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Shear Velocity Structure in the Eastern United States
from the Inversion of Surface Wave
Group and Phase Velocities

by

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and

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upper mantle indicate that all of the data can be explained satisfactorily by a single model which is isotropic in its elastic properties. This model includes a surface layer of low-velocity sediments, and is characterized by upper crustal shear-wave velocities which increase from 3.4 to 3.7 km/sec with increasing depth, by lower crustal shear-wave velocities between 4.0 and 3.8 km/sec, and by an upper-most mantle shear-wave velocity of about 4.8 km/sec. The fundamental-mode data of this study do not require a low-velocity zone in either the crust or upper mantle. Higher-mode velocities and wave forms, although having potentially greater resolving power than the fundamental-mode data, were too inconsistent to contribute to the elucidation of detailed features of the crust-upper mantle model.

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Abstract

Single-station group velocities at periods between 4 and 70 seconds and single-station phase velocities at periods between 10 and 50 seconds are obtained for a broad region of the eastern United States for both Rayleigh and Love waves. These data were obtained by the single-station method using earthquakes with known fault-plane solutions in the central United States. The group velocity data are combined with earlier acquired attenuation data to yield Q values for Rayleigh and Love waves in the eastern United States. Inversions of the dispersion curves for shear-velocity models of the crust and upper mantle indicate that all of the data can be explained satisfactorily by a single model which is isotropic in its elastic properties. This model includes a surface layer of low-velocity sediments, and is characterized by upper crustal shear-wave velocities which increase from 3.4 to 3.7 km/sec with increasing depth, by lower crustal shear-wave velocities between 4.0 and 3.8 km/sec, and by an upper-most mantle shear-wave velocity of about 4.8 km/sec. The fundamental-mode data of this study do not require a low-velocity zone in either the crust or upper mantle. Higher-mode velocities and wave forms, although having potentially greater resolving power than the fundamental-mode data, were too inconsistent to contribute to the elucidation of detailed features of the crust-upper mantle model.

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Introduction

Surface wave group and phase velocities have provided important information on the properties of the crust and upper mantle in various regions of the Earth. Their usefulness is partly due to their sensitivity to shear wave structure; thus they can provide information which is often difficult to obtain from body wave studies. A second advantage is that no special assumptions or procedures are necessary to invert surface wave data for models which include low-velocity zones. Most surface wave studies of crust-upper mantle structure have employed either phase or group velocities. Studies by Bloch et al. (1969) and Yu and Mitchell (1978) have, however, emphasized the desirability of employing both phase and group velocities to invert for structure.

Surface waves are most useful for studying structure in regions that are devoid of large-scale lateral property changes. Results of seismic refraction and other geophysical studies suggest that the eastern and central portions of North America are such suitable regions. Crust-upper mantle models for those areas have previously been obtained by Brune and Dorman (1963) using Love and Rayleigh waves crossing the Canadian shield, by McEvelly (1964) also using Love and Rayleigh waves in the central United States, and by Mitchell (1965) who used only Rayleigh waves to study a broader region of the east-central United States. Biswas and Knopoff (1974) used long-period Rayleigh waves to obtain upper mantle models for the central United States, as well as for other regions of North America. In later sections, we will refer particularly to the models of McEvelly (1964) since they were obtained for a region which corresponds most closely with the

region of the present study. All of the above references employed only phase velocities. We have used group velocity, as well as phase velocity, data in an attempt to obtain a new, more detailed, crust-upper mantle model for the east-central United States.

An interesting feature of the study by McEvelly (1964), was that a single isotropic model for the crust and upper mantle could not be found which would explain both the Love wave and Rayleigh wave data. He gave three possible explanations for that result: (1) that the upper mantle or/and lower crust are anisotropic in their elastic properties, (2) that systematic errors occur in the fundamental-mode phase velocities due to interference by higher modes, and (3) that his model, which included three crustal layers and four upper mantle layers was too simplified to completely explain his data. Thatcher and Brune (1969) suggested that the second of these explanations was likely to be the correct one. Boore (1969), however, argued that higher-mode interference would produce a large scatter, but no uniform bias, if data were obtained from an ensemble of sources.

Two developments since the study of McEvelly (1964) have led us to attempt to obtain a new shear-velocity model for the central and eastern United States. First, a new data set including both group and phase velocities has been obtained using central United States earthquakes with known fault-plane solutions. Second, inversion techniques are now available which permit us to consider models which are more complicated than those of McEvelly (1964).

We will invert the fundamental-mode Love- and Rayleigh-wave data, both simultaneously and separately, to investigate the possibility of anisotropy in crust-upper mantle properties. This approach of using isotropic inversion methods to study anisotropy is not strictly valid (Crampin, 1970, 1976). It will, however, provide an indication of the severity of any anisotropy

that might be present. In addition, synthetic seismograms, incorporating the fundamental mode and several higher modes will also be constructed in an attempt to investigate features of the resulting models which cannot be resolved using only the fundamental mode.

Sources

A fault-plane solution for only one central United States earthquake, that of November 9, 1968, has been determined using teleseismic P-waves (Stauder and Nuttli, 1970). Although numerous other earthquakes have occurred there, especially in the New Madrid seismic zone, they have been too small to be amenable to study by the classical method of teleseismic P-wave first motions. Mitchell (1973a,b) however, showed that surface waves are a valuable tool for determining depths and fault-plane solutions for relatively small earthquakes in that region in studies of the southeastern Missouri earthquake of October 21, 1965. He adopted a method similar to that devised by Tsai and Aki (1969). The use of surface waves has made it possible to obtain depths and fault-plane solutions for several central and eastern United States earthquakes (Herrmann, 1974, 1978). A number of these events will be used as sources of the surface data in the present study.

Data

Our data were obtained for paths between earthquake sources in the central United States (Table 1 and Figure 1) and numerous stations in the central and eastern United States. Phase velocities were determined using the single-station method (Brune et al., 1960). By using earthquakes with known fault-plane solutions, the phase and group velocities can be corrected for the initial phase and group delay at the source (Knopoff and Schwab, 1968).

The group velocities of this study were obtained using the multiple-filter method (Dziewonski et al., 1969; Herrmann, 1973). Spectral amplitudes associated with the fundamental mode can be separated from higher modes and noise; thus a fundamental-mode group velocity value can be obtained which is not contaminated by higher-mode interference.

The group velocities of this study were obtained as by-products of source mechanism and surface wave attenuation determinations (Herrmann, 1974). Those determinations included data from numerous paths to seismograph stations in both eastern and western North America. That body of data, after correction for the group delay at the source, is plotted in Figure 2. The large scatter in the data can be attributed to several factors, including regional variations in surface wave velocities, the inclusion of data with low signal-noise ratios, possible mode misidentification, and incorrect adjustment for initial phase due to departures of the actual focal mechanism from the assumed fault-plane solution.

In order to obtain a reliable data set pertaining to the eastern United States, the data of Figure 2 have been windowed in several ways. An azimuth window (340° - 135°) and a distance window (300 - 2000 km) were

first applied to the data. The azimuth window restricts our region of study to a section of the United States centered south of the Great Lakes, and omits data for paths in the western and southern United States where significant changes in shear wave structure may occur (Biswas and Knopoff, 1974). The distance window removes data obtained at short distances where possible incorrect initial phase assumptions cause the largest errors in phase and group velocity determinations, and at large distances where effects of lateral structural changes, such as surface wave lateral refraction and multi-pathing are most severe. The data have also been windowed according to their amplitudes. For each event, data having spectral amplitudes smaller than one-fifth the maximum amplitude at the given period were excluded. This process removes data which are associated with nodes in the surface wave source radiation patterns, and which consequently have a low signal-noise ratio. In addition, a few group velocity points which are clearly removed from the main trend of the data, and probably correspond to noise or higher-mode maxima rather than to the fundamental mode, were deleted from the final data set. The windowed data appear in Figure 3. It is clear that a considerable reduction of scatter has resulted from the windowing process.

The great preponderance of surface wave paths are restricted to a region which extends northward from the source region to the Great Lakes and eastward from the source region to the Appalachian Mountains (see Figure 1). A few short segments of path extend into Canada, to the east coast, and to the southeastern United States. Therefore our data pertain largely to the northeastern and north central United States. It is worthwhile to inquire about the extent of lateral variations in elastic properties through these regions since significant lateral variations

could bias our results. We have addressed this question by considering pertinent results of earlier surface wave studies, seismic refraction studies, and investigations of Pn travel-times.

The most useful surface wave results bearing on this question are those of Pilant (1972). Those data appear as contour maps of Rayleigh wave phase velocities across the United States. Across the region of the present study Pilant's velocities vary between 3.40 and 3.55 km/sec at a period of 16 seconds, between 3.55 and 3.70 km/sec at a period of 20 seconds, between 3.70 and 3.80 km/sec at a period of 25 seconds, between 3.90 and 4.00 km/sec at a period of 33 seconds, between 3.95 and 4.05 km/sec at a period of 41 seconds, between 4.00 and 4.10 km/sec at a period of 51 seconds, and between 4.05 and 4.10 km/sec at a period of 64 seconds. Those variations are comparable to the uncertainty of our observations. In general, Pilant's maps suggest that although some lateral variations occur across the region of interest, they are relatively minor.

Seismic refraction studies in the central United States (e.g. Stewart, 1968, and Ocala and Meyer, 1973) as well as more easterly portions of the central plains (Steinhart and Meyer, 1961) indicate crustal thicknesses between 40 and 45 km. In the New York-Pennsylvania region, however, a crustal thickness of 36 km was obtained (Katz, 1955). Along the eastern seaboard, and outside of our region of interest, the crust apparently thins considerably. Pn velocities, as compiled by Herrin and Taggart (1965), are relatively uniform throughout our region of study, varying only between 8.1 and 8.2 km/sec.

Phase velocities were determined using the better recorded events of Table 1. Spectral phases were obtained, and corrected for the initial phase at the source. Phase velocities were calculated using the single-station method (Brune et al., 1960). In using that method, a family of phase velocity

curves are determined, each corresponding to a different integer N , which occurs because there is an uncertainty of $2N\pi$ in the observed phases.

The value of the integer has been obtained by comparing the resulting phase velocities for various N values to the phase velocities of McEvelly (1964) and Biswas and Knopoff (1974) at longer periods.

Thatcher and Brune (1969) have pointed out that the velocities of long-period, fundamental-mode Love waves can be in error if there is significant interference with higher-mode Love waves. Difficulties due to interference with higher modes are apt to be most severe if the distance between the source and receiver includes a segment of oceanic path. Although there are no portions of oceanic path in the present study, it is still worthwhile to investigate the level of excitation of higher modes relative to that of the fundamental mode. Spectral amplitudes observed at a recording station will depend upon the depth and focal mechanism of the earthquake source, as well as upon the elastic and anelastic properties of the crust and upper mantle between the source and receiver. Spectral amplitudes for the fundamental Love mode and nine higher modes were computed using the formulation of Levshin and Yanson (1971) for the case of the earthquake of October 21, 1965 (Figure 4). The crust-upper mantle model used in the calculations is that obtained from Rayleigh waves as described in the following section. Spectra were determined for two source depths: 4 km, as obtained for this earthquake by Mitchell (1973a), and 15 km. The latter depth was chosen in order to illustrate the effect which variations in source depth have on the spectra. In both cases, at periods greater than about 10 seconds, the higher-mode amplitudes are down by more than an order of magnitude below those of the fundamental mode. We expect therefore, that our Love

wave data will not be significantly biased by the presence of higher modes.

Uncertainties in surface wave group and phase velocities can be determined from travel-time uncertainties associated with digitizing errors, inexact estimates of the initial phase at the source, uncertainties in origin time and source location, the effects of source finiteness, and the limited precision with which we can read group and phase travel times. We estimate the maximum uncertainties in the group and phase travel times to be about 8 seconds. At a distance of 1000 km, this time uncertainty leads to velocity uncertainties between 0.1 and 0.15 km/sec. The data points of Figure 3 exhibit a spread of about that much. The data used for inversion and their standard deviations are plotted in Figure 5. The standard deviations were calculated for all periods at which there were three or more data points. In cases where two data points were available, a standard deviation of 0.1 was used, and in cases where only one data point was obtained, a standard deviation of 0.15 was used.

Table 2 presents the mean group and phase velocities obtained in this study. Combining the group velocity data with values for surface wave attenuation coefficients (Mitchell, 1973a; Herrman and Mitchell, 1975) for the eastern United States allows us to tabulate Q values for Rayleigh and Love waves in that region. These values are also listed in Table 2.

Our observed group and phase velocity data are compared with theoretical values for three models of McEvelly (1964) in Figure 5. That set of models was derived from phase velocity data only and includes (1) a model which is isotropic at all depths, (2) a model which is anisotropic only in the upper mantle, and (3) a model which is anisotropic in both the upper mantle and lower crust. The last of

these models provided the best fit to McEvilly's data. The comparisons in Figure 5 indicate that the phase velocities for McEvilly's model which is anisotropic in both the lower crust and upper mantle fit the present set of phase velocity data quite well. This result indicates that the phase velocity data of McEvilly (1964) were not severely biased due to higher-mode interference.

The Rayleigh wave group velocities predicted by the same model also fit our observed data well, except at the shortest periods. The lower observed group velocities at those periods suggest that a low-velocity sedimentary layer is required for any eastern United States model. None of the Love wave group velocity curves predicted by McEvilly's models correspond very well to our observed group velocities. Fitting these data, as well as the phase velocities and Rayleigh wave group velocities, will apparently require a more complicated model for the eastern United States than any which is presently available.

Models Obtained From Inversion

Shear wave models were obtained from the surface wave data using modern inversion theory (Bachus and Gilbert, 1970) in stochastic form (Der et al., 1970; Jordan and Franklin, 1971). The program we have used for the inversions was written by W. L. Rodi and employs the surface wave algorithm of Harkrider (1964) and efficient methods for computing partial derivatives (Rodi et al., 1975). We have assumed that the compressional velocities and densities are known and have solved only for the shear velocities. Several compressional wave crustal models for the eastern United States are available. Two of these (McCamy and Meyer, 1966; Mitchell and Hashim, 1977) pertain to the region near the sources used in this study. These models are quite similar to one another and both include three crustal layers, in addition to surface sediments. We have taken the model of Mitchell and Hashim (1977) as the crustal compressional wave model for this region and those velocities have been fixed in all of our inversion attempts. Our starting shear wave model for the crust was determined from the compressional wave model by assuming a Poisson ratio value of 0.25. The starting upper mantle model is based upon the model of McEvelly (1964), but has been modified to better explain our group velocity data. The starting shear velocity model appears as a dotted line in each of the models of Figure 6.

Figure 6 presents shear velocity models and their standard deviations resulting from the inversion of the combined Rayleigh- and Love-wave data, from the Rayleigh-wave data only, and from the Love-wave data only. The combined inversion yielded a shear wave value of 1.53 km/sec for the upper 0.25 km of the model and an average value of 3.06 km/sec for the next 0.75 km. Large uncertainties are associated with these values. Resolving kernels for these models are given in Figure 7 and comparisons between the observed and theoretical velocities for the models appear in Figure 8. During the inversion process the

theoretical phase and group velocities were corrected for sphericity according to methods described by Schwab and Knopoff (1972).

As indicated by Figure 6, there are no significant differences between the three models, except perhaps in the lower crust where the shear velocities obtained from the inversion of Love waves (SH) are faster than the shear wave velocities obtained from the inversion of Rayleigh waves (SV) or of the combined data. With the exception of only one value, the upper mantle velocities are always within one standard deviation of the starting model. It is apparent, when the standard deviations and resolving kernels are considered, that an upper mantle low-velocity zone, if present, cannot be resolved by the data. This result was also noted by Biswas and Knopoff (1974). The small velocity decrease at a depth of 65 km occurs because a velocity decrease was present in the starting model at that depth and is not resolvable in the final model.

The relative sharpness of the resolving kernels for crustal depths suggests that we should be able to resolve much more detail in the crust than in the mantle. There are no significant differences among the three models of the crust at depths shallower than about 30 km. At greater depths, however, the shear velocities obtained using Rayleigh waves are lower than those obtained using Love waves. They also exhibit a decrease in value with increasing depth. These results perhaps suggest the possibility of both a low-velocity zone and polarization anisotropy near the base of the crust. Before reaching these conclusions, however, it is necessary to consider the resolving kernels and standard deviations corresponding to that depth range. Figure 7 indicates that the resolving kernels for those depths have a width of about 25 km or slightly more.

Since the thickness of the possible low-velocity layer is less than 20 km, this feature, if present, cannot be resolved using the present fundamental-mode dispersion data.

The theoretical velocities for the three models appear in Figure 8. Almost all of the data agree with the theoretical curves within one standard deviation. The only clear exception is one Love wave phase velocity value at a period of 35 seconds which differs from the theoretical velocity for the model obtained from the combined inversion by a small amount. As might be expected, the theoretical velocity values for the models obtained from separate inversions of the Love- and Rayleigh-wave data correspond more closely to the observed values than do the theoretical values for the model from the combined inversion. Although this better fit could occur because of anisotropy in the crust or upper mantle in the eastern United States, we do not consider that the improvement in fit is substantial enough to warrant that conclusion at the present time.

Synthetic Seismograms

The results of inversion of the fundamental-mode dispersion data indicate that there are no resolvable differences among the three models of Figure 6. It might be possible, however, to use higher-mode information to investigate features of these models in more detail. For example, the shear velocity at the top of the mantle (S_n) is higher for the model obtained using both Rayleigh and Love wave (R+L) data than it is for the model obtained using only Rayleigh wave (R) data. The R+L model compensates for the higher S_n velocities by having lower shear wave velocities in the lower crust than those which occur in the R model.

A further question which we might address using the synthetic seismograms concerns the existence of the low-velocity zone in the lower crust, suggested by two of the models in Figure 6. As stated earlier, we cannot resolve this feature using fundamental-mode dispersion data alone. Jordan and Frazer (1975) have inferred such a low-velocity zone in the lower crust of eastern Canada from the amplitudes of S to P converted waves at the base of the crust.

Higher mode data could contribute detailed information on the shear velocity of the crust if it is of sufficient quality. The higher mode group velocity data which we obtained, however, exhibits considerable scatter, and is not of sufficient quality to use for inversions for structures. We have attempted to study possible detailed features of our models by computing synthetic seismograms which include the effects of the fundamental and several higher modes (Tsai and Aki, 1970), and comparing these to the observed seismograms at several stations. The synthetics incorporate anelastic attenuation using values from Herrmann and Mitchell (1975).

Figure 9 shows the observed Rayleigh wave seismograms (vertical component), and the Love wave seismograms (transverse component) obtained from a coordinate transformation of the observed east-west and north-south components, for the October 21, 1965 earthquake, as recorded at four stations. These stations, MDS, AAM, BLA, and ATL, were selected because they are located at azimuths which avoid nodes in the surface wave radiation patterns; thus they provide well-recorded surface waves with a good signal-noise ratio.

The synthetic seismograms for the R+L, the R, and the L models appear in Figure 10. It is apparent that differences between the Rayleigh wave seismograms and between the Love wave seismograms for the different models are not very great. It is also apparent that although the major features of the synthetics agree with those of the observed seismograms, detailed comparisons are not possible. We therefore have no basis for selecting one model over the other, either in comparing model R to R+L, or in comparing model L to R+L. Moreover, there is no reason to propose more than a single isotropic model to satisfy the presently available surface wave data for the eastern United States.

Synthetic seismograms were also computed for models with and without a low-velocity zone in the lower crust. Differences between the synthetic seismograms for the two kinds of models at the various stations were even less substantial than those of Figure 10. It is not possible therefore, to come to any conclusions concerning the existence or absence of a low-velocity zone in the lower crust of the eastern United States from the presently available surface wave data.

Conclusions

An isotropic model for the crust and upper mantle of the eastern United States can be obtained which satisfies both fundamental-mode Love and Rayleigh wave phase and group velocity data within one standard deviation. Shear velocities in the upper crust of this model increase from 3.4 km/sec to 3.7 km/sec with increasing depth. Shear velocities in the lower crust lie between 4.0 and 3.8 km/sec. Although the model resulting from the inversion process exhibits decreasing velocity with increasing depth in the lower crust, this resulting low-velocity zone cannot be resolved using the data available to us. The uppermost mantle for this model has a shear wave velocity of about 4.8 km/sec. Although a modest low-velocity zone may be present in the upper mantle, one is not required.

Models resulting from separate inversions of the Rayleigh and Love wave data provide slightly better agreement with the observed dispersion data. This result suggests that the lower crust and/or upper mantle of this region could be anisotropic in its elastic properties. The observations are not definitive enough, however, to establish the existence of anisotropy. For the present, therefore, a single isotropic model for the crust and upper mantle of the eastern United States is sufficient to explain all of the surface wave data for that region.

The short-period group velocity data for the eastern United States require the existence of a layer of low-velocity sediments overlying the crust.

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References

- Bachus, G., and F. Gilbert (1970). Uniqueness in the inversion of gross earth data, Phil. Trans. Roy. Soc. (London), ser. A, 266, 123-192.
- Biswas, N.N., and L. Knopoff (1974). The structure of the upper mantle under the United States from the dispersion of Rayleigh waves, Geophys. J. Roy. ast. Soc., 36, 515-539.
- Bloch, S., A.L. Hales, and M. Landisman (1969). Velocities in the crust and upper mantle of southern Africa from multi-mode surface wave dispersion, Bull. Seism. Soc. Am., 59, 1599-1629.
- Boore, D.M. (1969). Effect of higher mode contamination on measured Love wave phase velocities, J. Geophys. Res., 74, 6612-6616.
- Brune, J. and J. Dorman (1963). Seismic waves and earth structure in the Canadian shield, Bull. Seism. Soc. Am., 53, 167-210.
- Brune, J., J. Nafe, and J. Oliver (1960). A simplified method for the analysis and synthesis of dispersed wave trains, J. Geophys. Res., 65, 287-304.
- Crampin, S. (1970). The dispersion of surface waves in multilayered anisotropic media, Geophys. J. Roy. ast. Soc., 21, 387-402.
- Crampin, S. (1976). A comment on 'The early structural evolution and anisotropy of the oceanic upper mantle,' Geophys. J. Roy. ast. Soc., 46, 193-197.
- Der, Z., R. Masse, and M. Landisman (1970). Effects of observational errors on the resolution of surface waves at intermediate distances, J. Geophys. Res., 75, 3399-3409.
- Dziewonski, A.M., S. Bloch, and M. Landisman (1969). A technique for the analysis of transient seismic signals, Bull. Seism. Soc. Am., 59, 427-444.

- Franklin, J.N. (1970). Well-posed stochastic extensions of ill-posed linear problems, J. Math. Anal. Appl., 31, 682-716.
- Harkrider, D.G. (1964). Surface waves in multi-layered media, 1. Rayleigh and Love waves from buried sources in a multi-layered elastic half-space, Bull. Seism. Soc. Am., 54, 627-629.
- Herrin, E.T., and J. Taggart (1966). Regional variations of Pn velocity and their effect on the location of epicenters, Bull. Seism. Soc. Am., 52, 1037-1046.
- Herrmann, R.B. (1973). Some aspects of band-pass filtering of surface waves, Bull. Seism. Soc. Am., 63, 663-671.
- Herrmann, R.B. (1974). Surface wave generation by central United States earthquakes, Ph.D. Dissertation, Saint Louis University, 262 pp.
- Herrmann, R.B. (1978). A seismological study of two Attica, New York earthquakes, Bull. Seism. Soc. Am., 68, 641-651.
- Herrmann, R.B., and B.J. Mitchell (1975). Statistical analysis and interpretation of surface wave anelastic attenuation data for the stable interior of North America, Bull. Seism. Soc. Am., 65, 1115-1128.
- Jordan, T.H., and J.N. Franklin (1971). Optimal solutions to a linear inverse problem in geophysics, Proc. Nat. Acad. Sci., 68, 291-293.
- Jordan, T.H., and L.N. Frazer (1975). Crustal and upper mantle structure from Sp phases, J Geophys. Res., 80, 1504-1518.
- Katz, S. (1955). Seismic study of crustal structure in Pennsylvania and New York, Bull. Seism. Soc. Am., 45, 303-325.
- Knopoff, L., and F.A. Schwab (1968). Apparent initial phase of a source of Rayleigh waves, J. Geophys. Res., 73, 755-760.
- McCamy, K., and R.P. Meyer (1966). Crustal results of fixed multiple shots in the Mississippi embayment, in The Earth Beneath the Continents Mono. 10., Am. Geophys. Union, 370-381.

- McEvelly, T.V. (1964). Central U.S. crust-upper mantle structure from Love and Rayleigh wave phase velocity inversion, Bull. Seism. Soc. Am., 54, 1997-2015.
- Mitchell, B.J. (1965). Surface wave dispersion and crustal structure across the central United States, M.S. thesis, Univ. of Minnesota, 81 pp.
- Mitchell, B.J. (1973a). Radiation and attenuation of Rayleigh waves from the southeastern Missouri earthquake of October 21, 1965, J. Geophys. Res., 78, 886-800.
- Mitchell, B.J. (1973b). Surface wave attenuation and crustal anelasticity in central North America, Bull. Seism. Soc. Am., 63, 1057-1071.
- Mitchell, B.J. and B.M. Hashim (1977). Seismic velocity determinations in the New Madrid seismic zone: A new method using local earthquakes, Bull. Seism. Soc. Am., 67, 413-424.
- Ocala, L.C., and R.P. Meyer, Central North American rift system, 1, Structure of the axial zone from seismic and gravimetric data, J. Geophys. Res., 5173-5194.
- Rodi, W.L., P. Glover, T.M.C. Li, and S.S. Alexander (1975). A fast accurate method for computing group-velocity partial derivatives for Rayleigh and Love modes, Bull. Seism. Soc. Am., 65, 1105-1114.
- Schwab, F.A. and L. Knopoff (1972). Fast surface wave and free mode computations, in Methods of Computational Physics, Vol. 11, Seismology: Surface Waves and Earth Oscillations, ed. by B.A. Bolt, 87-180, Academic Press, New York.
- Steinhart, J.S., and R.P. Meyer (1961). Explosion studies of continental structure, Publ. 622, Carnegie Inst. of Wash., Washington, D.C., 409 pp.

- Stauder, W., and O.W. Nuttli (1970). Seismic studies: South central Illinois earthquake of November 9, 1968, Bull. Seism. Soc. Am., 60, 973-981.
- Stewart, S.W. (1968). Crustal structure in Missouri by seismic refraction methods, Bull. Seism. Soc. Am., 58, 291-324.
- Thatcher, W., and J.N. Brune (1969). Higher mode interference and observed anomalous apparent Love wave phase velocities, J. Geophys. Res., 74, 6603-6611.
- Tsai, Y.B., and K. Aki (1969). Simultaneous determination of seismic moment and attenuation of surface waves, Bull. Seism. Soc. Am., 59, 275-287.
- Tsai, Y.B., and K. Aki (1970). Precise focal depth determination from amplitude spectra of surface waves, J. Geophys. Res., 75, 5729-5743.
- Yu, G.K., and B.J. Mitchell (1978). Regionalized shear velocity models of the Pacific upper mantle from observed Love and Rayleigh wave dispersion, Geophys. J. Roy. ast. Soc., submitted.

Table 1. Earthquake Parameters (from Herrmann, 1974)

Date	Origin Time	Lat.	Long.	Depth(km)	mb	T		P	
						Trend	Plunge	Trend	Plunge
2 Feb. 62	06 43 34.0	36.5	89.6	8	4.3	278	54	52	26
3 Mar 63	17 30 11.4	36.7	90.1	13	4.8	76	31	174	10
14 Aug. 65	13 13 56.6	37.2	89.3	1	3.8	147	1	237	17
*21 Oct. 65	02 04 38.3	37.5	91.0	4	4.9	163	5	305	84
4 June 67	16 14 13.6	33.6	90.9	10	4.5	156	22	248	7
21 July 67	09 14 48.9	37.5	90.4	15	4.3	50	5	314	52
9 Nov. 68	17 01 42.0	38.0	88.5	20	5.5	192	82	97	1
1 Jan. 69	23 35 36.2	34.8	92.6	9	4.4	228	65	330	7
17 Nov. 70	02 13 55.1	35.9N	89.9W	14	4.4	173	32	270	10

*Parameters from this earthquake were taken from Mitchell (1973a).

Table 2

Mean Group Velocities, Phase Velocities, Attenuation Coefficients, and Q values for the Eastern United States

T(sec)	U_R (km/sec)	C_R (km/sec)	$*Y_R$ (km ⁻¹)	Q_R	U_L (km/sec)	C_L (km/sec)	$*Y_L$ (km ⁻¹)	Q_L
4	2.894 + .201				3.258 + .087			
5	2.950 .102		5.10 x 10 ⁻⁴	418	3.348 .087		7.20 x 10 ⁻⁴	261
6	2.996 .073		5.50	318	3.399 .077			
7	3.036 .070		5.87	252	3.444 .084			
8	3.075 .064		4.06	315	3.434 .071		3.35	341
9	3.087 .057		2.45	462	3.440 .070		2.07	490
10	3.104 .059	3.301 + .045	3.12	324	3.443 .070	3.629	0.44	2074
12	3.099 .058	3.373 .034	2.70	313	3.457 .062	3.712 + .039	-0.05	
14	3.092 .061	3.440 .030	1.37	530	3.492 .035		0.07	9180
16	3.088 .067	3.490 .039	1.76	361	3.453 .055	3.807 .012	0.80	711
18	3.097 .068	3.529 .046	1.01	558	3.476 .058	3.914 .035	0.67	749
20	3.135 .076	3.582 .052	1.19	421	3.474 .049	3.990 .018	1.01	448
22	3.140 .077	3.610 .060	1.47	309	3.491 .045	3.984 .075	1.02	401
24	3.177 .069	3.657 .061	1.50	275	3.512 .041	4.068 .048	0.84	444
26	3.225 .094	3.725 .072	1.06	354	3.534 .049	4.137 .083	0.59	580
28	3.251 .084	3.786 .070	1.27	272	3.534 .049	4.163 .065	0.60	529
30	3.362 .098	3.811 .079	1.31	238	3.549 .050	4.141 .065	0.93	317
35	3.515 .127	3.916 .113	0.75	341	3.584 .067	4.350 .071	0.46	544
40	3.580 .143	4.021	1.25	176	3.653 .067	4.421 .084	1.13	190
45	3.745 .136	3.913			3.748 .073	4.434	0.02	9313
50	3.856 .072	4.071			3.865 .065	4.562		
60	3.870				3.960 .132			
70	3.864							

*From Herrmann and Mitchell (1975)

Standard deviations for U and C are not given for cases with two or fewer observations.

Figure Captions

Figure 1. Map indicating the location of earthquake sources and seismograph stations in eastern North America from which data used for surface wave inversion were obtained.

Figure 2. All group velocity data obtained using the central United States earthquakes of Table 1. These data were obtained from numerous stations throughout eastern and western North America without regard to regionality or signal-noise ratio.

Figure 3. Group velocity data which pertains only to eastern North America between the sources and seismograph stations shown in Figure 1. These data remain after windowing to exclude paths outside the distance range 300-2000 km, and outside the azimuth range 340° - 135° . Paths along azimuths from nodes in the surface wave radiation patterns have also been excluded.

Figure 4. Theoretical amplitude spectra for the fundamental-mode (F) and nine higher-mode Love waves generated by the earthquake of October 21, 1965 at a depth of 4 km (left), and by an earthquake with the same fault-plane solution, but at a depth of 15 km (right).

Figure 5. Phase and group velocity data and the standard deviations which were used for inversion plotted with theoretical values for models of McEvelly (1965). The solid lines, the short dashes, and the long dashes pertain to models which are isotropic, anisotropic, anisotropic in the upper mantle, and anisotropic in the lower crust and upper mantle, respectively.

Figure 6. Shear velocity models with standard deviations which have resulted from the inversion of both Rayleigh and Love waves (left), Rayleigh waves only (center), and Love waves only (right). The dotted lines denote the starting model used for the inversion process.

Figure 7. Resolving kernels for the models of Figure 6.

Figure 8. Phase and group velocity data used for inversion and theoretical values pertaining to the models of Figure 6. The solid lines correspond to the inversion of the combined Love- and Rayleigh-wave data and the dashed lines correspond to separate inversions of the Love- and Rayleigh-wave data.

Figure 9. Seismograms recorded at stations MDS, AAM, BLA, and ATL for the earthquake of October 21, 1965. The seismograms for transverse motion were obtained from a coordinate transformation of the east-west and north-south components. The zero times correspond to the beginning of digitization.

Figure 10. Synthetic seismograms corresponding to the event and stations of Figure 9 for models obtained from a simultaneous inversion of Rayleigh and Love wave data (left), from an inversion of Rayleigh data only (upper right), and from an inversion of Love wave data only (lower right). The starting times for each three-minute segment do not necessarily correspond to the starting times in Figure 9. The two second noise results because a boxcar filter with a sharp cutoff at 0.5 hz was used in the generation of the seismograms.

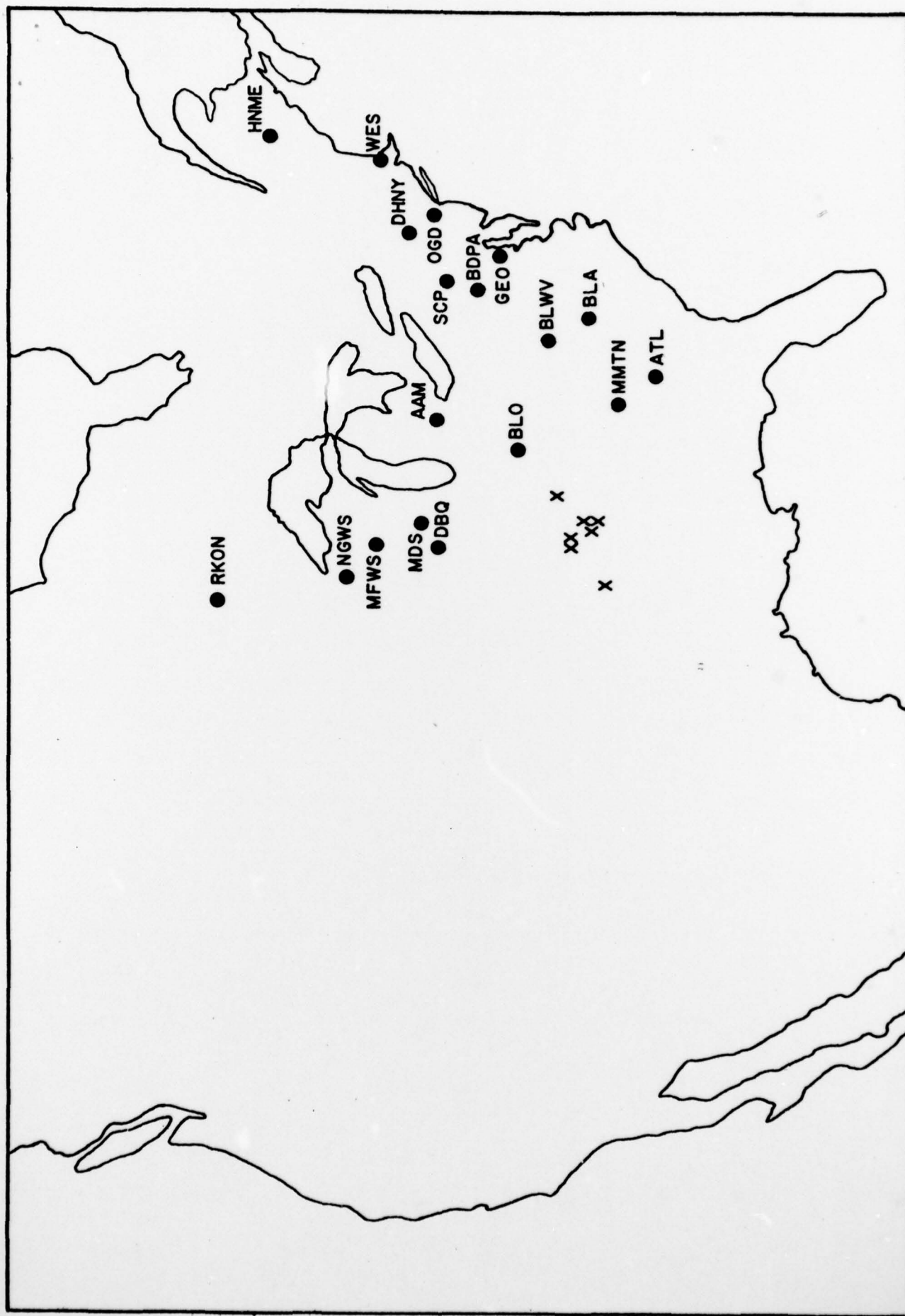


Figure 1

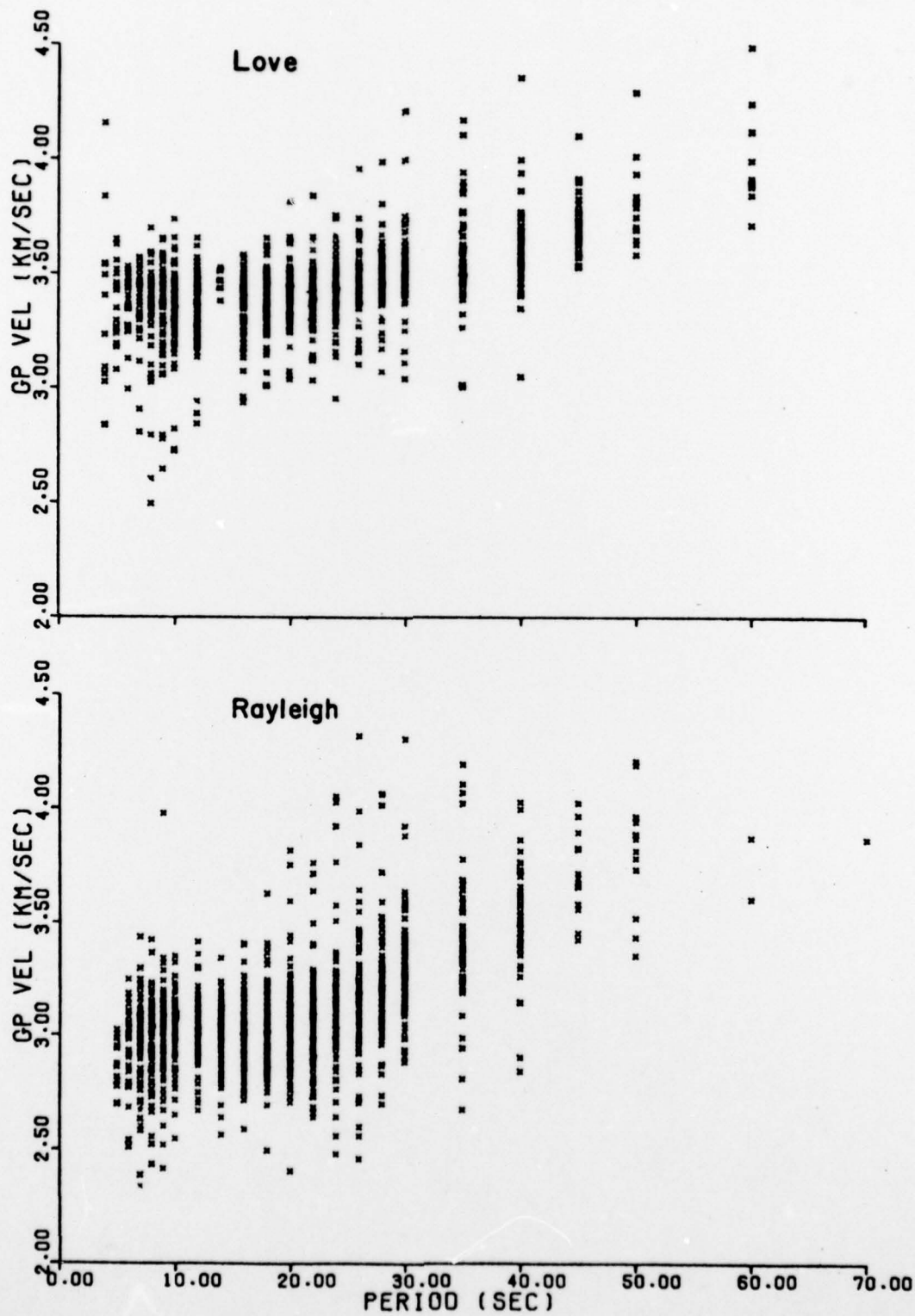
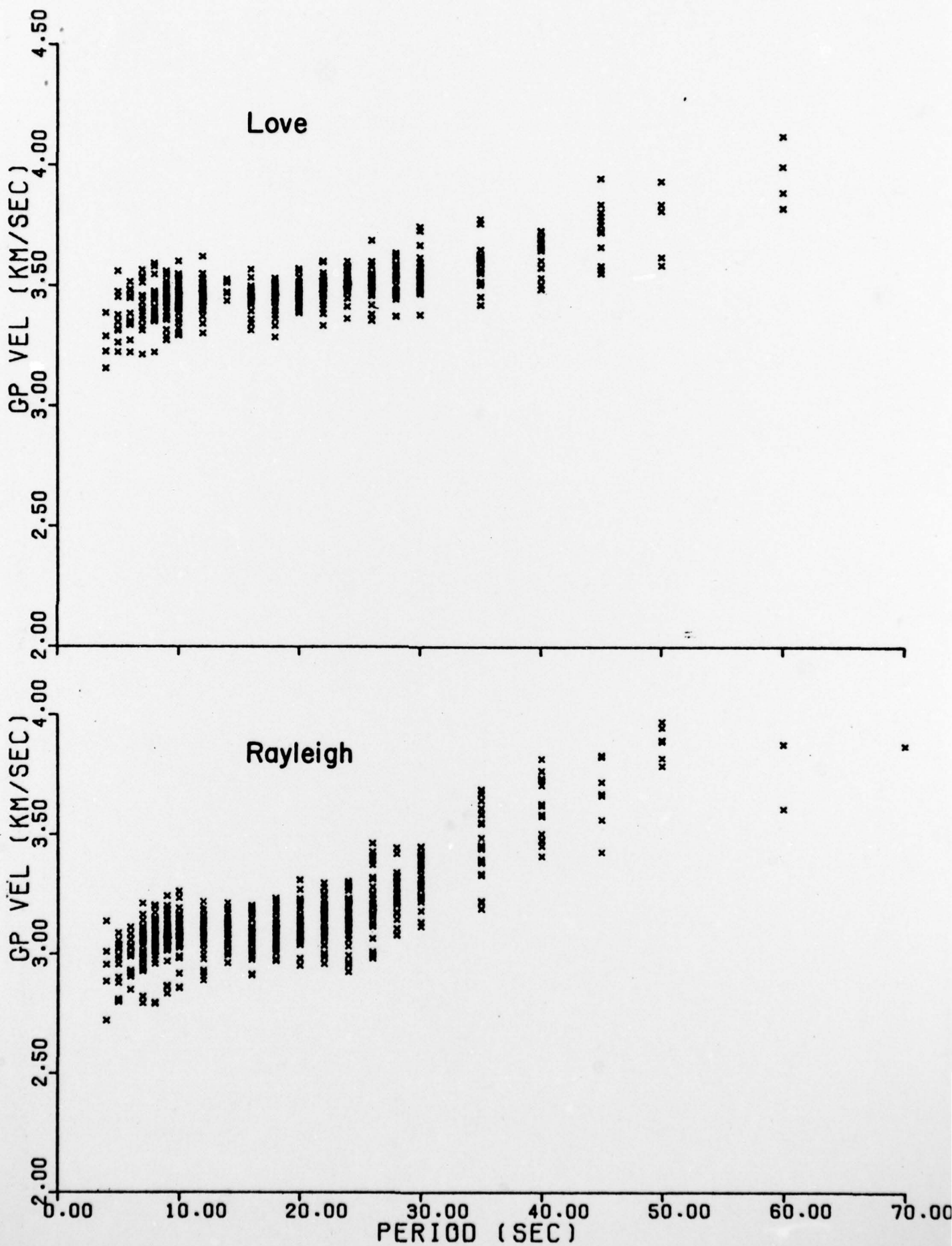


Figure 2



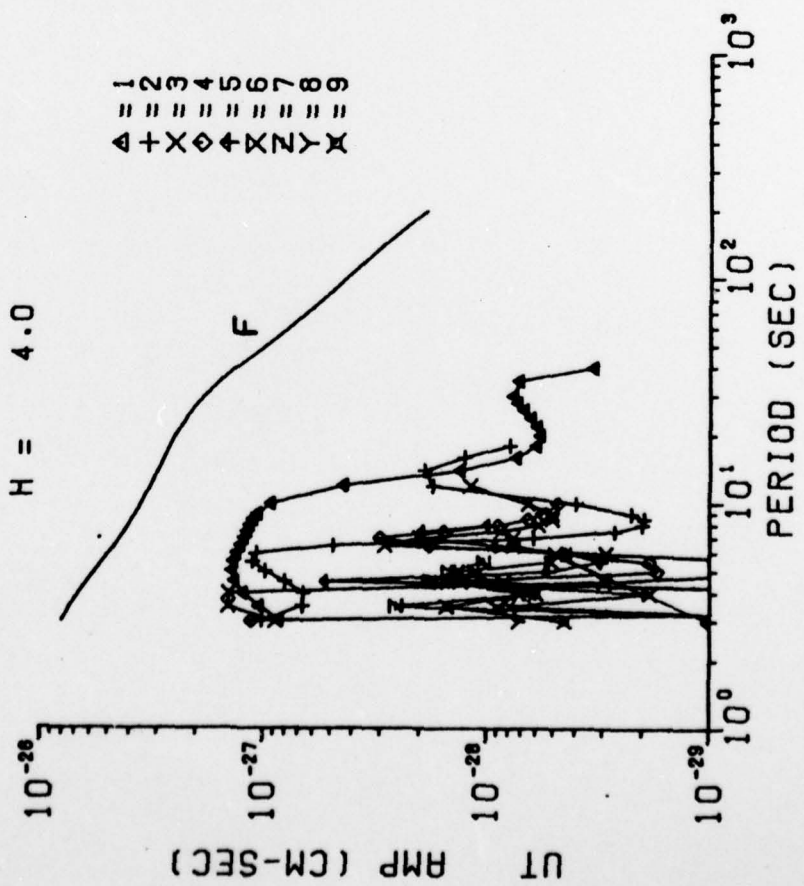
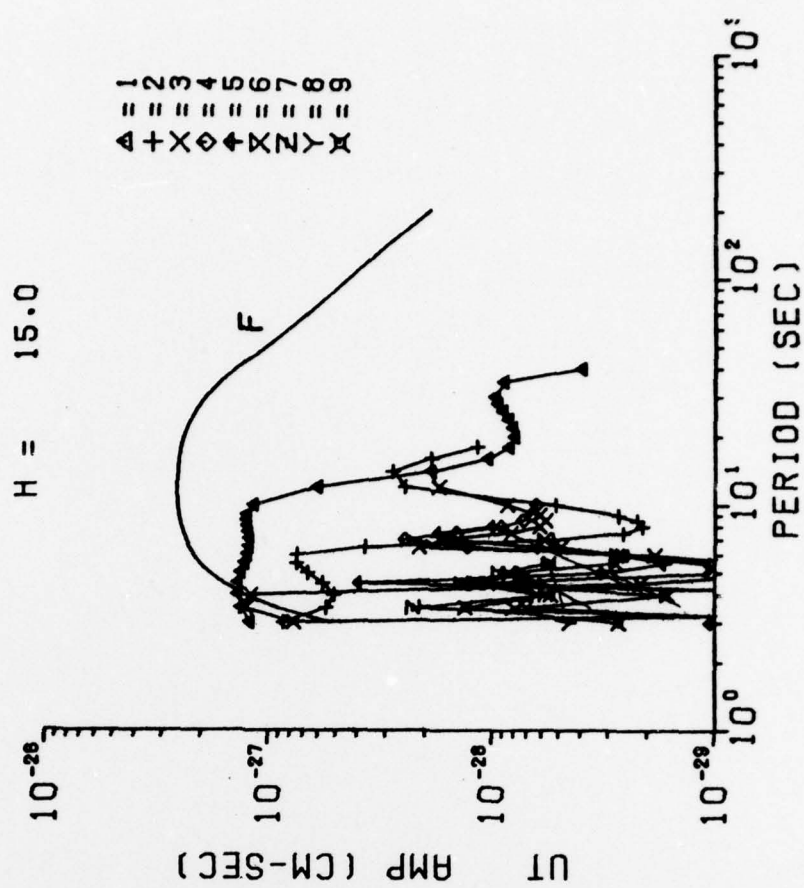


Figure 4

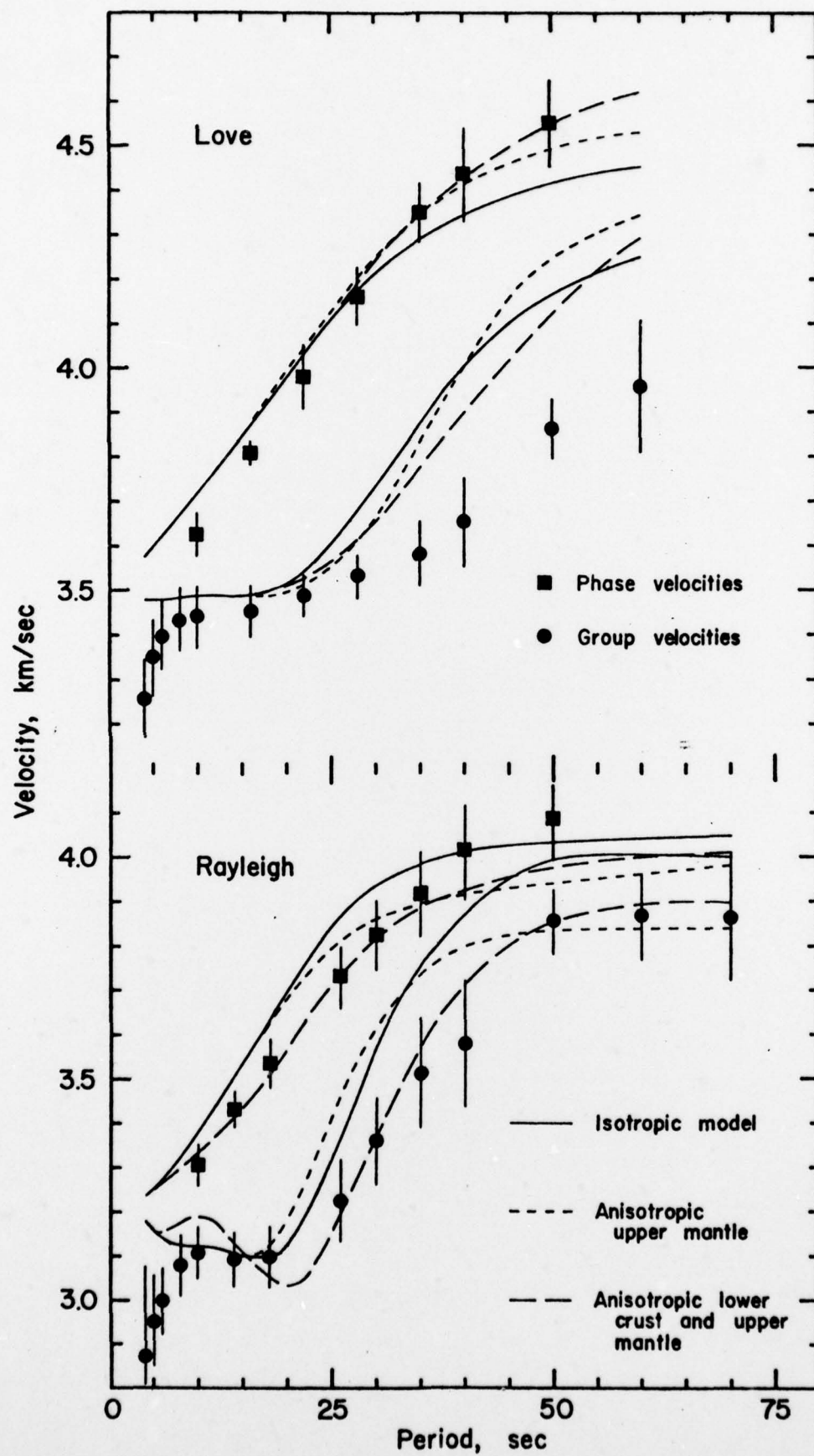


Figure 5

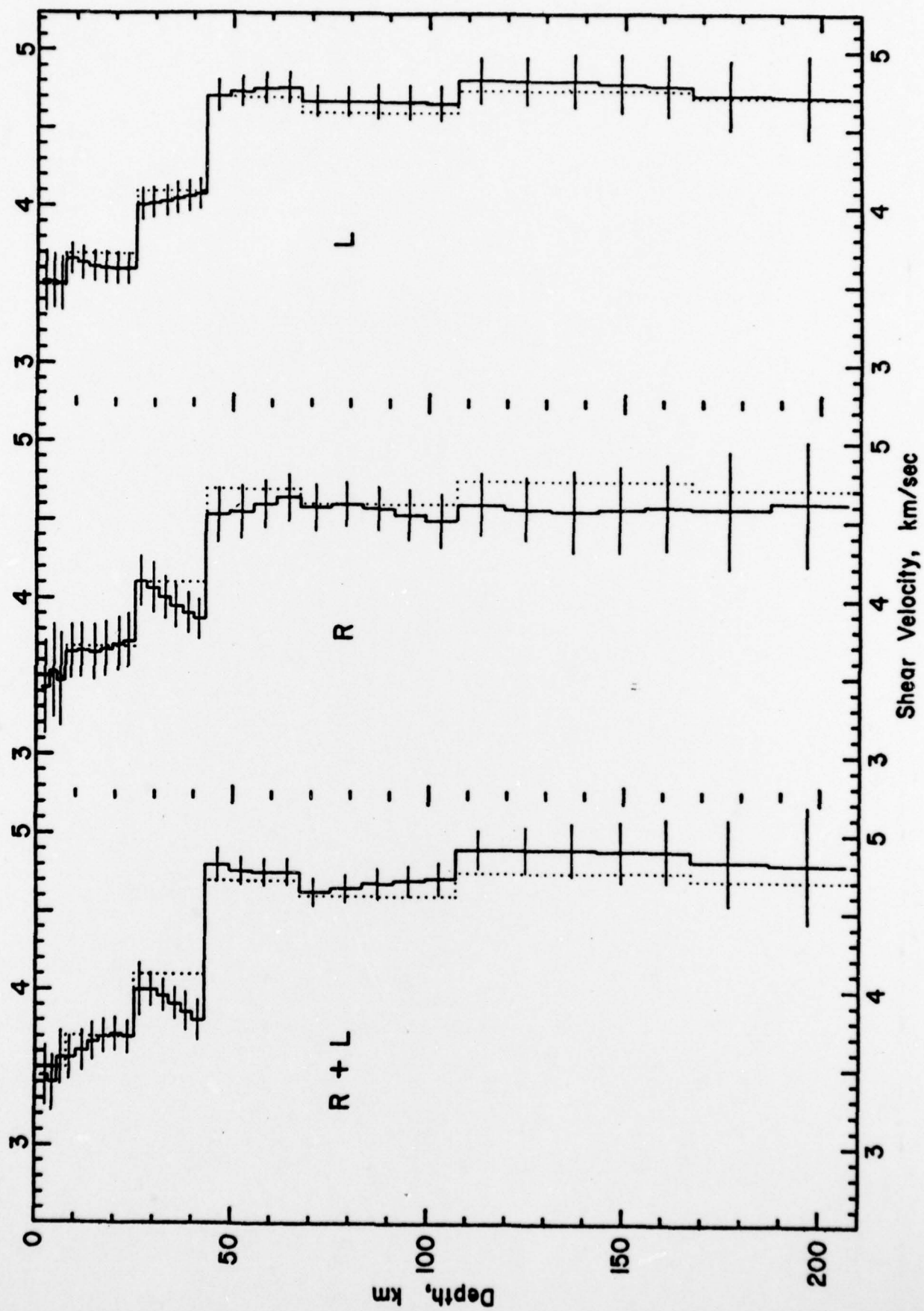


Figure 6

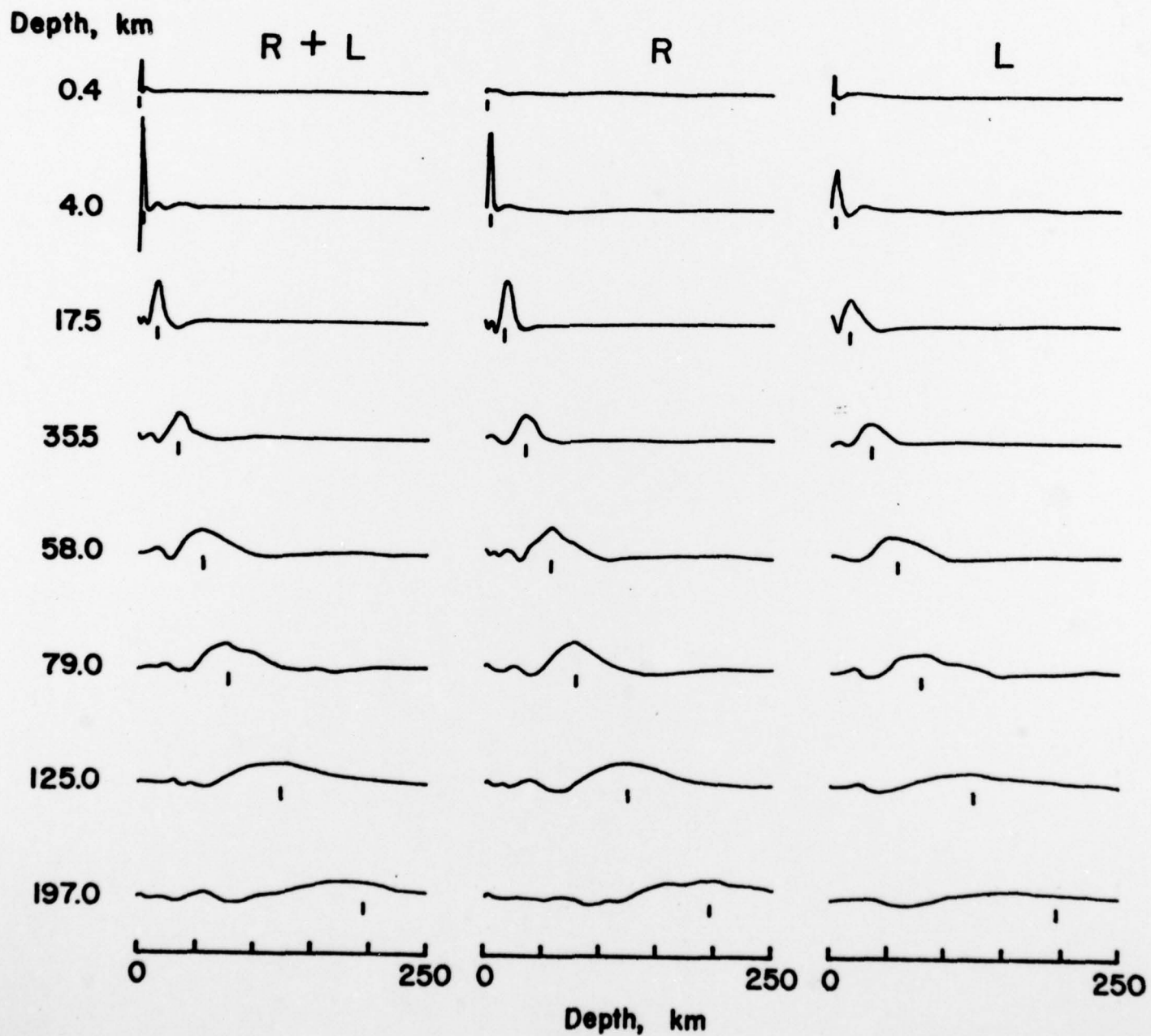


Figure 7

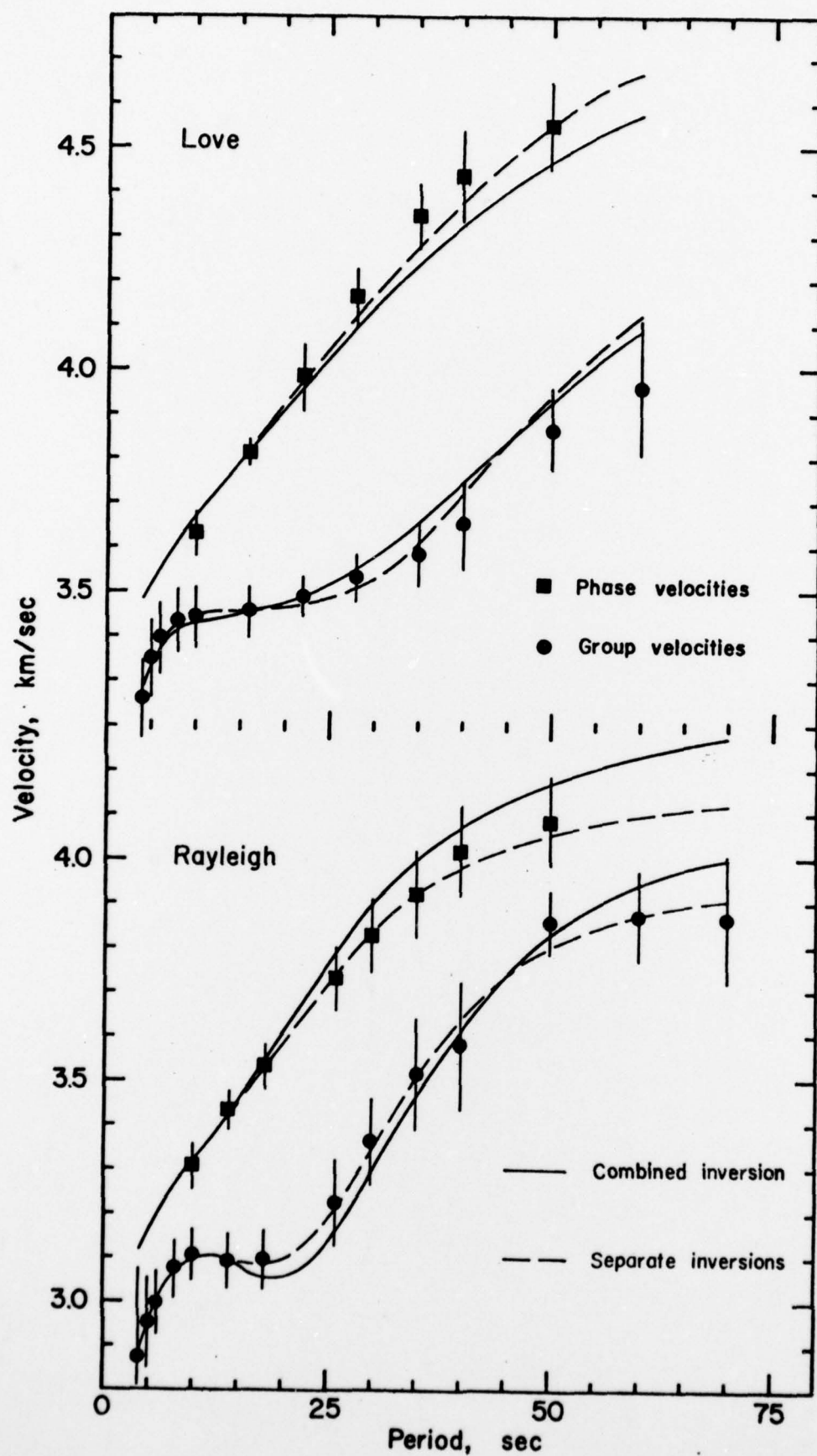
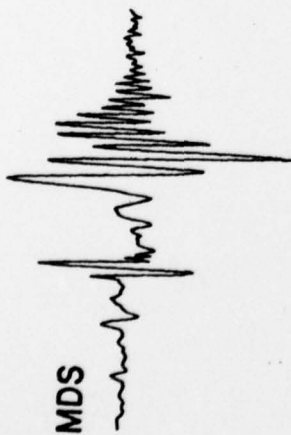


Figure 8

Rayleigh

MDS



AAM



BLA



ATL



Love

MDS



AAM



BLA



ATL

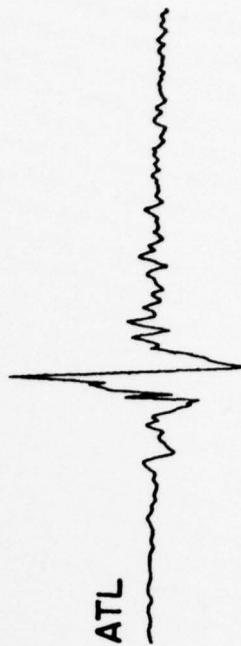
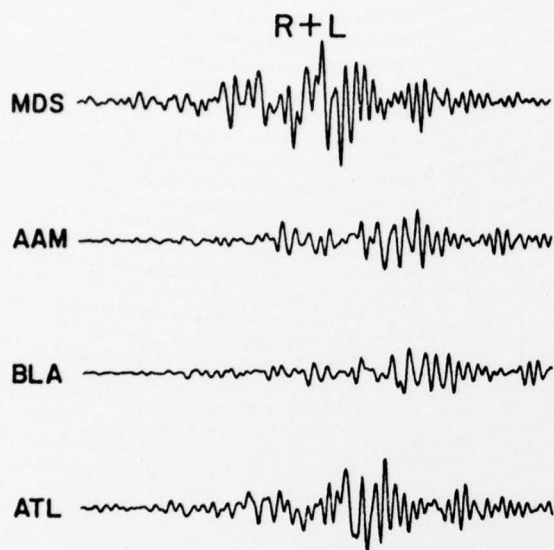


Figure 9

Rayleigh



Love

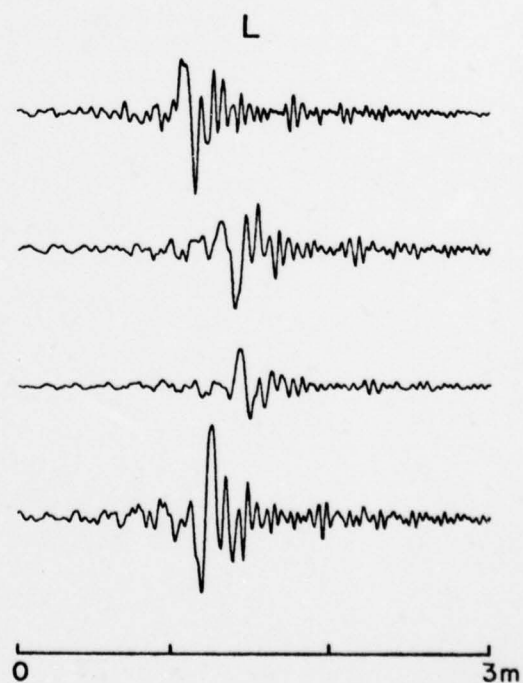
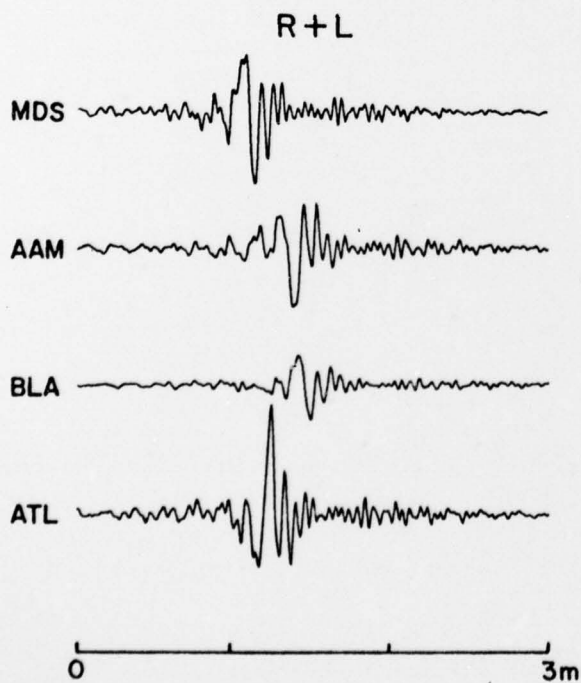


Figure 10