32.00	3.879	0.098
	34.00	3.933	0.106
	36.00	3.977	0.122
	38.00	4.018	0.135
	40.00	4.083	0.148
	42.00	4.135	0.164

Wave Form Modeling of The Observed Data From RYD Station

The generation of synthetic seismograms for point sources in simply layered structures has made rapid advances in the past decades. Two approaches involving Laplace transform and Fourier transform techniques were pursued. The Cagniard-de Hoop technique is based on the Laplace transform and is usually referred to as the generalized ray method (Helmberger, 1968). In this method the solution is constructed by tracking the individual seismic arrivals ray by ray from the source to receiver. This method is valid at high frequencies and works well at predicting particular phases, but is poorly suited to models with many layers and larger distances when a complete seismogram is desired. The other approach involves expressing the solutions in terms of a double integral transformation over wave number and frequency (Hudson, 1969). This is the full wave theory approach, in which the complete solution, rather than the individual rays, is considered. The full wave integration theory is explained in detail in Herrmann (1978, 1979), Wang and Herrmann (1980), and Wang (1981).

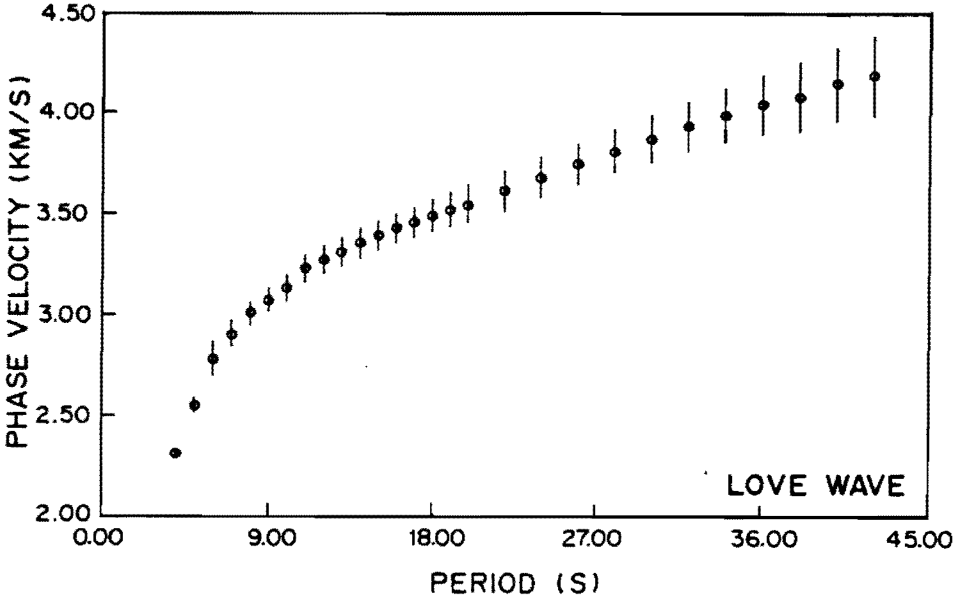

FIG. 4. Love wave phase velocity dispersion curve. See Fig. 3 for explanations.

Seber and Mitchell (1992) have studied the effects of the different source and fault parameters on surface wave amplitude spectra. The effects of focal depth, source time function, fault parameters (dip, strike, and slip), and the effect of Q were studied. The spectra were computed for certain cases in which each parameter was allowed to vary through a range of values while holding the others fixed at certain values. The effect of depth was found to influence the energy at shorter periods of surface waves. The Love wave spectrum for the longer periods is much less sensitive to the focal depth than it is for the shorter periods. The Rayleigh wave spectrum, however, were found sensitive to the focal depth at both short and long periods.

The effect of fault parameters were found to affect the Love and Rayleigh waves with most significant changes occurring at the longer periods, while the effect of the crustal Q values was found to be dominant at short periods. In particular, Q_β values for the upper crust have a large effect on the amplitude spectra and it was found to play the most important role in shaping the Rayleigh and Love wave spectra especially at short periods.

In this study, the full wave integral method as described by Herrmann and Wang (1985) is used to compute synthetic seismograms of surface waves in plane layered medium due to a point source. The medium consists of homogeneous plane layers overlying an elastic halfspace. Wang and Herrmann (1980) used the wave theory integral method to obtain 10 Green's functions required to describe the wave field due to an arbitrary point dislocation source. The Green's functions were derived by introducing the compound matrices in the formulation of Haskell (1964). The implementation of the compound matrices and the use of a suitable numerical integra-

TABLE 3. Love wave phase velocity results.

Mode	Period (s)	Phase velocity (km/s)	Standard deviation
Fundamental	4.00	2.310	0.025
	5.00	2.548	0.041
	6.00	2.778	0.088
	7.00	2.901	0.068
	8.00	3.000	0.061
	9.00	3.070	0.064
	10.00	3.126	0.070
	11.00	3.225	0.070
	12.00	3.265	0.074
	13.00	3.303	0.077
	14.00	3.346	0.076
	15.00	3.384	0.075
	16.00	3.419	0.075
	17.00	3.451	0.077
	18.00	3.481	0.082
	19.00	3.510	0.089
	20.00	3.540	0.096
	22.00	3.604	0.104
	24.00	3.669	0.105
	26.00	3.734	0.105
	28.00	3.797	0.109
	30.00	3.859	0.117
	32.00	3.917	0.126
	34.00	3.972	0.136
	36.00	4.026	0.151
	38.00	4.066	0.176
	40.00	4.130	0.189
	42.00	4.172	0.202

tion method has made it possible to obtain high quality synthetic seismograms at a wide range of frequencies and for relatively complex seismic models.

Synthetic seismograms for the long period records of the events listed in Table 4 were computed. The strike, dip, and slip are obtained from the best double couple mechanism of the moment tensor solution published by the PDE Bulletins. The depth was also taken from the hypocentral location of the same bulletin. The focal mechanism parameters for the 13 Jul 1986 event are those for an earthquake that took place on 12 Jul 1986 from the same location. Since the mechanism of the event on the 13*th* were not published, I used the main event mechanism assuming that it is representative of the aftershock.

The resulting synthetic seismograms were computed using the above information and the inverted shear wave velocity models for the different paths. In addition, a general Q model that consists of upper crust with compressional wave quality factor $Q_\alpha = 160$ and a lower crust with $Q_\alpha = 300$ was used. The thicknesses of these two

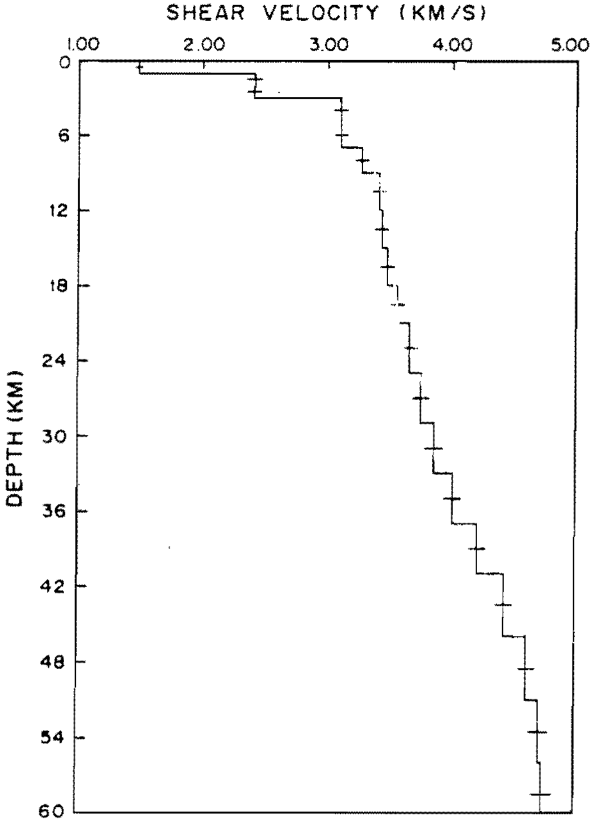

FIG. 5. Inverted shear wave crustal velocity structure for the Arabian platform. The model shows that the top 3 km is representative of the sedimentary cover rocks. The upper crust is considered to extend to about 25 km, and the Moho discontinuity is taken at 45 km (Mokhtar *et al.* (1993)).

TABLE 4. Earthquake source parameters used in computing synthetic seismograms.

Event no.*	Date	Lat.	Lon.	Depth	Orig. time	Mag.	Dip	Slip	Strike
1#	13 Jul. 1986	29.99N	51.52E	10.0	00:48:45.55	4.9	81	172	178
9	20 Dec. 1986	29.99N	51.62E	25.5	23:47:08.91	5.5	70	− 179	348
13	12 May 1987	28.17N	55.56E	40.2	07:15:13.10	5.2	34	104	278
	29 May 1987	34.08N	48.27E	40.9	06:27:50.77	4.9	80	2	218

* Event number as indicated in Table (1).

\# Event source parameter is that of the event that occurred on 12 July 1986 in the same location of the epicenter and in the same tectonic settings as listed in the Epicentral Data Report (EDR).

layers were fixed at 20 km each and they are underlain by an upper mantle lithosphere (or lid) with Q_α fixed at 1000. These values were modified from the models by Badri (1991) (personnel communications). Seber (1990) found that the average crus-

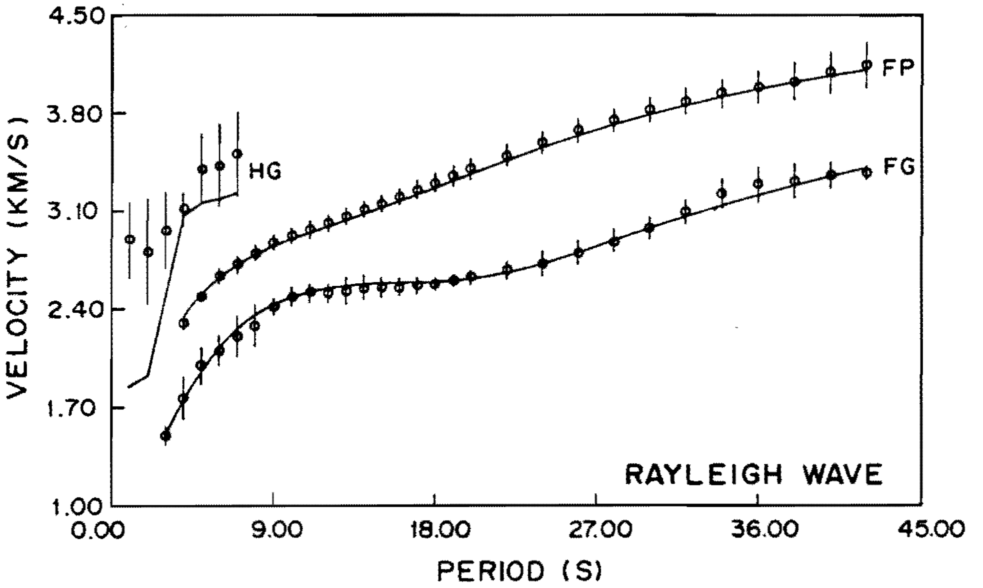

FIG. 6. The fitting between observed and theoretical Rayleigh wave dispersion data for the model in Fig. 5. FG is the fundamental mode group velocity curve, FP is the fundamental mode phase velocity curve, and HG is the first higher mode group velocity curve.

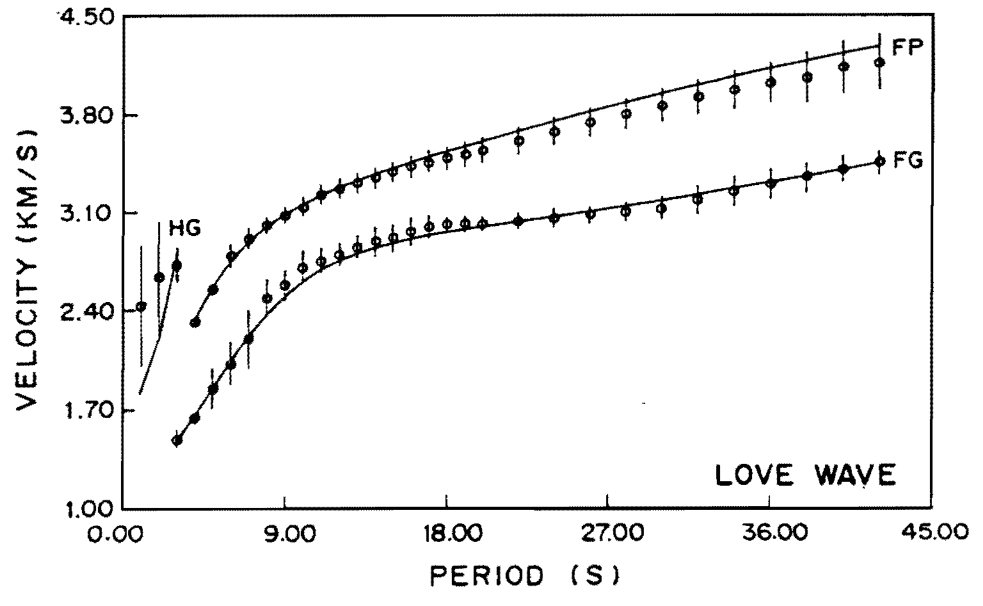

FIG. 7. The fitting between observed and theoretical Love wave dispersion data for the model in Fig. 5. FG is the fundamental mode group velocity curve, FP is the fundamental mode phase velocity curve, and HG is the first higher mode group velocity curve.

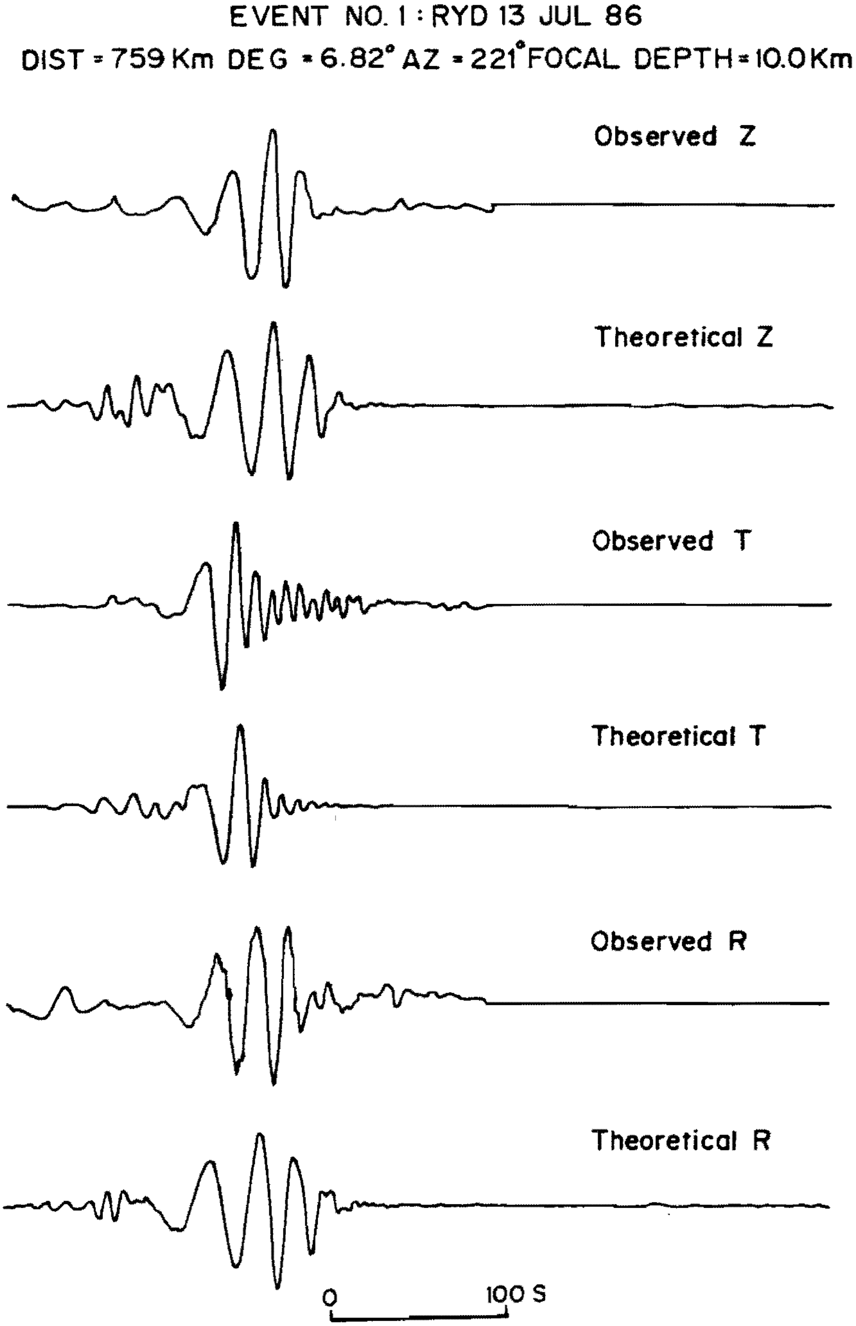

FIG. 8. Observed and synthetic seismograms of the long period components of RYD Station for the 13 July 1986 earthquake (event No. 1). *Z* is the vertical component, *T* is the transverse component, and *R* is the radial component.

tal Q_β values across the Arabian Peninsula vary between 50 and 150. In the coastal regions Q_β values are smaller than those of the inland regions. Q_β, the shear wave quality factor was taken to be about $\frac{1}{2}$ Q_α. Figure 8 represents the results of the wave form modeling of the Long period Z (vertical), T (transverse), and R (radial) components of the earthquake of 13 July 1986 (event No. 1). Observed and synthetic seismograms for the 20 December 1986 (event No. 9) event are shown in Fig. 9. The synthetics seismograms for the same event in Fig. 10 were computed using a depth of 10.0 km instead of 25.0 km. Figure 11 is the observed and synthetic seismograms of the earthquake on 12 May 1987 (event No. 13) and Fig. (12) is for the same event using a depth of 10.0 km. Figure 13 shows the observed and synthetic seismograms of the event on 29 May 1987 (event No. 14).

In all of these cases, the waveform modeling shows that the shear velocity model that has been obtained by Mokhtar and Al-Saeed (1994) and by this study produces synthetic seismograms that are very consistent with the observed seismograms. In addition, the upper crustal Q_β used is close to the values reported by Seber and Mitchell (1992) for the Arabian platform (65-85). Using Q_β of lower or higher values have resulted in poor matching between observed and synthetic seismograms. Figure 14 summaries the shear wave crustal structure and the Q_β models used in the production of the synthetic seismograms in this study.

Discussion and Conclusion

The decay of surface wave amplitude has been studied much less frequently because of the difficulty in making meaningful determinations (Mitchell *et al.* (1977)). The attenuation of surface wave amplitudes with distance may be produced by a number of factors, including geometrical spreading, scattering, lateral refraction, multipathing and anelasticity of the medium through which the wave travels. It is important, however, to try to obtain amplitude attenuation data because they provide a measure of the anelastic properties of the earth. Seismic ground motion hazard evaluations are usually based on probabilistic models which take into account several factors such as the spatial distribution, the recurrence rate of earthquakes, and the ground motion attenuation model for the region under investigation. The later is a critical factor in these studies and research has shown that much of the variability in estimates of the peak ground acceleration values are due to uncertainties inherent in ground attenuation models. Due to the lack of Q values in Arabia, several recent seismic risk analyses of the region have used attenuation models developed for regions that have similar tectonic settings (*e.g.*, Thenhause *et al.* (1986), Al-Noury and Ali (1986). Thenhause *et al.* (1986), for example, preferred a low Q model (Q 180 at 1 Hz) developed for western United States of America over the high Q model characteristic estimates of the peak ground acceleration in western Arabia.

In this study a straight forward method has been presented to estimate the attenuation characteristics of the crust of the Arabian platform. Unfortunately, the absolute magnification of the RYD Station long period component is not available.

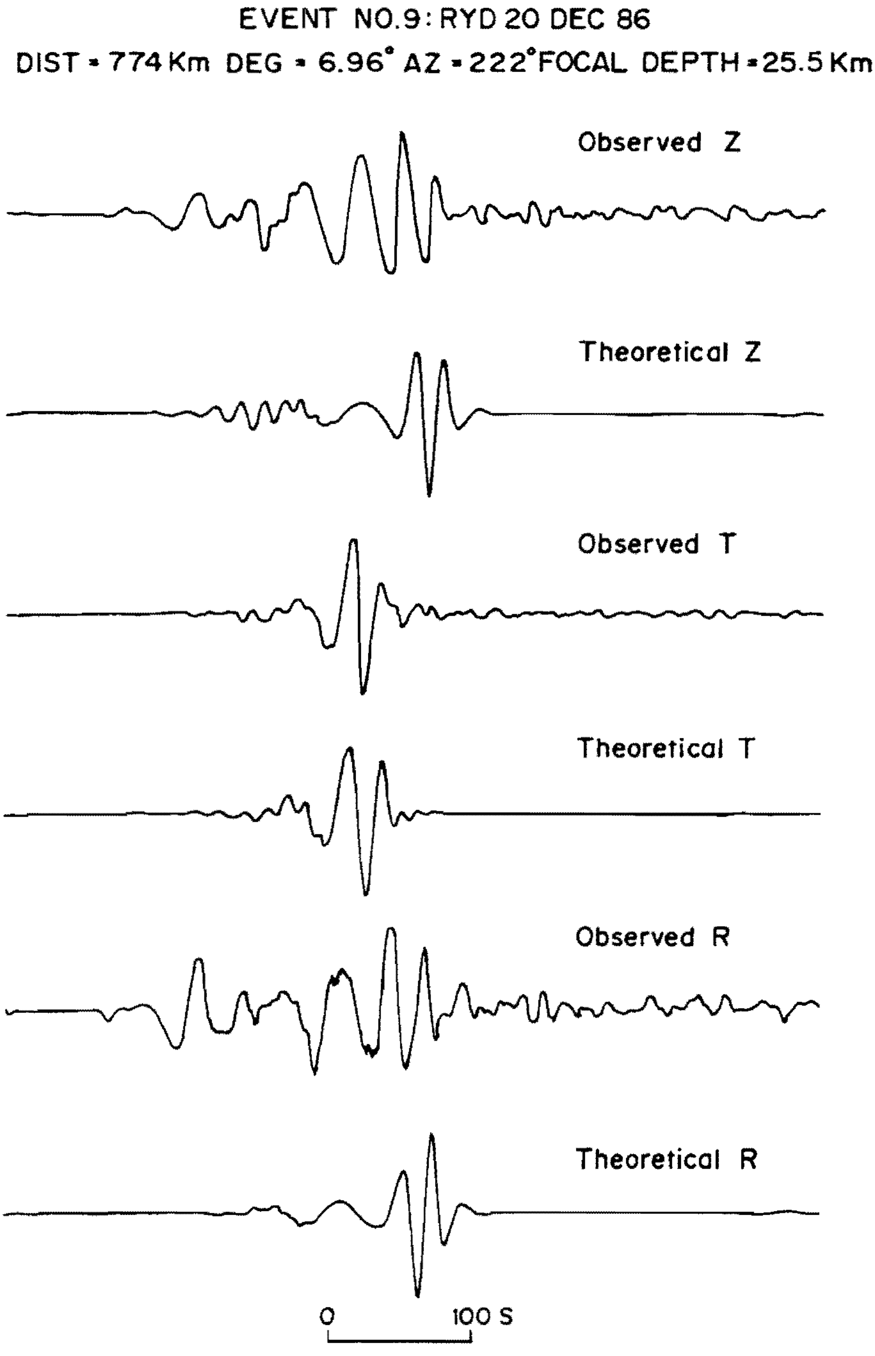

FIG. 9. Observed and synthetic seismograms of the long period components of RYD Station for the 20 December 1986 earthquake (event No. 9). The depth of focus used in computing the synthetic seismograms is 25.0 km.

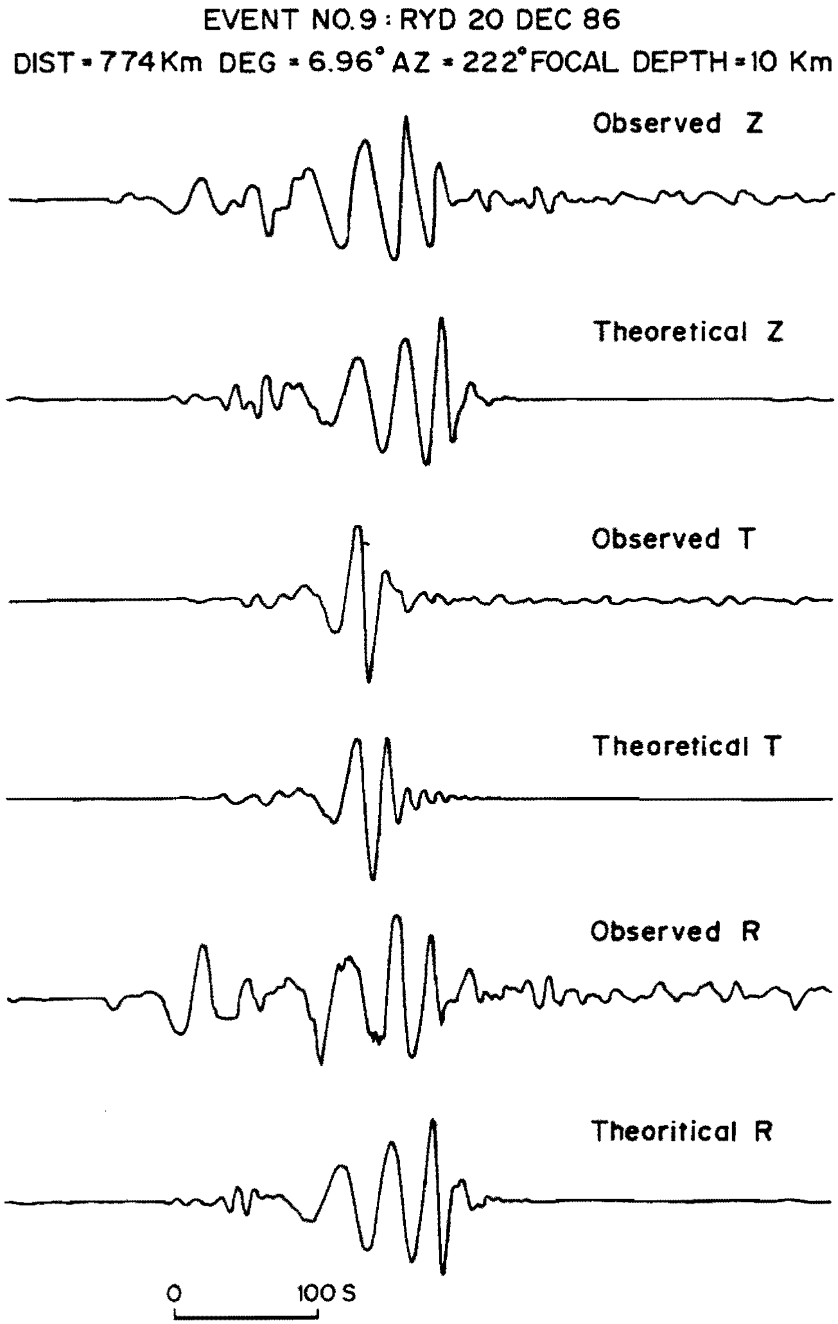

FIG. 10. Observed and synthetic seismograms of the long period components of RYD Station for the 20 December 1986 earthquake (event No. 9). The depth of focus used in computing the synthetic seismograms is 10.0 km.

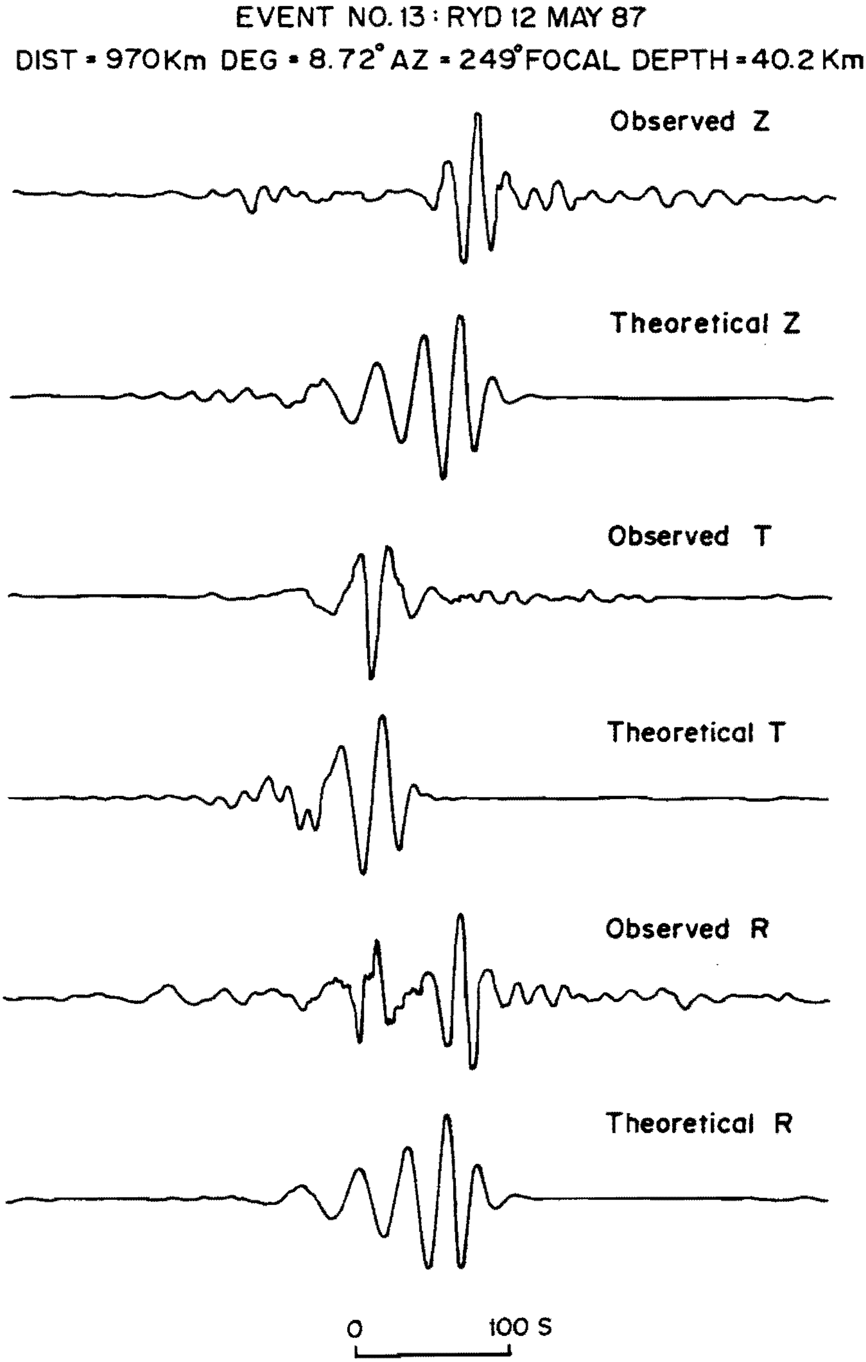

FIG. 11. Observed and synthetic seismograms of the long period components of RYD Station for the 12 May 1987 earthquake (event No. 13). The depth of focus used in computing the synthetic seismograms is 40.2 km.

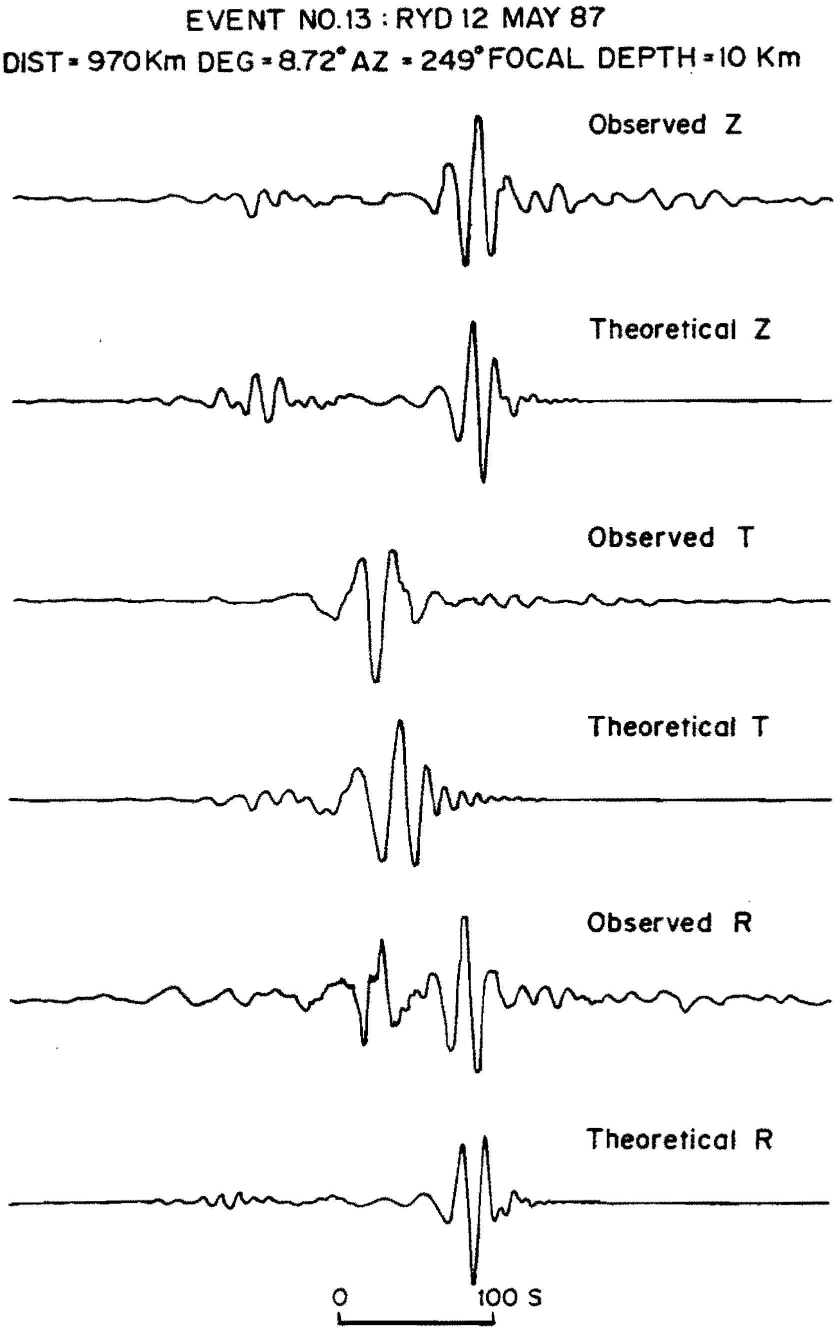

FIG. 12. Observed and synthetic seismograms of the long period components of RYD Station for the 12 May 1987 earthquake (event No. 13). The depth of focus used in computing the synthetic seismograms is 10.0 km.

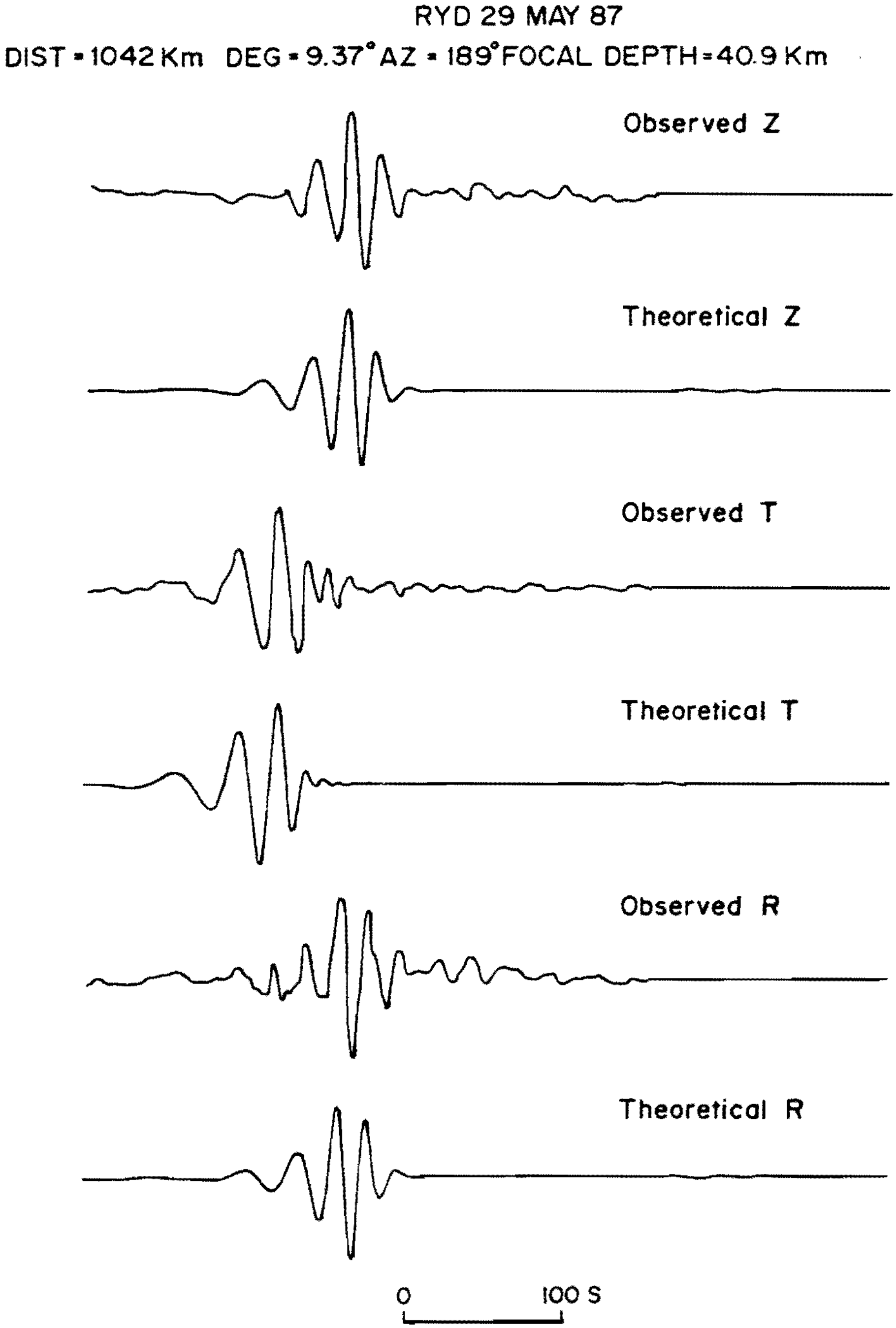

FIG. 13. Observed and synthetic seismograms of the long period components of RYD station for the 29 May 1987 earthquake.

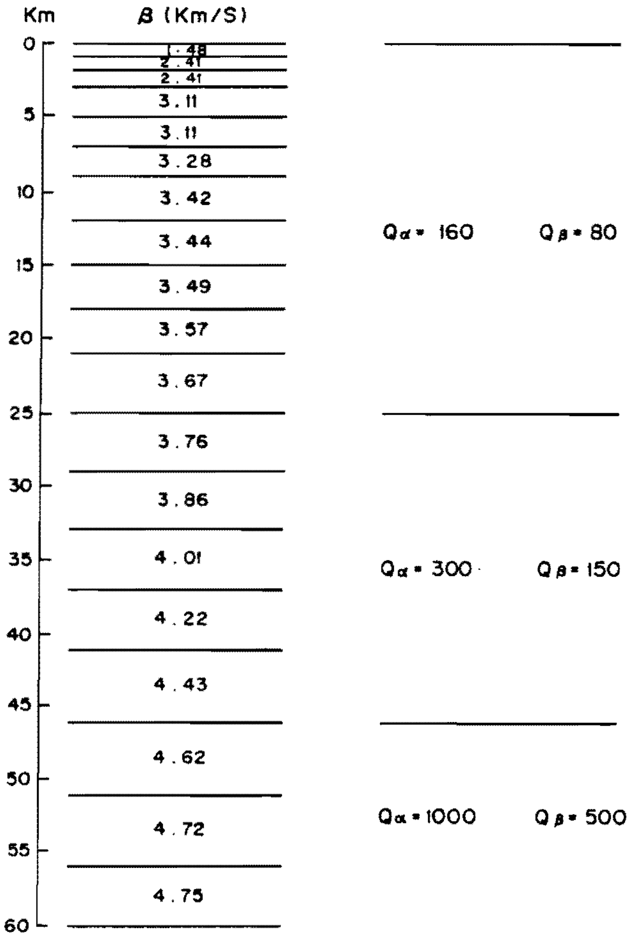

FIG. 14. Shear wave velocity and Q_β model of the Arabian platform used to produce synthetic seismograms.

Knowing the ground motion amplitude, one can modify the Q_β values of the upper crust used initially in order to obtain good matching of the synthetic amplitude with the observed ones in addition to waveform matching, which is mainly dependent on the seismic velocity model used.

The phase velocity dispersion of the Arabian platform is characteristic of that of continental crust. The phase velocities were found to increase rapidly from about 2.3 km/s at 4 seconds to about 2.9 km/s at 10 seconds. The low velocity at the short periods is due to the presence of the sedimentary cover section in the platform. At longer periods, the increase in phase velocity is more slower and the phase velocity reaches 4.1 km/s at 42 seconds. The shear velocity structure obtained by inversion of phase and group velocities of Rayleigh and Love wave is consistent with that of Mokhtar and Al-Saeed (1994) for this area.

The waveform modeling results indicate that the Q_β of the upper crust is about 75. This is consistent with previously published results by Seber and Mitchell (1992). Further investigations using appropriate seismic data are required to determine the quality factor of the lower crust in the region.

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سرعــة الــطــور للرصــيف العــربي وخــواص التــوهــين له باستخــدام نمذجــة شــكل الموجــة

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المستخلص . تمت دراسة التركيب القشري تحت الرصيف العربي باستخدام المعلومات الزلزالية المتاحة من محطة الرياض . وجدت سرعة الطور على طول ١٤ مسارًا باستخدام طريقة المصفى الموافق للطور . ووجد أن سرعات الطور تزيد من ٣,٢ كم/ث عند ٤ ثوان إلى ٤,١ كم/ث عنـد ٤٢ ثانية . جمعت السرعات للطور وللمجموعة واستخدمت في انعكاس لاستنباط نموذج زلزالي لموجة S . النموذج المتحصل عليه أدمج مع نموذج لـ Q يمثل المنطقة وحسبت الرسومات الزلزالية النظرية . وبمقارنة الرسومات الزلزالية المشاهدة والنظرية وجد أن نموذج سرعة القص المحسوب في هذه الدراسة قد أعطى توافقًا ممتازًا بين الرسومات المشاهدة وتلك المحسوبة . وهذه الطريقة يمكن استخدامها لدراسة خواص التوهين في الرصيف العربي وكذلك الطبق العربي ككل .