G-35-678

SUMMARY REPORT

A STUDY OF SEISMICITY AND EARTHQUAKE HAZARD IN NORTHERN ALABAMA AND ADJACENT PARTS OF TENNESSEE AND GEORGIA

By

Leland Timothy Long and Jeih-San Liow

Prepared for

GEOLOGICAL SURVEY OF ALABAMA
University of Alabama 35486

Report Period Covered: May 15, 1980 - July 31, 1985

Under

Contract No. GSA-80-3053

November 1985

GEORGIA INSTITUTE OF TECHNOLOGY
A UNIT OF THE UNIVERSITY SYSTEM OF GEORGIA
SCHOOL OF GEOPHYSICAL SCIENCES
ATLANTA, GEORGIA 30332

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Atlanta, Georgia 30332

TABLE OF CONTENTS

LIST OF FIGURES

LIST OF TABLES

Program Objectives for May 15, 1984, to July 31, 1985

The Seismic Net

Convert two stations (OCA and one to be determined) to three-component operation using S-13 seismometers.

Digital Recording

Use digital event recorders to record data from seven to nine seismic stations or from the three-component stations. The digital data will be used to study arrival times and amplitudes of secondary phases.

Crustal Velocity Studies

Continue to study the travel times of crustal phases and refine the velocity model for the crust. Seismic refraction data will be obtained using portable recorders to monitor commercial blasts. A two-dimensional ray tracing technique will be used to facilitate the understanding of the proposed crustal models.

Earthquake Locations

The results of the seismic refraction study will be used to make corrections to estimates of earthquake epicenters and focal depths. A new technique for location is being developed which uses independent computation methods for obtaining the origin time, epicenter, and depth.

Crustal Thickness Studies

A preliminary study of the depth of the crust-mantle transition zone will be extended by collection of more data. Work will continue on the identification of reflections from the base of the crust. Studies of the gravity field will be undertaken to refine the definition of crustal structures.

Focal Mechanism Studies

Study the focal mechanisms of the earthquakes in southeastern Tennessee, northern Alabama, and Georgia. Seismicity patterns will be examined for tectonic implications. Special emphasis will be placed on events occurring within the Georgia Tech seismic net.

Focal Mechanism Determination Techniques

The program to find focal mechanisms, FOCALSR, will be revised to include confidence levels, SH-wave data, and more convenient input and output interfaces.

Attenuation Studies

Determine the decay in amplitude of the Lg phase with distance and calibrate the magnitude scale currently being used.

Reports and Coordination with Adjacent Seismic Nets

Continue to provide reports on the seismicity of northern Alabama and southeastern Tennessee.

Summary of Progress on Program Objectives

The Seismic Net

The Georgia Tech-Geological Survey of Alabama Seismic Net is a twelve-station array extending along the Southern Appalachians from southeastern Tennessee to southwestern Alabama. The net is designed to monitor the southwestern terminus of the Southern Appalachian Seismic Zone. The seismic coverage is supplemented by stations near Rome, Georgia, supported by the Georgia Power Company, and stations near Carters Dam, Georgia, assisted by the U.S. Army Corps of Engineers, Mobil District. The Georgia Tech seismic network also includes stations in central Georgia and along the Savannah River. Seismic stations in the vicinity of the Richard B. Russell Reservoir are supported by the U.S. Army, Corps of Engineers, Savannah District. The distribution of stations operational during the period of July 1980 through July 1985 is shown in Figure 1. Table 1 lists the stations, their locations, and period of operation.

The seismic stations all consist of vertical-component short-period (free period 1.0 Hz) seismometers, broad-band amplifiers, and voltage-controlled oscillators, for data transmission over phone or radio communication systems. The field instrumentation is sealed in PVC pipe and housed in a cinder block enclosure on a cement pad. Station CBT was installed as a three-component short-period system in August 1984. An objective of the project for the May 1984 to July 1985 period was to convert two stations (OCA and one to be determined) to three-component operation using available S-13 seismometers. Station OCA was converted to three-component recording in August 1985, and station MLA is scheduled for installation as a three-component station in October 1985. Station ETG, which had been supported previously by the NRC and Georgia

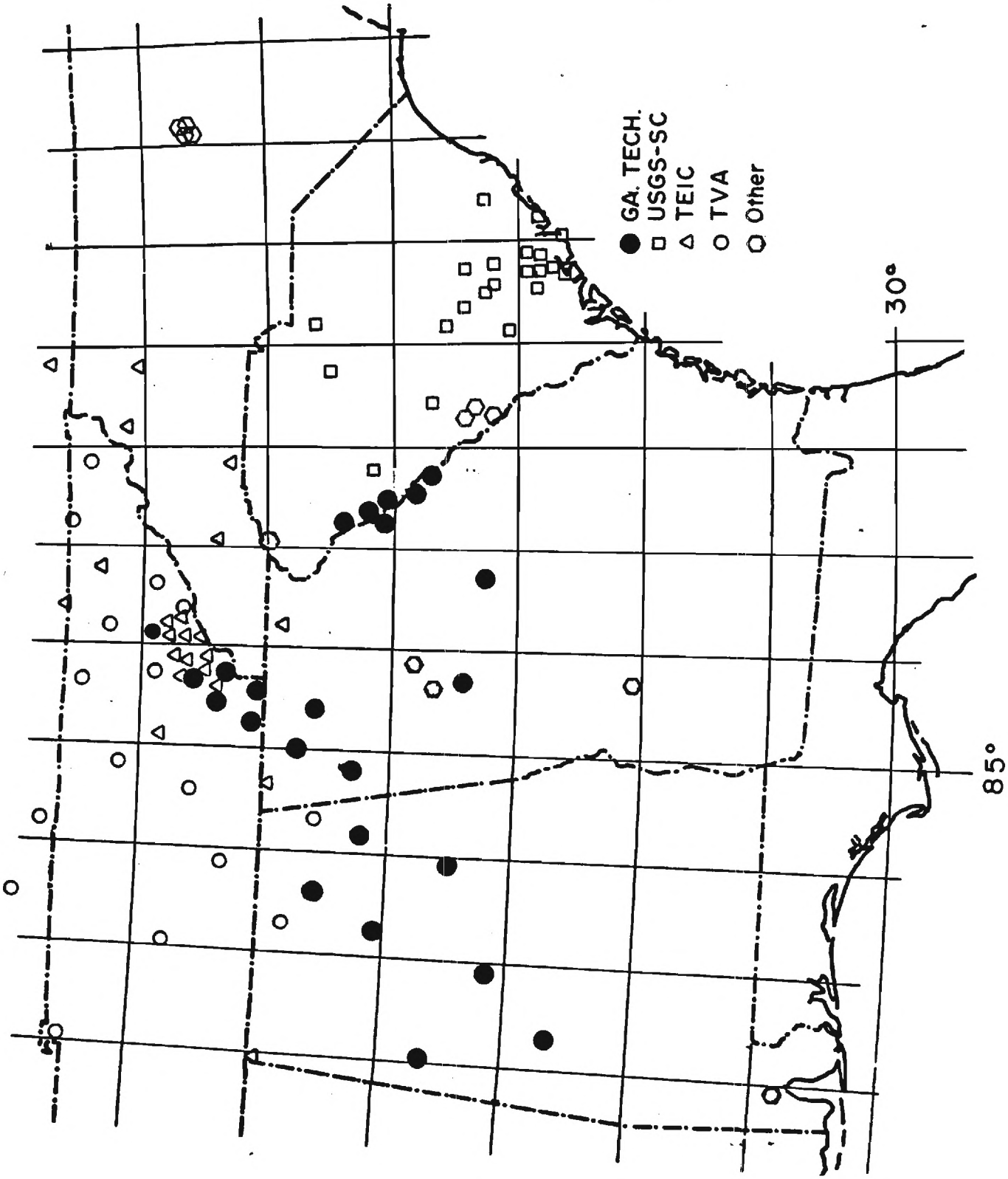


Figure 1. Locations of Seismic Stations in the Georgia Tech Seismic Net.

Table 1. Stations Names, Locations, and Period of Operation for Stations in the Georgia Tech Seismic Net.

STATION	LATITUDE (DEG-N)	LONGITUDE (DEG-W)	ELEVATION (KM)	OPERATION PERIOD
Northern and Central Alabama				
HGA	34.2602	85.8463	0.384	Since April 1981
HVA	34.0264	86.7692	0.195	Since April 1981
MLA	32.7055	87.6937	0.055	Since April 1981
OCA*	34.6138	86.4352	0.250	Since April 1981
TDA	33.5417	86.0247	0.181	Since April 1981
TSA	33.2562	87.0675	0.180	Since April 1981
BKA	33.6338	87.9690	0.122	Since April 1981
Southeastern Tennessee				
CBT**	35.5393	84.4207	0.357	Since April 1982
RCT	35.3453	84.6613	0.265	Since April 1982
RHT	35.0781	84.8825	0.299	Since April 1982
TLT	35.3012	84.2833	0.512	Since April 1982
DCT	35.0542	84.4194	0.508	Since April 1982
Clark Hill and Richard B. Russell Reservior, Georgia				
CH5	33.7332	82.3118	0.114	Since January 1976
CH6	33.8938	82.5291	0.130	Since January 1976
IVA	34.2721	82.7460	0.168	Since June 1980
BEV	34.0893	82.7333	0.158	Since June 1980
LDV	34.1478	82.6833	0.162	Since June 1980
CHF	34.0247	82.5867	0.152	Since Feburary 1984
Carter's Dam and Rock Mountain, Georgia				
CDG	34.6108	84.6667	0.333	Since January 1978
CRG	34.6589	84.5822	0.422	January 1978-August 1984
DALG***	34.7755	85.0137	0.457	Since August 1984
TVG	34.3772	85.3023	0.323	Since July 1977
Wallace Dam, Georgia				
ETG	33.2912	83.3507	0.137	Reinstalled July 1985

* OCA was converted to a three-component system in July 1985.
** CBT was converted to a three-component system in August 1984.
*** CRG was moved to DALG and was kept on operating.

Power Company was reinstalled in the summer of 1985. However, continuous recording has not been achieved because of unanticipated changes in land use.

Digital Recording

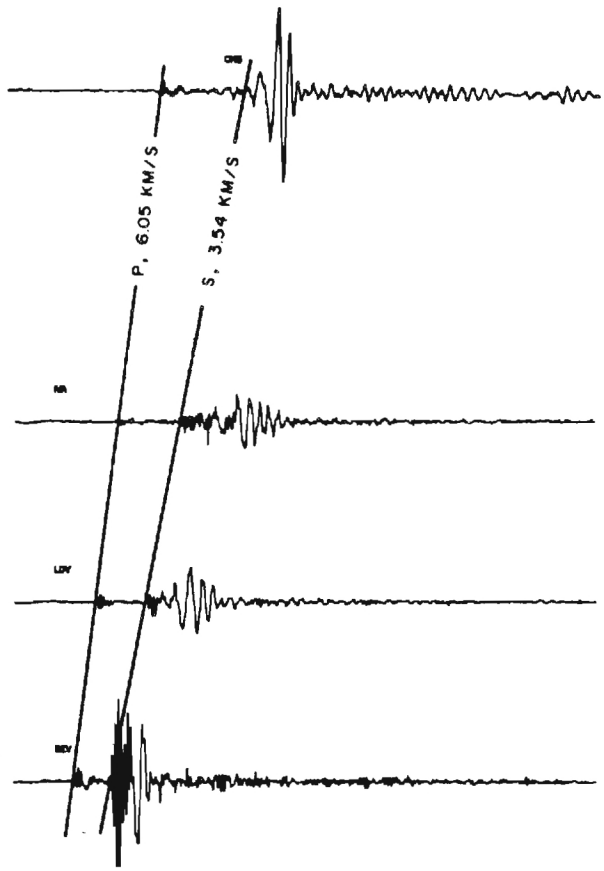
Digital recording capabilities are limited to three three-component digital event recorders. These instruments provide data on nine channels, or on seven channels if set up to trigger on a common station. One event recorder must be taken off line to save and plot the recorded events. A "C" language program has been written to read data from the event recorders and store the data on disk in coded or decoded form. The program "FLUSH" also will plot the seismograms on a dot matrix graphics printer. Alternately, the data can be transmitted to the Georgia Tech mainframe computers and saved on magnetic tape or analyzed using a Tectronics 4114 graphics terminal. The advantage of this system is access to digital recordings of some events and blasts. The disadvantages include incomplete recording of events and many missed events. Most of the recording efforts have been directed toward capturing blasts and events on the three-component stations.

The existing analog data have allowed first-order determination of event location and focal mechanism. Without an extensive increase in the number of stations, refinement in these results will be limited. The only realistic way to develop a significant increase in resolution and understanding of the seismicity is to record all channels on a digital monitor. Digital data will allow spectral analysis of the data. Spect5ral information can be used to study such parameters as attenuation, stress drop, and scattering. The ability to vary the plotting scale for digital data will allow identification of reflected phases and the fine structure of crustal layering.

Crustal Velocity Studies

The results of seismic refraction studies put together by Russell Propes in his Masters Thesis indicate a coherent velocity structure for the granitic crust in the southeastern United States. The model consists of a gradient in crustal velocity which is consistent with the depth variation in velocity predicted by laboratory studies. In this study, four data sets were examined for apparent velocity and the depth of penetration of the rays. None of the study areas were in the area of the East Continent Gravity High, which probably explains the higher (i.e., 6.7 km/s) velocities observed in some studies.

A refraction line in the Richard B. Russell Reservoir area was compiled from recordings of explosions from the several crushed stone quarries in the area. The refraction line extends northwest along the Savannah River. The line is perpendicular to the refraction line by Kean and Long (1980) in the Piedmont. The travel-time curve (Figure 2) indicates a P-wave velocity of 6.05 km/s, which is identical to the



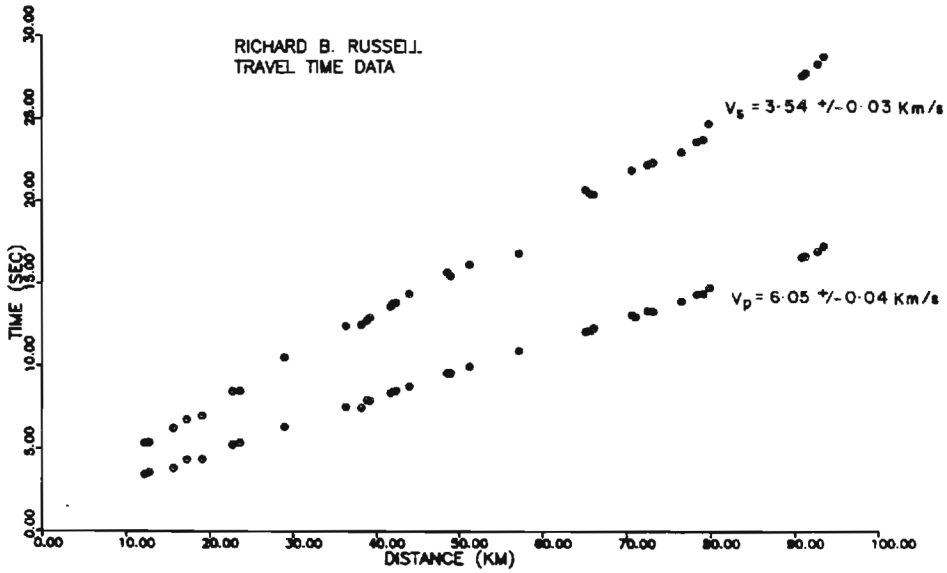


Figure 2. Travel-Time Curve for the Richard B. Russell Area.

velocity found by Kean and Long (1980) and thus does not support the existence of anisotropy suggested by Dorman (1974). Ray path modeling indicates that this velocity is representative of depths of 3 to 8 km.

The refraction line in southeastern Tennessee was centered near Ducktown, Tennessee, at the western edge of the Blue Ridge structural front. Industrial explosions from the Tennessee Chemical Company's Ducktown open pit copper mine were used as sources. The profile is a 35-kilometer line oriented approximately parallel to the regional strike of the Appalachian trend (N 17° E). Portable smoked paper systems were used to achieve a station spacing of two kilometers. The locations of the stations and blasts are shown in Figure 3. The travel time curve (Figure 3) indicates a P-wave velocity of 5.91 $\pm$ 0.08 km/s. Ray path modeling indicates that this velocity represents rocks in the depth range of 0 to 3 km.

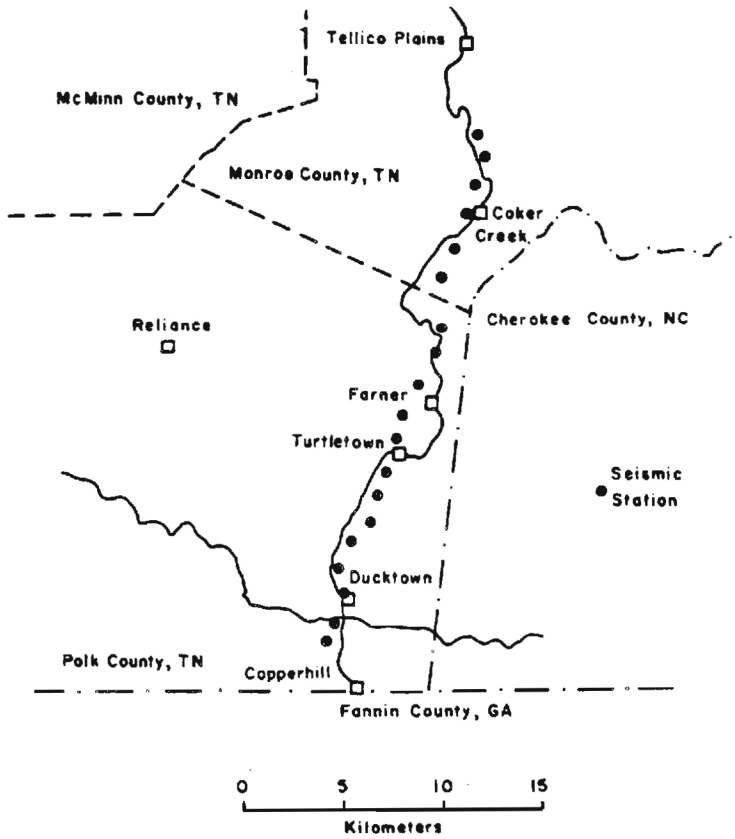
With larger explosions which can be recorded at greater distances, the deep crustal velocities can be determined as well as the shallow crustal velocity. The larger aperture of the Alabama Seismic Net was used to sample the deeper crust (see Figure 1). The travel time data are shown in Figure 4. A P-wave velocity of 6.15 km/s is indicated. Ray path modeling (Figure 5) shows that these velocities represent velocities in the depth range of 5 to 10 km.

Travel times from deep-focus (greater than 10 km) events can behave differently than travel times from events at the surface. The deep-focus events can sample the deeper layers of the crust. A total of 110 paired time and distance values have been collected for this study. A least squares linear regression was used to define the best fit line from which velocities can be computed. The arrivals define three seismic arrivals: Pn, P, and Lg (Figure 6). Although previous studies have interpreded similar data as higher velocity layers in the lower crust, attempts to distinguish the P* arrivals in this data set were futile. The travel-time curve which satisfies the observed arrivals corresponds to an apparent P-wave velocity of 6.3 km/s. Ray path modeling indicates that this velocity is representative of the 12 to 15 km depth range.

The four studies reported above sample the granitic crust at four separate depths. Travel times for events at different depths are shown in Figure 7. These indicate a velocity gradient exists in the crust, and the resulting model is shown in Figure 8. The gradient model will satisfy all the refraction lines observed in the granitic crust in the southeastern United States. The velocity structure can be generalized to a single model from the Georgia Piedmont to the Alabama Valley and Ridge (Figure 8).

Earthquake Locations

Depth determination in a hypocenter solution is made difficult by the coupling in the inversion matrix between origin time and depth. In



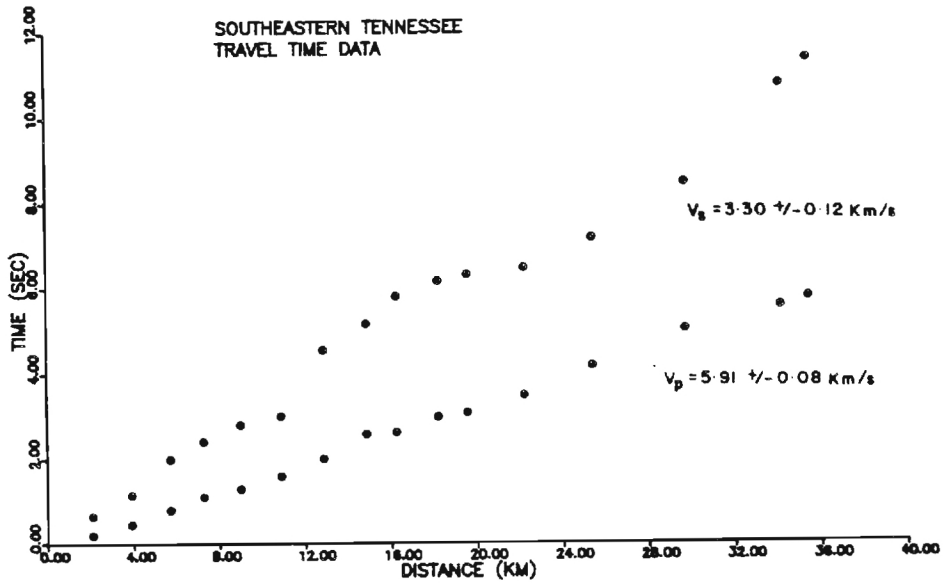


Figure 3. Locations of Recording Sites for the Ducktown Refraction Line and Southeastern Tennessee P-wave and S-Wave Travel-Time Data.

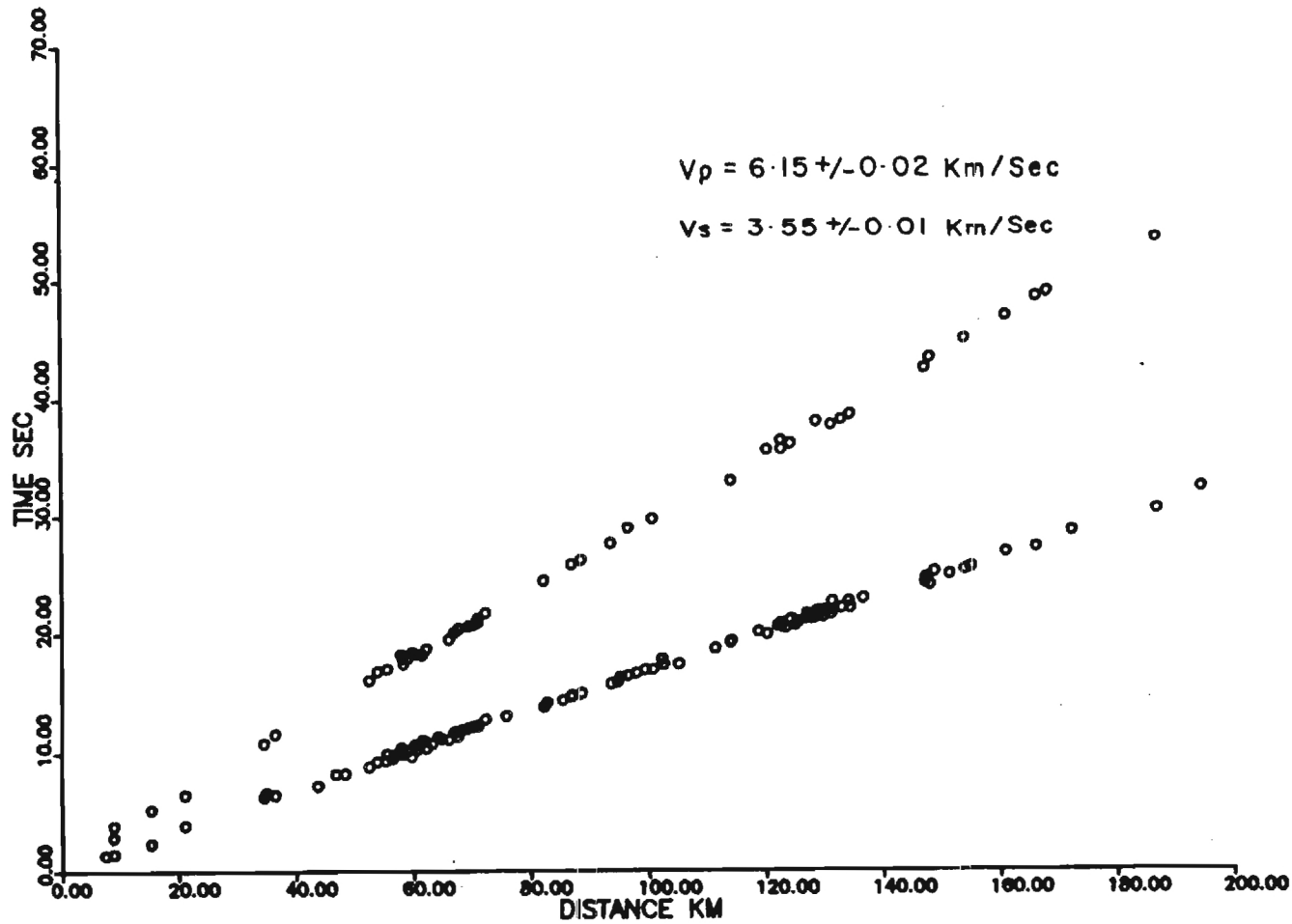


Figure 4. Alabama Travel-Time Relationship (after Liow et al., 1983).

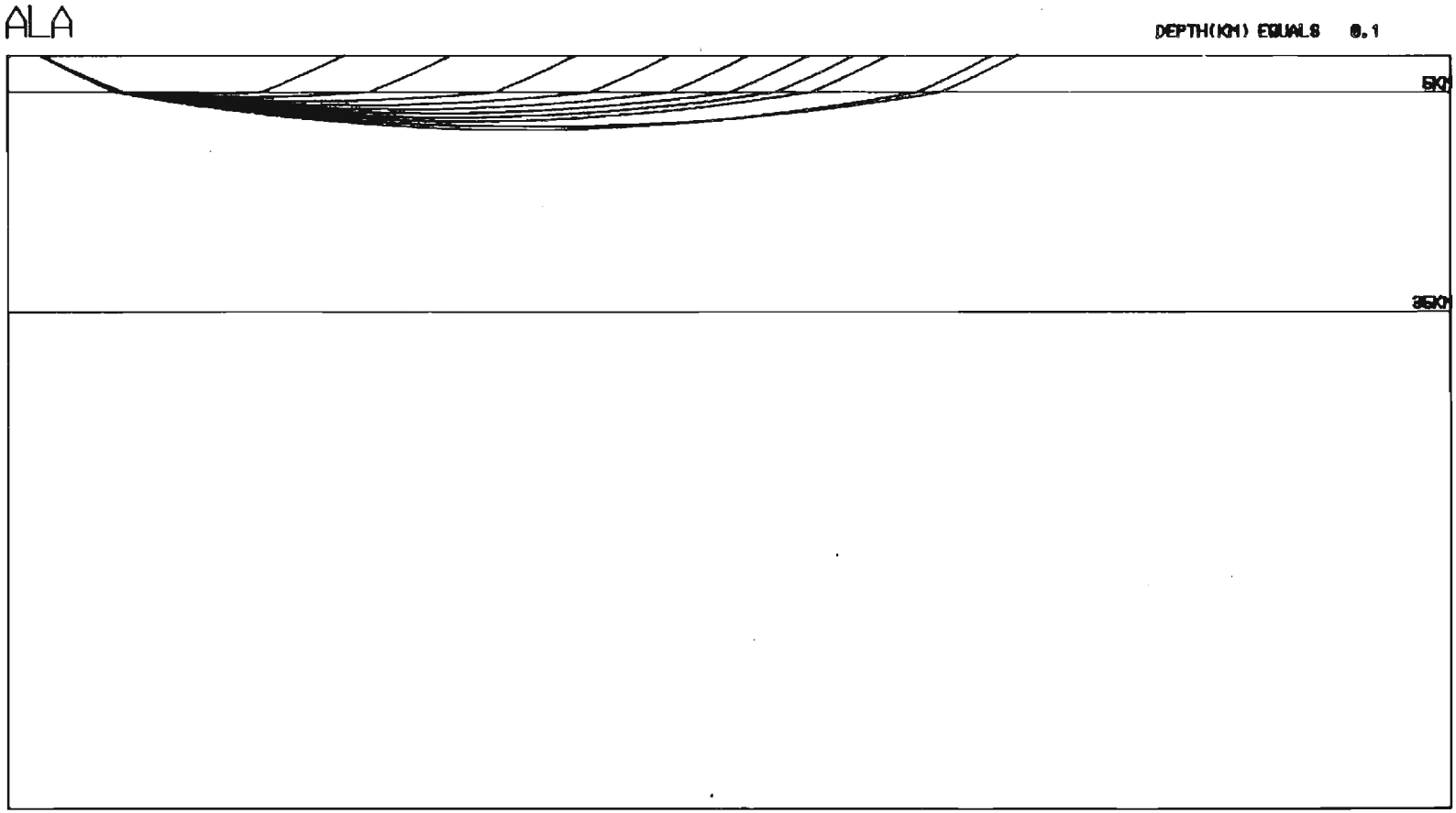


Figure 5. Ray Paths for Alabama Velocity Structure.

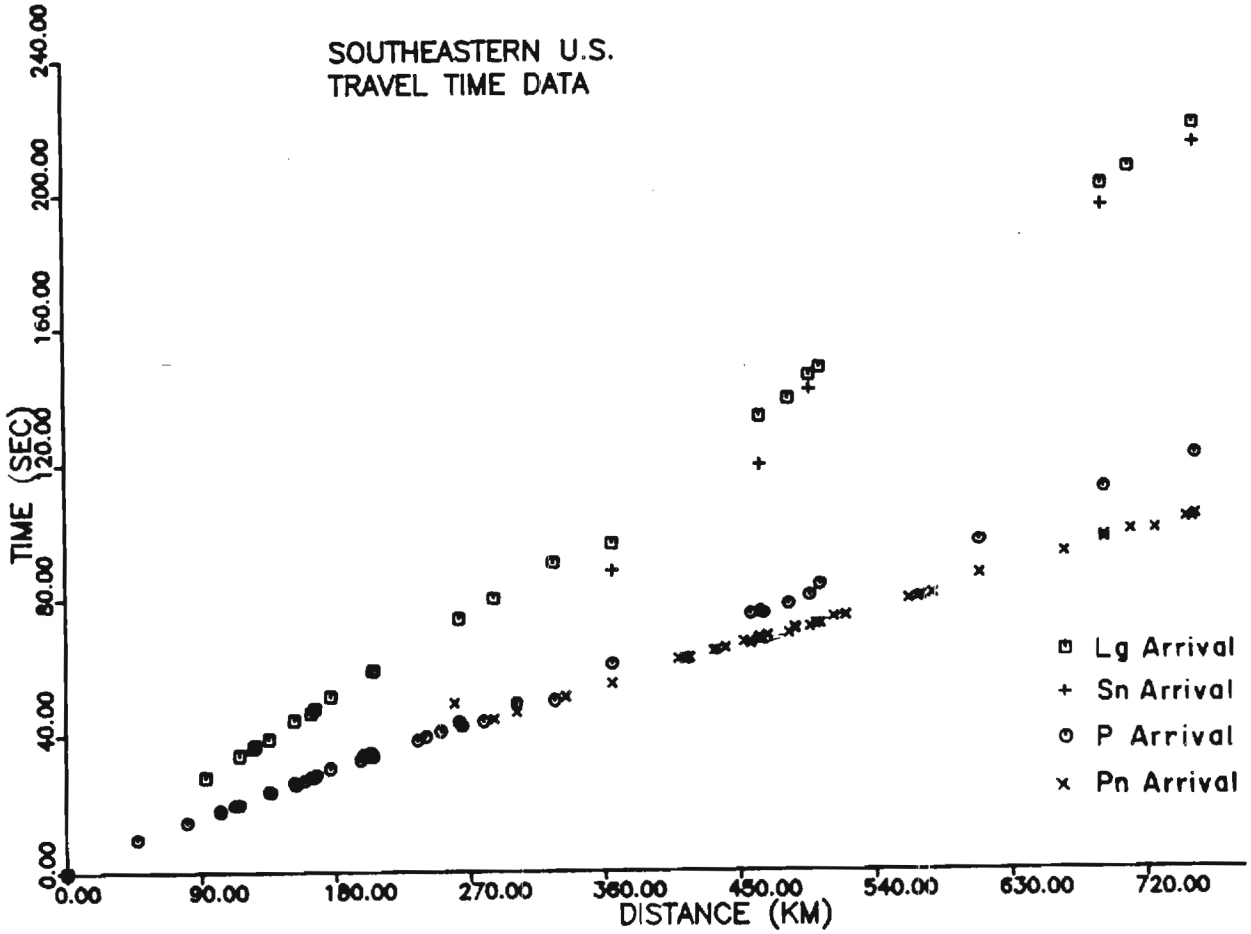


Figure 6. Regional Travel-Time Curve.

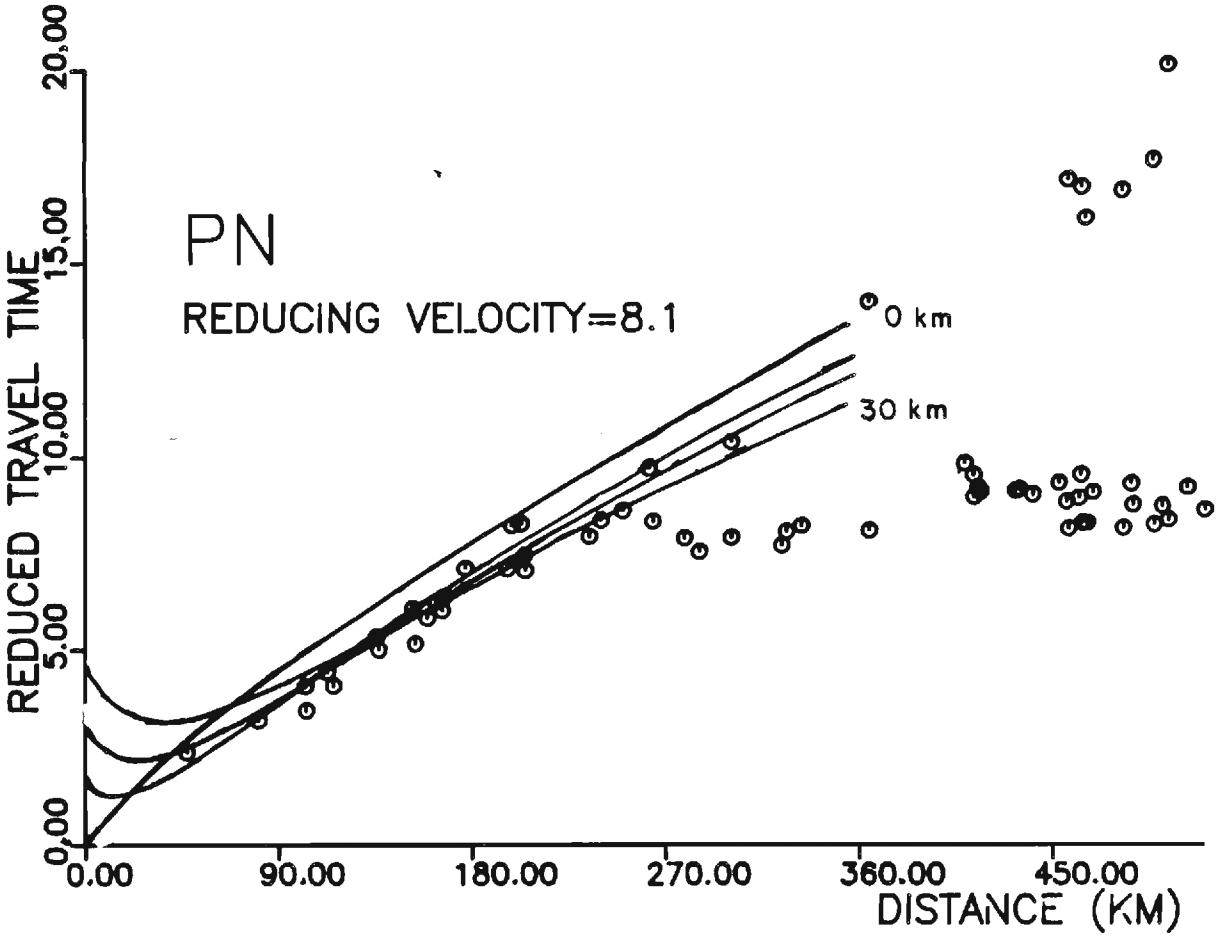


Figure 7. Southeastern Regional Reduced Travel-Time Curve for Source Depths from 0 to 30 Kilometers.

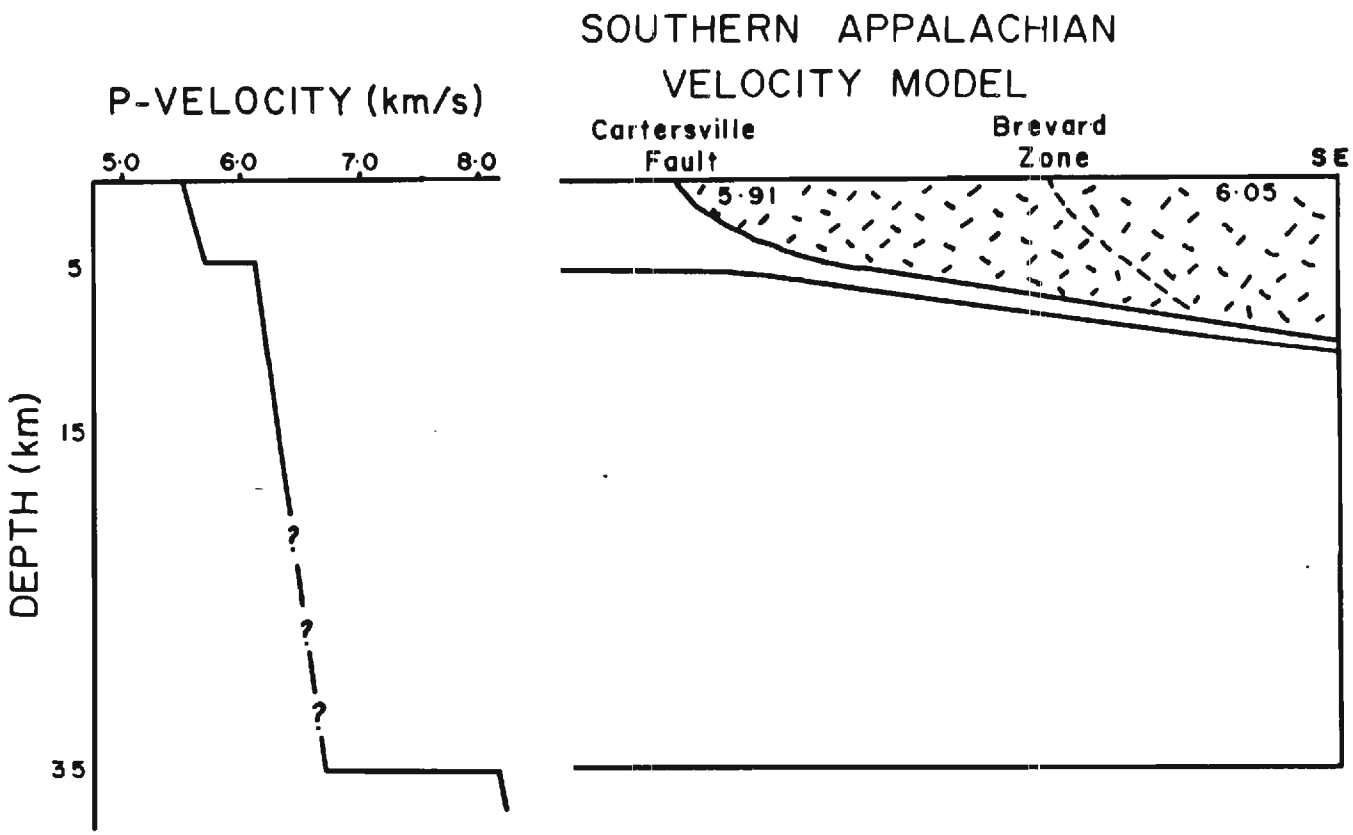


Figure 8. Regional Crustal Structure and Velocity Model for the Southeastern United States.

addition, the existence of velocity gradients and velocity inhomogenieties significantly perturb the relation between travel time and depth. We have made two modifications to the traditional location program to accommodate these difficulties. First, we have isolated the origin time computation by only using events with clear S-P phases to determine origin time. Second, we solve for location and depth independently to take advantage of separate systems of weights, each a unique function of distance. Finally, we have modified the travel-time computation subroutine to accommodate the gradient in velocity developed from refraction data. We expect this new location program to significantly improve the definition of hypocenters in southeastern Tennessee. The structural model for southeastern Tennessee is being tested for influence on location and will also be included in the location program as corrections to the travel-time data.

Crustal Thickness Studies

A method of computing equivalent crustal thicknesses for the various areas of the southeastern United States was developed from the time term method by Kean and Long (1980). This method was used by Long and Liow (1985) and again by Propes (1985). The recent data by Propes are shown in Figure 9 and are consistent with the previous results. The depths range from 45 km (or deeper) under the Southern Appalachians to 27 km near Charleston, South Carolina.

At station CBT in southeastern Tennessee, unique reflections off the crust-mantle interface have been identified. They indicate a crustal thickness of 55 km. The analysis of these reflections is presented in Appendix I.

Focal Mechanisms Studies

The October 9, 1984, event is the largest recent event to occur within the southeastern Tennessee and Alabama seismic net. This event has been studied and a preliminary report on the event is presented in Appendix II. Pending revision of the focal mechanism program, other events will be reevaluated for focal mechanism.

Focal Mechanism Determination Techniques

Program FOCALSR was developed at Georgia Tech in order to include the ratio of the amplitudes of P and S waves in determining domains of valid focal mechanism solutions. However, the interpretation of some of the seismograms revealed ambiguities in the ratio of P- and S-wave amplitudes. The two primary ambiguities relate to the angle of incidence and the effect of scattered arrivals. The effect of these ambiguities on focal mechanisms is to bias the solution toward strike-slip mechanisms. Since the majority of the events in southeastern Tennessee

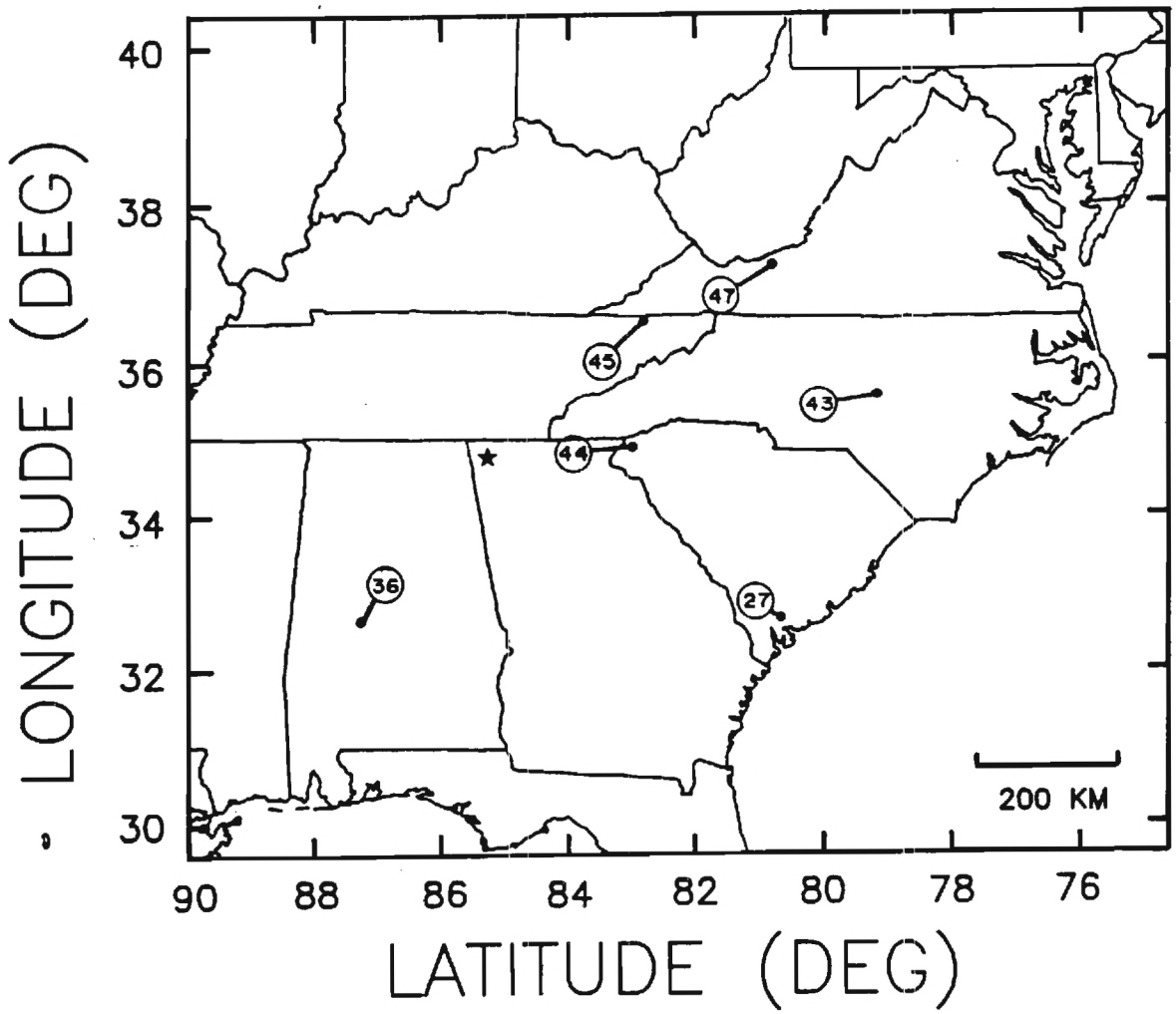


Figure 9. Equivalent Crustal Thicknesses in the Southeastern United States.

have a significant strike-slip component, the significance of the amplitude ratios needs to be examined. Preliminary evaluation (Zelt and Long, 1984) of the amplitude ratios indicate a correlation with regional geology. Other studies on the particle motion of the P wave indicate that only the first cycle represents a direct arrival from the source (Wilson, 1984). The following waves, the P-wave coda, have a dominantly horizontal particle motion which indicates near-surface scattering of the P wave.

Attenuation Studies

Efforts to define attenuation of various phases has been directed toward the Alabama Seismic Net. Earthquakes which excite the crustal channel are rare. Consequently, signals from explosions have been used. The main difficulty in the use of explosions is their strong excitation of surface waves which propagate as the fundamental mode in the near-surface sediments. Because they are often set off as line explosions on the face of a quarry wall, they may also be highly asymmetric. In our preliminary analysis we have attempted to isolate the Lg phase from the surface waves. The preliminary results imply a Q value significantly lower than the 600 to 1000 values in adjacent areas. The spectral data also indicate that energy in the 6 to 10 Hz range is propagated more efficiently than the lower frequencies. The preliminary interpretation of this frequency dependence on Q is that this is a surface layer effect.

Reports and Coordination with Adjacent Seismic Nets

Seismic arrivals and event locations are submitted to and included in the Southeastern United States Seismic Net Bulletin published out of Virginia Polytechnic and State University. A special report on the Columbus earthquake has been submitted to EARTHQUAKE NOTES.

Figure 10 shows the epicenters of earthquakes located in 1984-1985. The revised locations are given in Table 2.

EPICENTERS FROM MAY 1, 1984

TO JULY 31, 1985

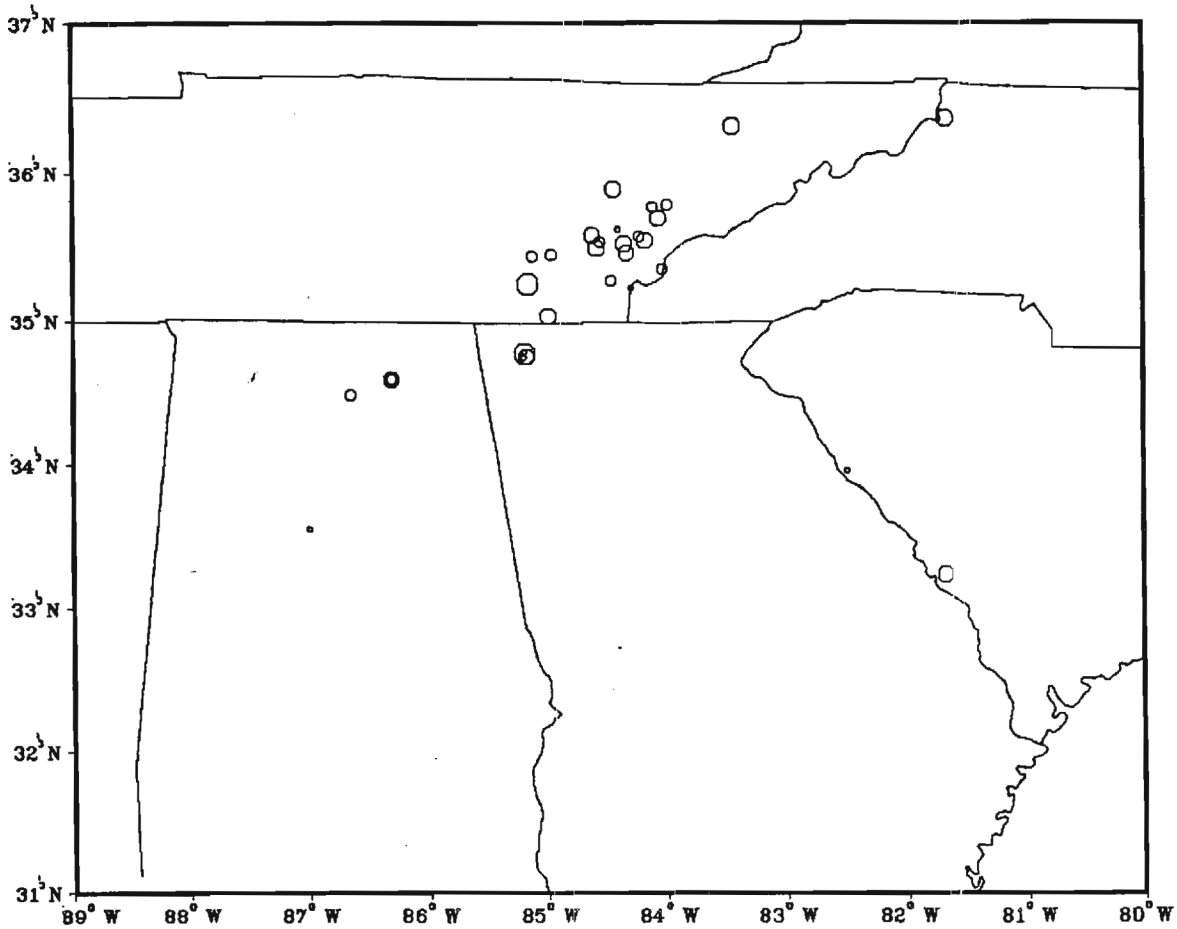


Figure 10. Locations of Earthquakes During the Period May 1, 1984, to July 31, 1985.

Table 2. Events Located by the Georgia Tech Seismic Net Between May 16, 1984, and July 31, 1985.

Yr/Mo/Da	Time (UTC)	Mb	Latitude	Longitude	Depth	
84/05/19	03:46:29.88	2.1	36.3069	83.4420	0.0*	Morristown, TN
84/05/20	07:48:24.06	1.1	35.6417	84.3856	0.0*	Athens, TN
84/05/25	10:15:38.59	2.1	35.5736	84.6246	24.53	Athens, TN
84/06/07	16:57:49.86	2.7	35.5381	84.1761	23.05	Madisonville, TN
84/06/23	14:03:57.09	1.5	35.7546	83.9718	19.05	Maryville, TN
84/08/09	02:42:35.24	2.9	34.6017	86.3036	15.4	Huntsville, AL
84/08/09	07:35:07.79	1.0	34.6059	86.3028	8.6	Huntsville, AL
84/08/24	19:47:39.90	1.4	34.4967	86.6419	21.6	Huntsville, AL
84/08/30	16:26:26.34	2.0	35.5613	84.3429	0.0*	Greenback, TN
84/09/21	20:46:22.48	1.5	35.3672	82.9786	0.0*	NC
84/10/09	11:54:26.25	3.7	34.7752	85.1929		Lafayette, GA
84/10/15	16:56:52.02	2.0	34.7538	85.1754	11.6	Lafayette, GA
84/10/22	18:58:41.20	2.7	36.3609	81.6802	2.1	GFM, NC
84/11/30	19:06:03.29	1.4	35.5300	84.5570	10.0	Madisonville, TN
84/12/17	18:48:27.64	2.2	35.4539	84.3349	17.2	Vonore, TN
84/12/23	07:22:44.28	1.8	35.4325	85.1275	18.6	Graysville, TN
85/01/25	02:13:34.23	1.8	35.4441	84.9682	18.6	Calderwood Dam, TN
85/03/09	14:29:57.67	2.3	35.0291	84.9942	11.0	Apison, TN
85/03/12	08:57:43.23	1.6	35.2687	84.4654	12.6	Servilla, TN
85/03/12	13:04:44.75	2.0	35.8882	83.4429	0.0*	Harrisburg, TN
85/04/09	21:41:00.73	1.3	35.7663	84.1120	21.7	Maryville, TN
85/04/10	10:53:59.25	2.4	35.6903	84.0674	18.6	Maryville, TN
85/04/20	04:21:02.39	2.5	35.4858	84.5871	22.2	Athens, TN
85/04/28	07:04:23.59	1.0	35.5646	84.2270	9.3	Madisonville, TN
85/05/23	05:29:38.66	0.4	35.2188	84.2929	0.0*	Tellicoplain, TN
85/06/09	00:38:42.55	2.5	33.2368	81.6837	0.0*	Barnwell, SC
85/06/14	00:30:37.78	1.8	35.3472	84.0312	0.0*	Tellicoplain, TN
85/07/12	18:20:28.25	3.0	35.244	85.166	19.7	Chickamauga Lake, TN

* Depth fixed at zero (not computed).

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APPENDIX I

ANALYSIS OF REFLECTIONS FROM THE BASE OF THE CRUST IN SOUTHEASTERN TENNESSEE

By

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Abstract

Three-component digital recordings of commercial blasts and a small earthquake located near station CBT in southeastern Tennessee show clear signatures for the shear wave reflected off the base of the crust. The travel times and particle motions indicate that the SmP phase was recorded from the blasts and the SmS phase was recorded from the earthquake. The near-vertical shear wave was preceded by 0.7s by a conversion from S to P near the base of the Paleozoic sediments. The computed crustal thickness of 55 km (±1.5 km) agrees to within 3 km with results obtained by the time term method for sites 25 to 40 km away. The similarity of the reflected waveform to the direct S wave limits the thickness of the crust-mantle transition zone near station CBT to less than 2 km.

Introduction

The seismic reflection method has rapidly improved the resolution of crustal structures. In the Southern Appalachians the COCORP traverse (Cook and others, 1983) has revolutionized tectonic concepts and advanced our understanding of structures at shallow to mid-crustal depths. However, resolution of the crust-mantle transition zone is inconsistent. In some regions the Mohorovicic discontinuity identified from refraction data also generates a moderately strong signature in reflection data. One area where refraction data (Kean and Long, 1980) and reflection data give similar results for the thickness of the crust is the Piedmont of Georgia along the COCORP Southern Appalachian traverse. Another region is the British marginal sea in the BIRPS data (Peddy, 1984). In other regions a signature for the crust-mantle transition zone can not be identified. The Tennessee segment of the Southern Appalachian traverse is an example of an area lacking reflections from the crust-mantle transition zone. The lack of such reflections in seismic reflection profiles could be explained by the greater depth to the base of the crust or the limited penetration of seismic energy through the overlying geological structures.

In the Southern Appalachian area, the COCORP seismic reflection traverse covers four different geological provinces. Reflections from the mantle are observed only in the Piedmont. In the Cumberland Plateau

in eastern Tennessee, where no reflections from the crust-mantle transition zone are observed, Owen and others (1984) analyzed the crustal structures by using teleseismic data recorded on a broadband station (CPO). They found a transitional crust-mantle boundary extending from depths of 40 to 55 km under the Cumberland Plateau in central Tennessee. They modeled the shear wave velocity in the transitional zone as a laminated gradient. Also, evidence for a thin laminar crust-mantle boundary has been observed from the COCORP Southern Appalachian traverse in the Piedmont Province (Hale and Thompson, 1983). The COCORP reflection line across the Valley and Ridge Province of Tennessee does not show reflections at times corresponding to the expected arrival times for reflections from the crust-mantle boundary. The lack of sharp reflections in the Valley and Ridge indicates either that a sharp crust-mantle boundary does not exist or, perhaps, that it does exist and the reflections are scattered and masked by the complicated overlying crust.

In contrast to the lack of evidence for Moho reflections in the COCORP data and from the analysis of Owens and others (1984), we have observed reflections from some events in the vicinity of our three-component seismic station CBT, which is located in the Valley and Ridge Province in southeastern Tennessee. The digital recording at CBT significantly improved the resolution of these phases and, perhaps, explains why they were not identified on analog records. The purpose of this paper is to analyze and discuss these reflections and their implications concerning the depth and character of the crust-mantle transition zone. First, we discuss the depth of the crust-mantle boundary based on an analysis of travel times. Second, we compare the observed data with theoretical reflections from different transitional boundary models.

Crustal Thickness Analysis

The three-component seismic station at Christenburg, Tennessee (CBT), has been monitored with a digital event recorder since August 1984. On three of the commercial blasts (events 1, 2, and 3) and one earthquake (event 4), there exists a phase located near 21 seconds after the P arrivals. The three blasts used in this paper are located within a distance of 10 km to the northeast of CBT (Figure 1). Figure 2 shows the three-component trace of event 2 as an example. Event 4 (see the trace in Figure 3), which is located 17.4 km to the southeast of CBT, is a magnitude 2.2 earthquake.

A modification of the standard location procedure used in HYPO71 (Lee and Lahr, 1975) was used here to locate the four events. The modification consists principally of independently computing origin time, depth, and epicenter. First, the S-P time was used to calculate the origin time assuming only a constant Poisson's ratio. This procedure makes the origin time estimate independent of the location or depth. Then, the origin time was held constant and the latitude and longitude were found by a traditional iteration of a least squares solution for a correction to the epicenter. For the three blasts, the depth was fixed at the surface. For the earthquake, the hypocenter was

obtained by iteration of alternate independent determinations of the epicenter and depth. The earthquake had a focal depth of 15.1 km (see Table 1 for hypocenter data).

For the blasts, the ground motions for the reflections are dominated by vertical particle displacements. Plots of the particle motion of the reflection (Figures 4a, 4b and 4c) indicate that the reflections arrive at the surface with angles less than 10 degrees from the vertical. The horizontal movements were typically less than 10 percent of the vertical motion. Therefore, we assume that the incident wave (or the up-going wave) is a P wave. The down-going wave can either be a P or an S wave. However, if both paths were P waves, the computed crust-mantle boundary would be unreasonably deep (e.g., greater than 70 km). Consequently, we conclude that the observed reflections are S waves converted to P waves at the crust-mantle discontinuity.

For the earthquake, the reflection arrival is strong on the vertical component and the horizontal component. A plot of the particle motion (Figure 4d and Figure 5) indicates that the first 0.2 seconds of the reflection is vertically polarized. Afterwards, the particle motion is a mixture of horizontal and vertical movements. The horizontal movement can be identified from the EW-NS plot of Figure 4e. We deduced that the incident wave (or the up-going wave) is an S wave, and interpreted the pure vertical movement in the beginning part of the reflection as a conversion of the up-going S wave to a P wave. This converted P wave indicates the existence of a discontinuity 1.6 ($\pm$0.3) km below the surface, possibly within the sedimentary layer. Again, the down-going wave has to be an S wave so that the computed crust-mantle boundary will not be unreasonably deep. This arrival is the S wave reflected as an S wave at the crust-mantle boundary.

Figure 6 compares the reflections with the shape of their direct S waves. The similarities are evident and thus support the conclusion that the reflections were derived from the shear waves. The reflection on the horizontal component of the earthquake was used in the comparison because the reflection on the vertical component was preceded and disturbed by the conversion from S to P near the base of the sediments.

The seismic station CBT is located in the Valley and Ridge Province, which is characterized by a cover of Paleozoic sediments two to three kilometers thick. The velocity model used here to compute the depth to the crust-mantle boundary was derived from the northern Alabama area (Propes, 1985). The model consists of a 3-km-thick surface layer overlying a granitic crust with a velocity gradient at least down to a depth of 20 km (Figure 7). We think this model represents an appropriate velocity structure in the vicinity of CBT. Whether a continuation of the velocity gradient is representative of the lower crust for this area remains unclear, but the deviation of the computed crust-mantle boundary due to velocity uncertainty in the lower crust is small. For example. in the computation, the P wave velocity is 6.12 km/s at a depth of 5 km and 6.3 km/s at a depth of 15 km; we use the same gradient to extrapolate the velocity down to the crust-mantle boundary. This gives a

7.02 km/s P-wave velocity at a depth of 55 km, which could be higher than an expected 6.7 km/s P-wave velocity in the lower crust. However, the differences in the computation of the depths will be smaller than 1 km because of this difference.

Ray tracing (Propes, 1985) was used to compute the theoretical travel times and epicenter distances along the ray path of the reflection. By comparing these theoretical travel times and epicenter distances with the observed data, we obtain a depth for the crust-mantle boundary of 55 km. Table 2 shows the theoretical and the observed data. The average deviation of the epicenter distances is ± 0.05 km. The average deviation of the travel time is ± 0.26 second, corresponding to, approximately, a variation of ± 1.45 km in the computed depth of the crust-mantle boundary.

Analysis of Thickness of the Crust-Mantle Transition

In order to examine the character of the reflections from different types of boundaries, we use a Z transform technique (Robinson and Treital, 1980). A scattering matrix was obtained between the waves at the top of a multi-layered medium and the waves at the bottom of that medium. By controlling the two-way travel time in each layer and the total number of layers, we can generate reflections from different types of boundaries.

In this study we treated the reflections as traveling at normal incidence to simplify the computation. A five-layered transitional boundary was used in this study. The velocity is 6.7 km/s at the top, corresponding to an arbitrary P-wave velocity for the lower crust, and 8.2 km/s at the bottom, corresponding to the P-wave velocity in the mantle. The velocity was increased uniformly within in the transitional zone (Figure 8). The source wavelet was modeled after arrivals observed on earthquakes occurring in this area. The dominant frequency is approximately 4 Hz. From Figure 8, the reflection begins to spread out and no longer resembles its original pulse shape when the total thickness of the transitional boundary exceeds one quarter of its wavelength. The dominant frequency of the reflections recorded by the short period seismic stations is 4 Hz, and thus the maximum thickness for a transitional boundary to allow the reflections to preserve their shape is less than 2.0 km for the P wave and 1.0 km for the S wave.

Discussion and Conclusion

From our travel-time data, we computed an average depth to the crust-mantle boundary of 55 km. The range of crustal depths is ± 1.5 km. Another uncertainty in depth is due to the uncertainty in wave velocity in the lower crust. Velocity uncertainty would contribute less than 1 km of uncertainty to depth. The depth variation computed between the two most distant events, event 2 and event 4, is 1.6 km over a 25 km

horizontal distance, indicating that if the boundary is dipping, the dip would be less than 3.6 degrees in the vicinity of CBT.

Figure 9 shows computations for the depth of the crust-mantle transition zone from a time term analysis (Long and Liow, 1985). Because the time term depths were computed with a constant P-wave velocity of 6.15 km/s, the time term depths are shallower and need to be adjusted to compare them with the gradient velocity model used in this reflection study. The difference in the computed depths from these two velocity models is 3.2 km. This implies that, for instance, the 49.5 km depth near the Great Smoky Fault area on Figure 9 will increase from 49.5 km to 52.7 km. The area of the present study (see Figure 9) does not overlap any result from the time term analysis. However, if we assume the depth to the crust-mantle transition zone follows the surface geological structures trending northeast, the depths of the crust-mantle boundary of 49.5 km (52.7 km after correction) and 48.1 km (51.3 km after correction) on Figure 9 are very close to our present result (within the estimates of uncertainty).

The reflections all resemble the character of the direct S wave and, according to our theoretical seismograms, the thickness of a transition zone has to be less than 1.0 km in order to preserve the original shape of the S waves. A 15-km-thick transitional zone for the crust-mantle boundary, as proposed by Owens and others (1984), will distort the shape of the incident wave and generate a longer duration of the reflections, a phenomenon not observed in our data. As to how the velocity structure varies within the transitional zone, whether it increases uniformly or irregularly, we are not able to tell with our "thin" model and the 4 Hz waves we recorded in this study. In order to improve the resolution, recordings of the higher frequency waves would be needed.

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Figure Captions

Figure 1. Relative blast sites and earthquake epicenter to seismic station CBT and other seismic stations used in location. Solid circles are blast sites. Star represents earthquake epicenter.

Figure 2. An example of the blast trace (event 2) on digital three-component record. Moho reflection is on the vertical direction.

Figure 3. The earthquake trace (event 4) on digital three-component record. Moho reflection appears on both N-S and vertical directions.

Figure 4. Particle motion plots of the reflections: event 1 (4a), event 2 (4b), and event 3 (4c). 4d and 4e are the first 0.2 second and the rest portion of the reflection from event .

Figure 5. Comparison of the reflection of event 4 between the vertical component and the horizontal N-S component. The first 0.2 second of the reflection on the vertical component is considered a secondary conversion from the sediment-crust boundary.

Figure 6. Comparisons between the direct shear waves and the Moho reflections.

Figure 7. Crustal model used in computing the depth to the crust-mantle boundary (Propes, 1985).

Figure 8. Synthesized vertical P reflections from a five-layered transitional boundary related to the total thickness of the transitional boundary.

Figure 9. Results of the depth of the Moho discontinuity from the time-term method (Long and Liow, 1985). Shaded square is the approximate area studied in this paper.

Table Captions

Table 1. Hypocenter data for the four events used in this paper.

Table 2. Theoretical and observed travel-time data based on a depth of Moho at 55 km. A ray tracing program (Propes, 1985) is used in computing the theoretical travel time.

Event	1	2	3	4
Date	841030	841114	841121	841217
O-T	16:05:27.6	15:45:39.3	14:34:36.5	18:48:28.2
Lat.	35.564	35.615	35.598	35.390
Long.	84.325	84.400	84.362	84.359
Depth(km)	0	0	0	15.1
M-Res.(s)	0.36	0.44	0.42	0.19
Error of Ellipse	1.07	0.74	0.72	0.12

Table 1.

Event	1	2	3	4
Type	S to P	S to P	S to P	S to S
AZM(deg)	72.4	12.5	39.1	161.3
O-D(km)	9.2	8.6	8.4	17.4
T-D(km)	9.16	8.55	8.43	17.47
O-T(s)	22.8	23.46	23.01	25.3
T-T(s)	23.04	23.02	23.02	25.14
Error(s)	-0.24	+0.44	-0.01	-0.16

Table 2.

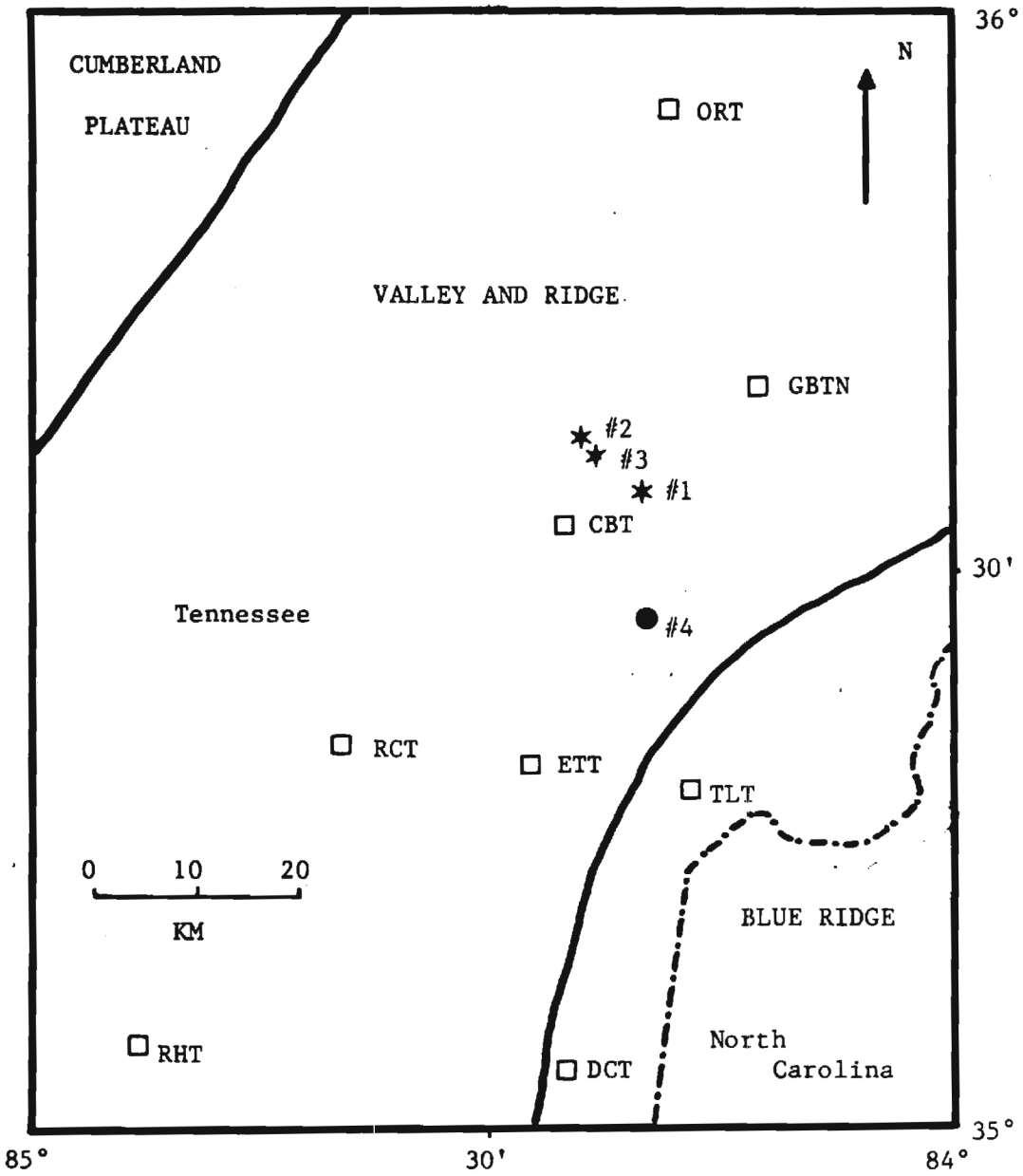

CUMBERLAND
PLATEAU
ORT
N
VALLEY AND RIDGE
GBTN
#2
#3
#1
CBT
Tennessee
#4
RCT
ETT
TLT
0
10
20
KM
BLUE RIDGE
RHT
DCT
North
Carolina
36°
30'
35°
85°
30'
84°

Figure 1.

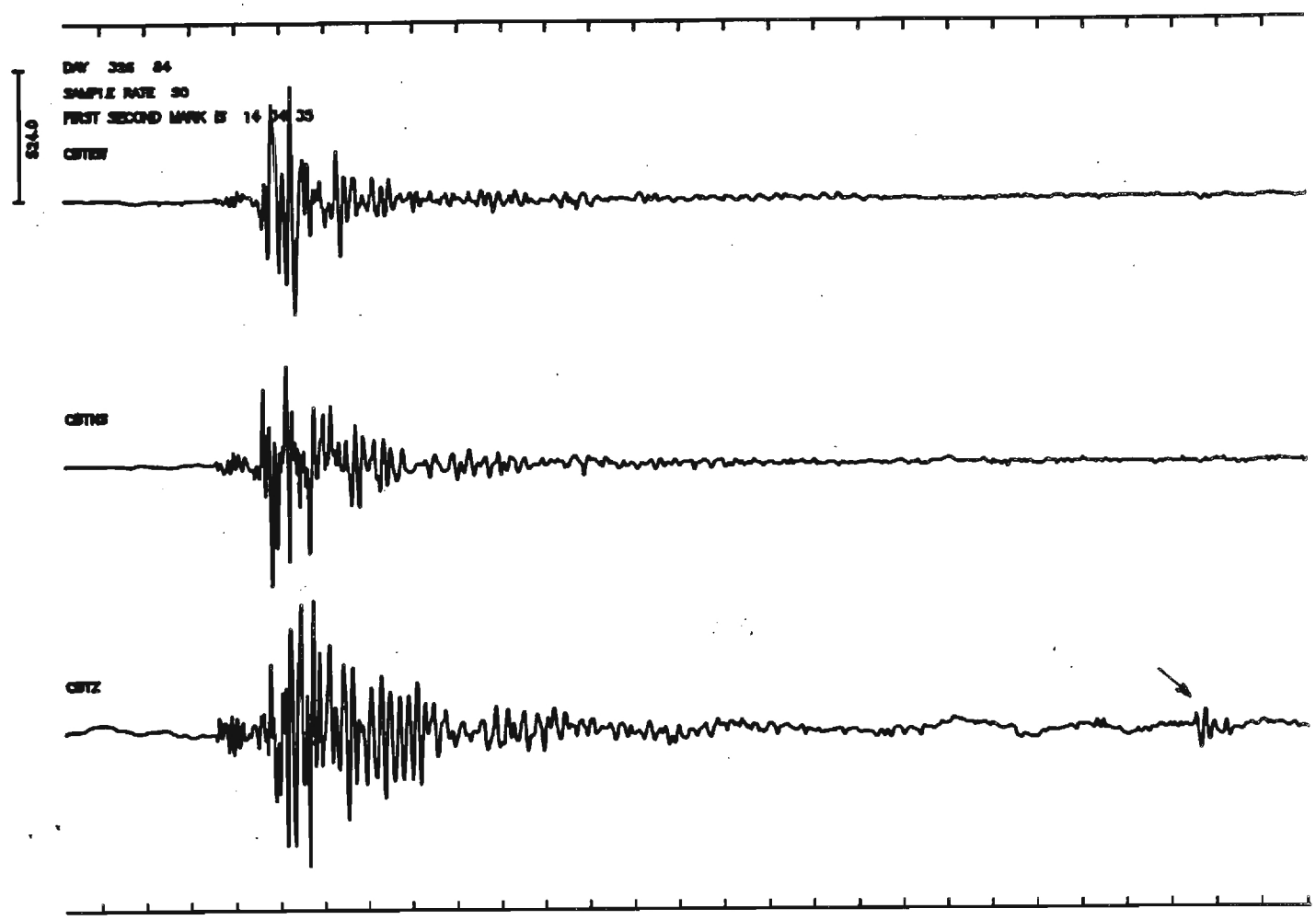


Figure 2.

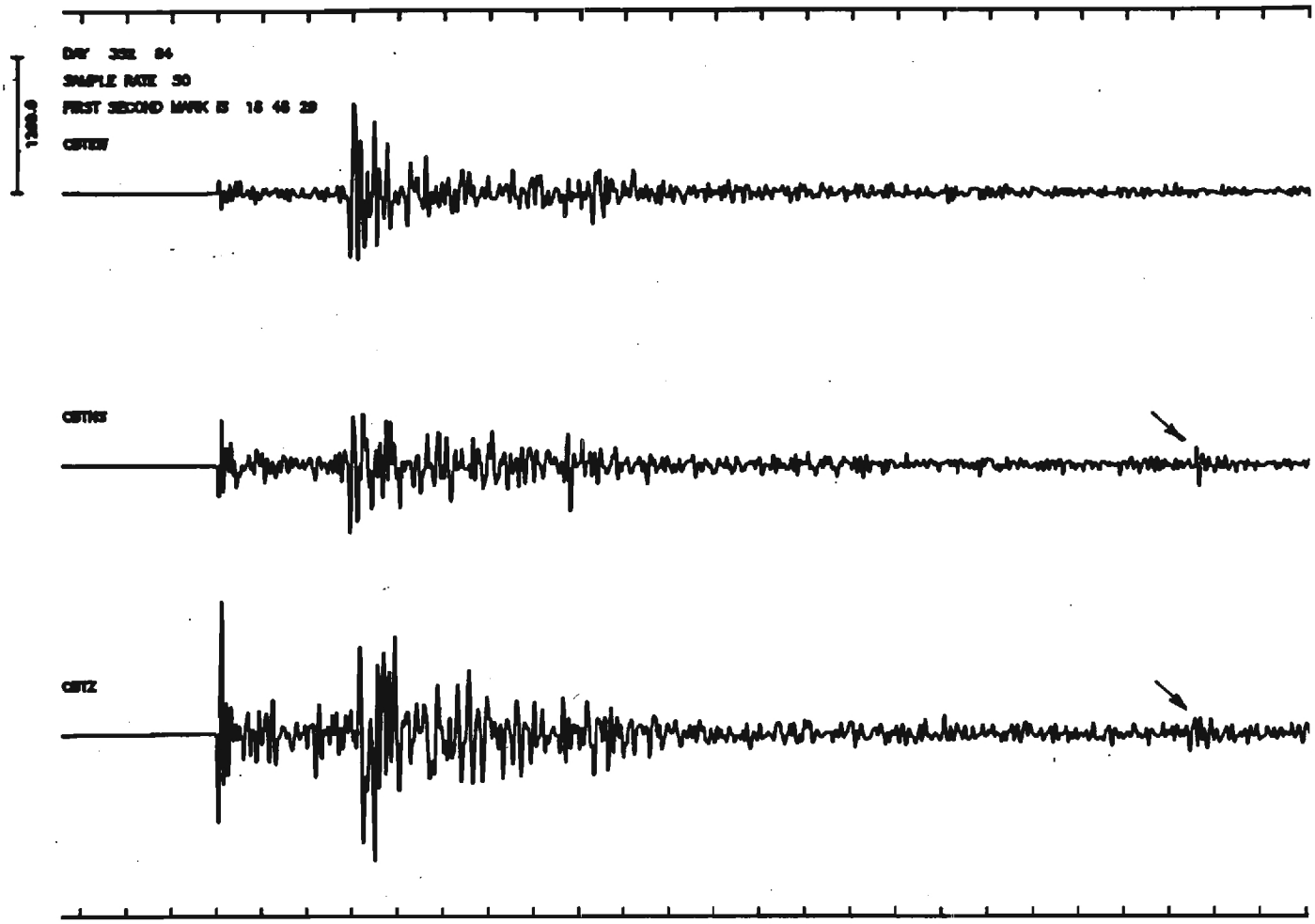
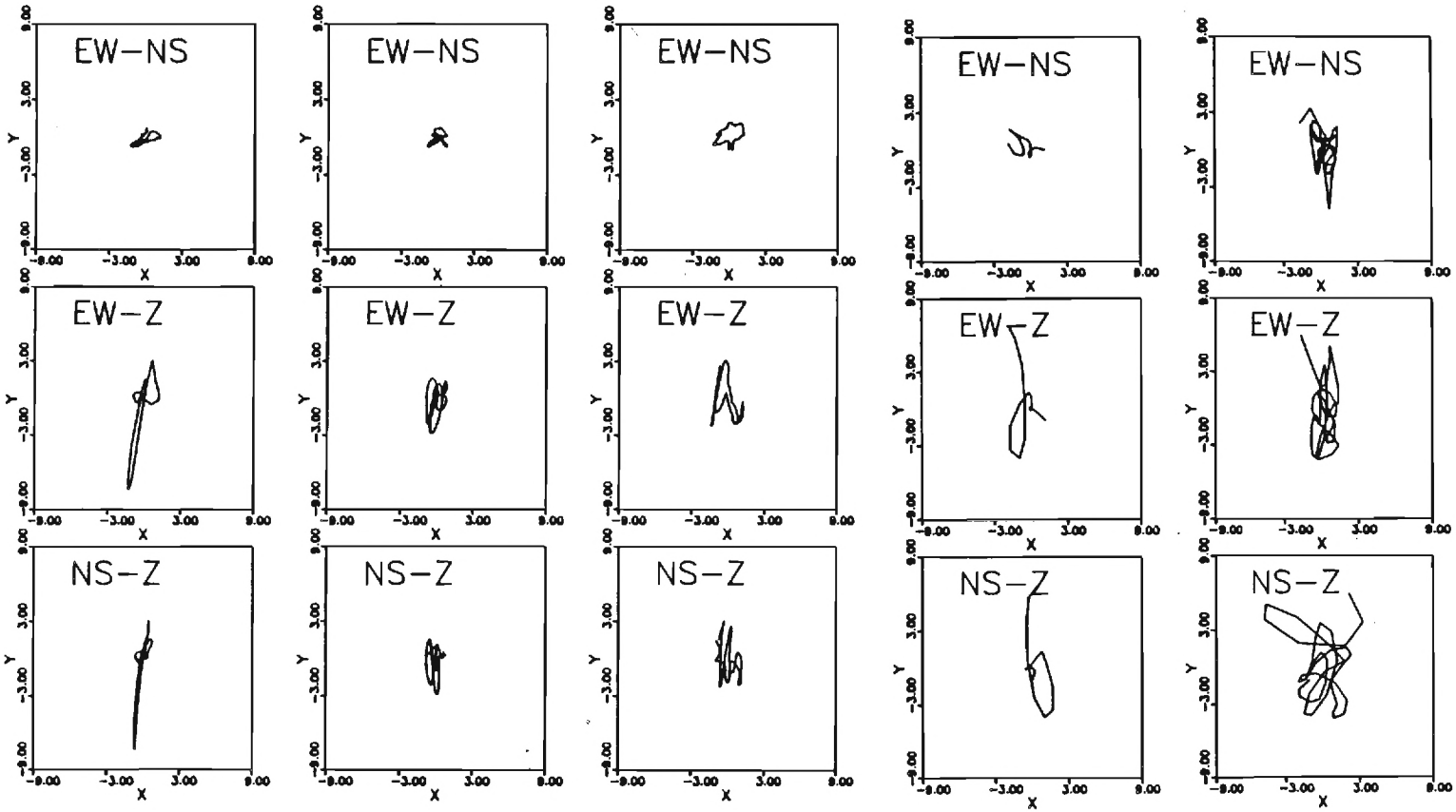

DAY 352 84
SAMPLE RATE 50
FIRST SECOND MARK IS 18 48 29
CBTEW
CBTNS
CBTZ

Figure 3.



4a. 4b. 4c. 4d. 4e.

Figure 4.

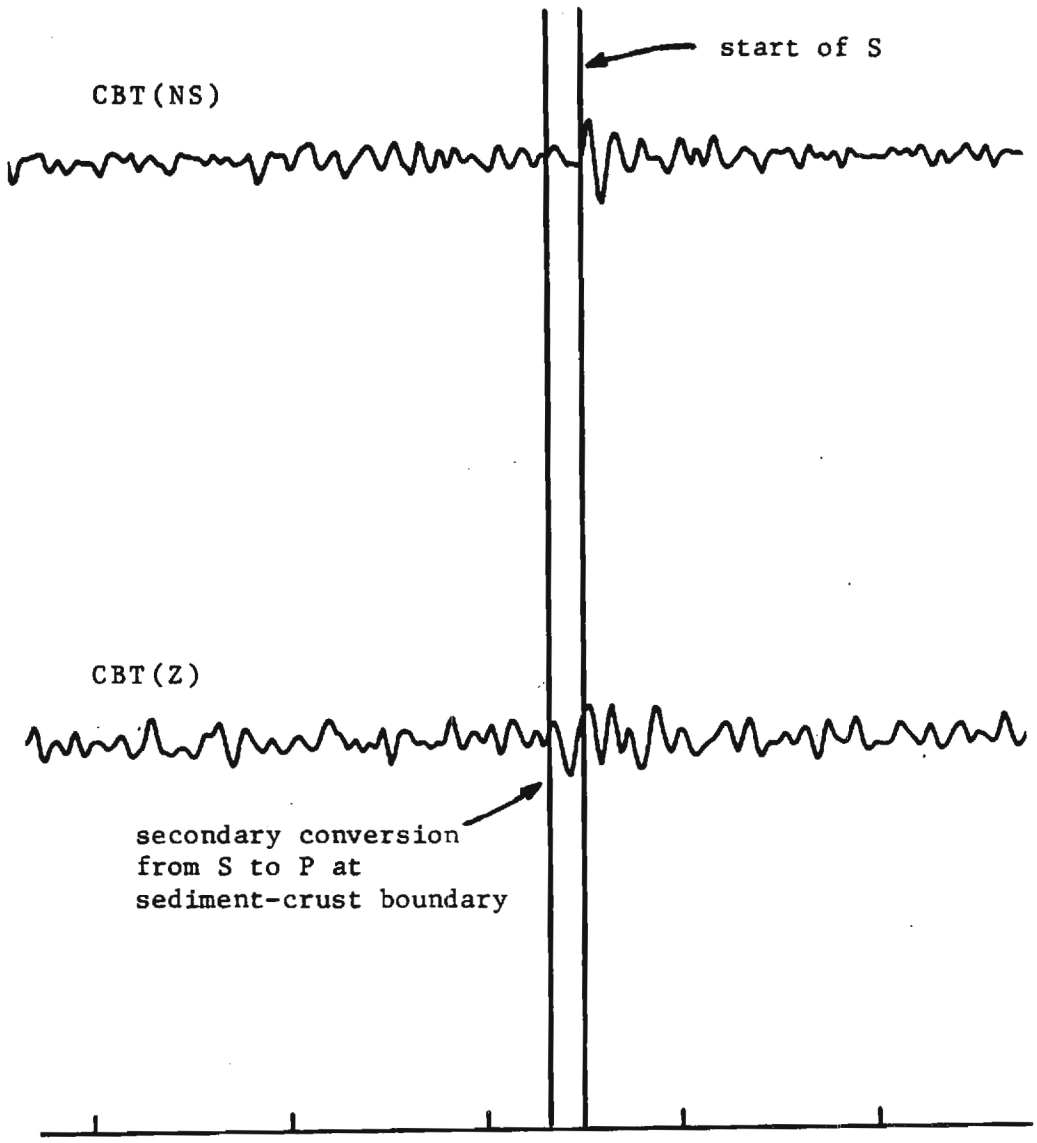

start of S
CBT(NS)
CBT(Z)
secondary conversion
from S to P at
sediment-crust boundary

Figure 5.

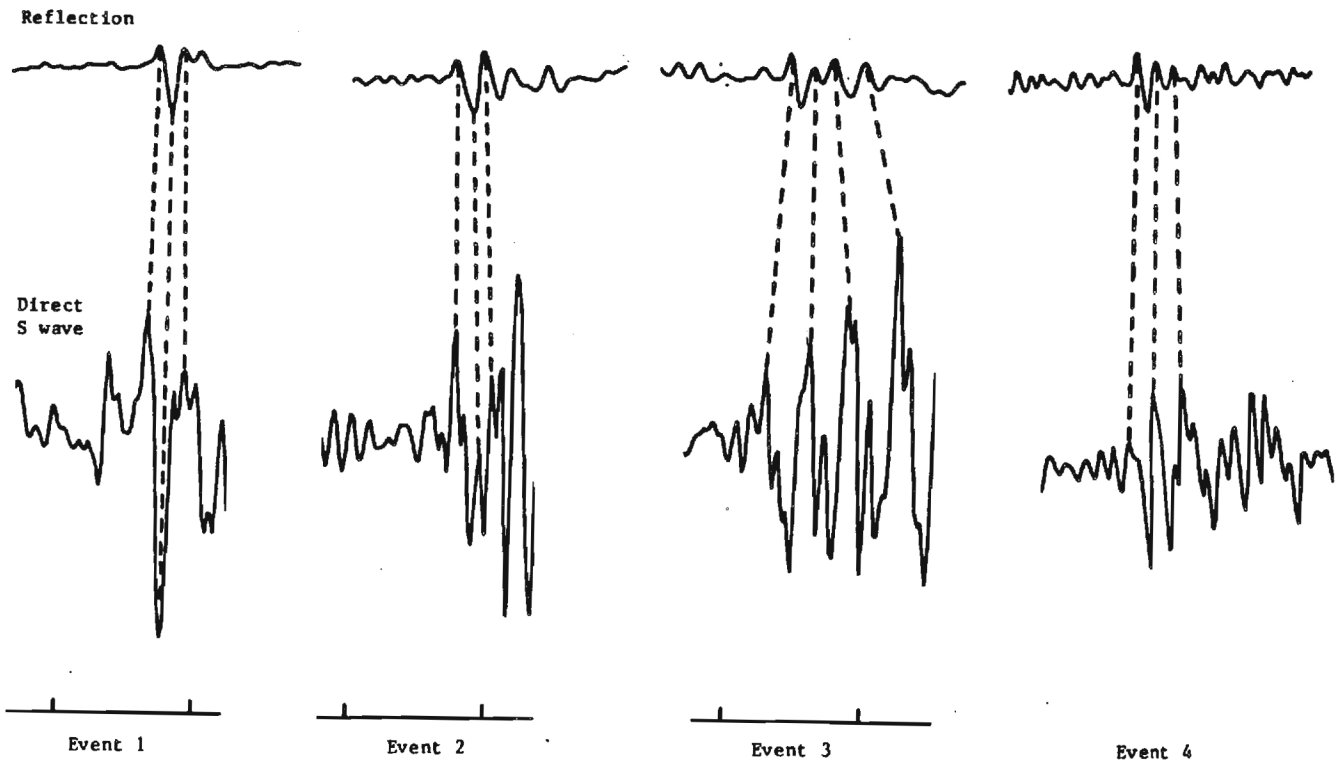


Figure 6.

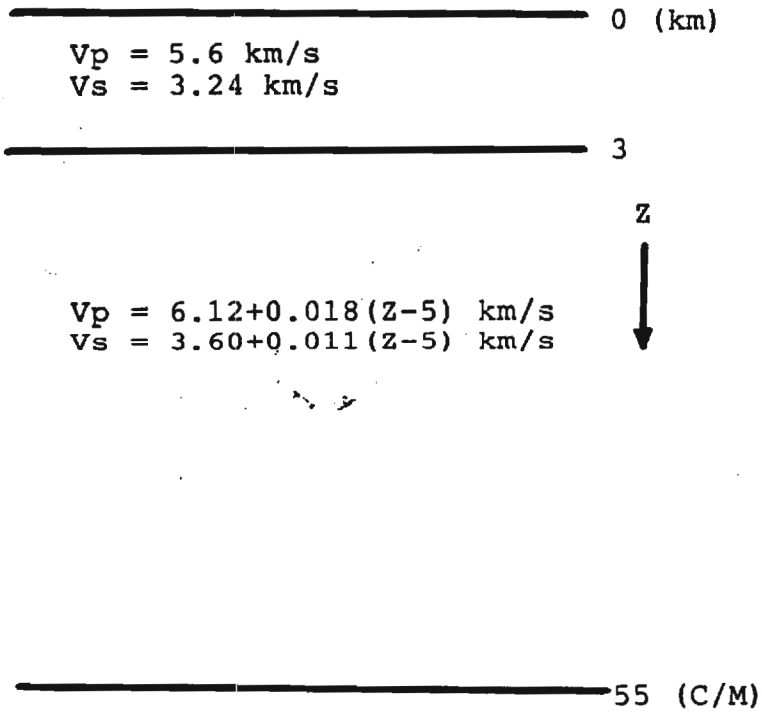

0 (km)
Vp = 5.6 km/s
Vs = 3.24 km/s
3
Z
Vp = 6.12+0.018(Z-5) km/s
Vs = 3.60+0.011(Z-5) km/s
55 (C/M)

Figure 7.

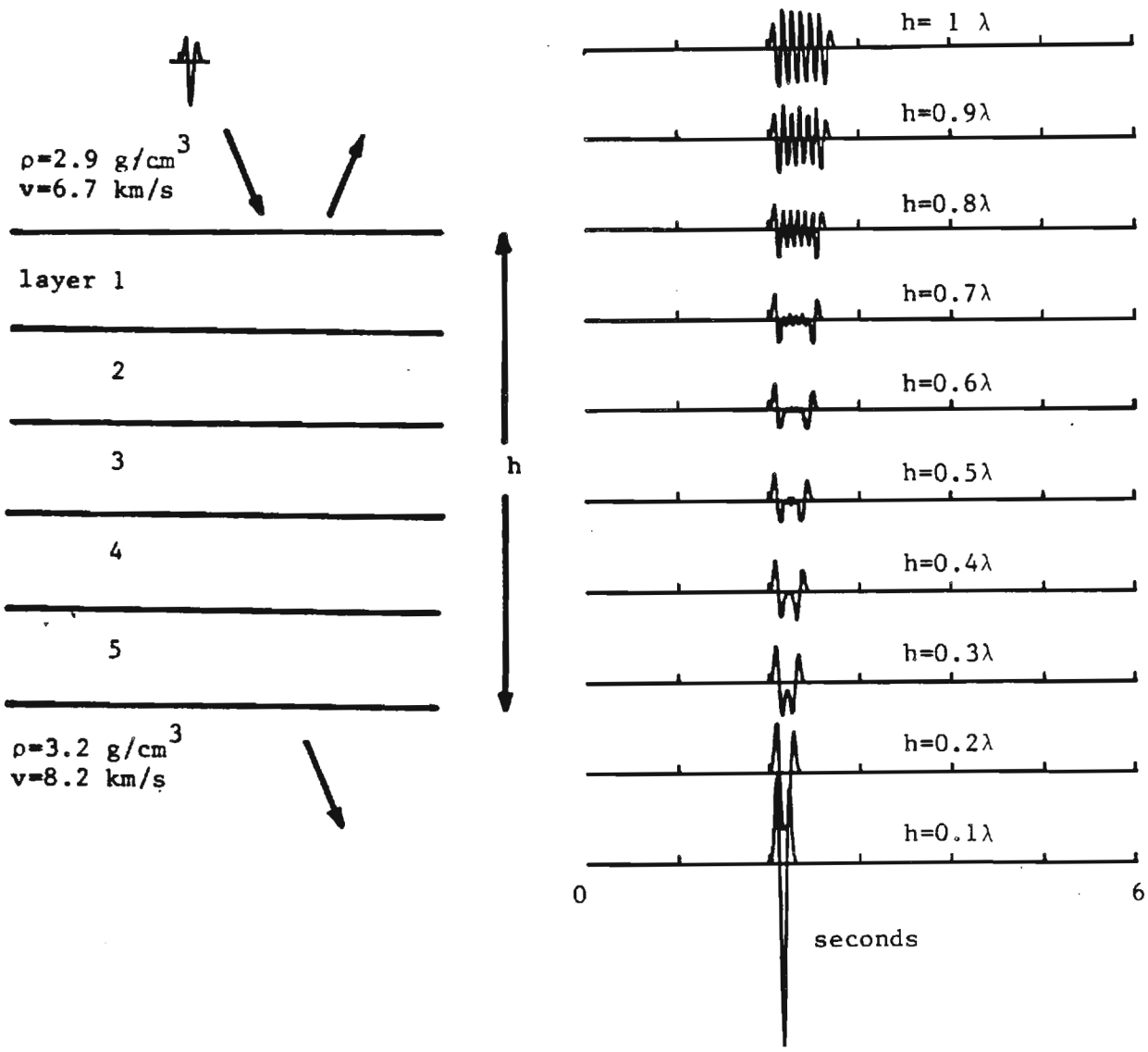

ρ=2.9 g/cm3
v=6.7 km/s
layer 1
2
3
4
5
ρ=3.2 g/cm3
v=8.2 km/s
h
h= 1 λ
h=0.9λ
h=0.8λ
h=0.7λ
h=0.6λ
h=0.5λ
h=0.4λ
h=0.3λ
h=0.2λ
h=0.1λ
0
6
seconds

Figure 8.

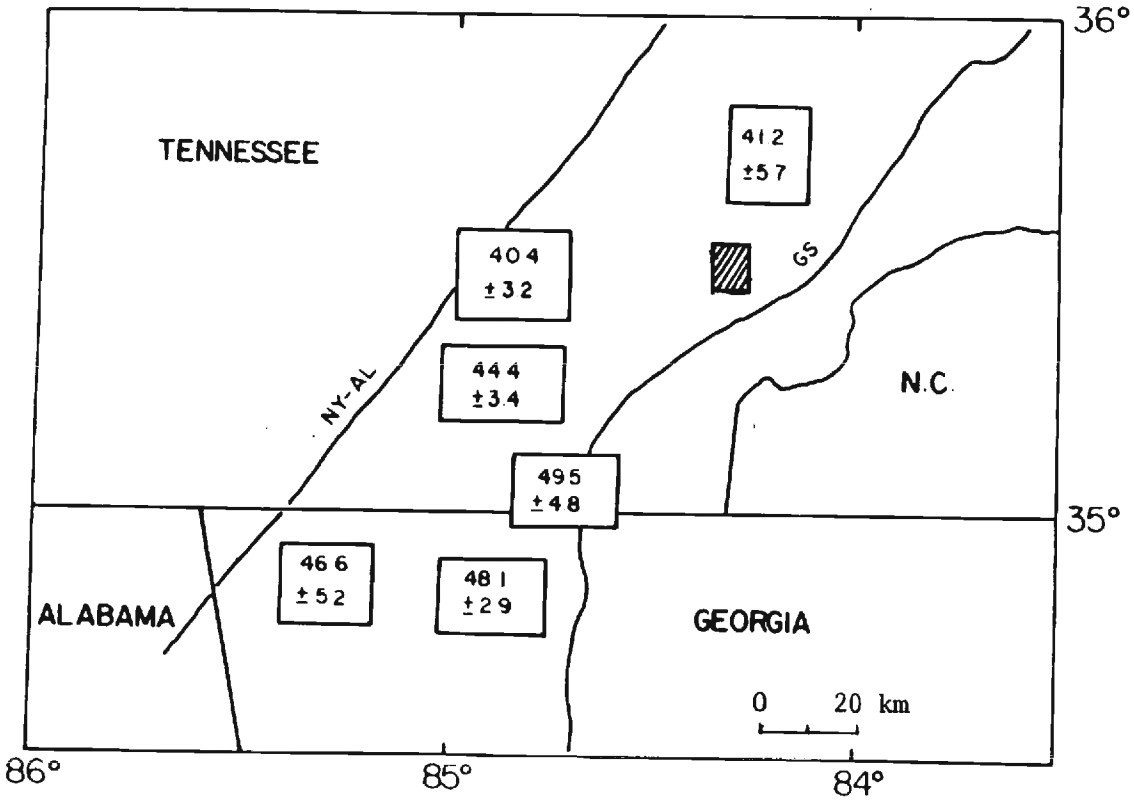

36°
TENNESSEE
412
±57
404
±32
GS
444
±34
NY-AL
N.C.
495
±48
35°
466
±52
481
±29
ALABAMA
GEORGIA
0 20 km
86°
85°
84°

Figure 9.

APPENDIX II

THE NORTH GEORGIA EARTHQUAKE OF OCTOBER 9,1984

By

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Abstract

In the early morning of October 9, 1984, northwestern Georgia was shaken by a magnitude m_b = 3.5 earthquake. This event is typical of the general seismicity experienced in northwestern Georgia and southeastern Tennessee. The maximum intensity was IV and it was felt over 12,000 km^2. The depth of focus was 10 km, which places the earthquake below the Paleozoic sediments. A strike-slip focal mechanism was indicated with right-lateral slip on the north-trending near-vertical plane. The equivalent fault radius was 1.14 km. A stress drop of 75 bars was computed for a moment of 3.5×10^{23} dyne-cm. An aftershock survey detected only one significant (m_b>1.5) aftershock which occurred on October 15, 1984.

Introduction

Early in the morning of October 9, 1984, northwestern Georgia and adjacent areas of Alabama and Tennessee were shaken by a small earthquake of magnitude 3.5. The epicenter was located in the Southern Appalachian Seismic Zone defined by Bollinger (1973). In describing the results of five years of cooperative efforts by southeastern seismic net operators, Johnston et al. (1985) noted that the greatest concentration of seismic activity in the Southern Appalachian Seismic Zone was in southeastern Tennessee. The October 9, 1984, event occurred at the southwestern edge of this concentration of recent activity (Figure 1). In this century, about 10 events were reported felt and were located within about 30 km of the October 9, 1984, event (Tables Ia and Ib). Fewer earthquakes have occurred at scattered locations to the southwest in the extension of the Southern Appalachian Seismic Zone into Alabama (Steigert, 1984).

Intensity

An intensity survey for the October 9, 1984, event was conducted by visiting 39 communities. In addition, the U.S. Geological Survey in Denver kindly shared responses to an intensity questionnaire mailed to postmasters in the region. In all we documented 135 responses. The greatest intensities were reported in the populous Chattanooga area, where walls creaked and windows were broken. The event occurred at 6:55 am local time in Georgia and 5:55 am local time in Alabama. At this hour of the day, the difference in local time could have affected the perception of intensity. The intensity reports from Alabama were more sparse; only a small percent were in Alabama.

The area where the earthquake was felt with intensity IV covers an elliptical area of about 6000 km^2 around the epicenter and stretching farthest to the west (Figure 2). Also, isolated felt areas were found in several directions at distances of 100 to 150 km. One such area occurred 130 km northwest of the epicenter in southern Tennessee. A single contour which encloses all felt regions has an area of 12,000 km^2.

Much of the region which experienced the October 9, 1984, earthquake also felt the February 18, 1964, earthquake. The 1964 earthquake was felt in Chattooga, Dade, and Walker counties in Georgia, and DeKalb county in Alabama. The highest intensity was V(MM) and was observed in Lyerly and Menlo, Georgia. The epicenter of the 1964 event was 20 km west-southwest of the 1984 earthquake.

Summerville, Georgia, is in Chattooga County, Georgia, 30 km south of the epicenters of the 1964 and 1984 earthquakes. A search was made of microfilmed issues of the Summerville News which date back to 1889, and were published weekly. All front pages from 1889 to 1965 were examined, except for a few which were missing. Descriptions were found for earthquakes in 1902 and 1964 that indicate these events were perceived with about the same intensity as the 1984 earthquake.

Aftershock Survey

Within two days, teams from Georgia Tech and the Tennessee Earthquake Information Center were in the field with portable seismic recorders. The Tennessee Valley Authority provided additional instrumentation. Continuous coverage with portable smoked paper recorders was

obtained from a few hours after the main event through October 29, 1984. The recording locations are indicated in Figure 3 and in Table II. In all, only eight events were identified (Table Ic), and only one was large enough to provide significant data. The one large aftershock occurred on October 15, 1984, at 16:56 UTC, six days after the main event.

Location and Focal Mechanism

We used the S-P time from the digital recording of station CBT to fix the origin time in both depth and epicenter computations. The depth and epicenter were then computed independent of each other and the origin time. The epicenter of the main event was 34° 46.51' $\pm$0.8 km, 85° 11.57' $\pm$0.9 km (see Table III for complete station and location data). The epicenter is about 10 km east-northeast of the town of LaFayette. The focal depth from only first motions was 15 km $\pm$10.

Because the depth of focus for the main event was not well constrained by the first arrivals alone, the Pn delay times were also used to constrain the depth. Direct P and Pn arrival times are given in Figure 4. The observed velocity for the direct P wave was 6.3 km/s. A velocity of 6.13 km/s was observed by Dainty et al. (1984) to the southwest. However, Owens et al. (1984) noted that a velocity structure typical of a continental rift is found to the north. The 6.3 km/s velocity we observed is related either to a southward extension of the rift structure observed to the north in Tennessee or to an increase in velocity with depth. The theoretical curves for the P wave from hypocentral depths from 0 to 40 km compared to the observed arrival times only constrain the depth to less than 30 km. However, the Pn delay

times can constrain the depth-of-focus to ± 5 km. The stations in the 180 to 300 km range are located in areas where from previous refraction studies (Dainty *et al.*, 1984; Kean and Long, 1979) the crustal thickness is known to be 35 km. Recent data (Long and Liow, 1985) indicate a crustal thickness of 45 ± 3 km in northwestern Georgia. The delay times indicate a two-way thickness of 80 km, requiring a focal depth of 10 ± 5 km.

The aftershock, for which close-in stations were available, was located 2 km to the southeast (34° 45.23', 85° 10.53') of the main event. The depth of the aftershock was well constrained by the portable smoked-paper recorders. There were two stations within 20 km. The depth was 10.5 ± 0.64. Given the location uncertainties, the aftershock could have occurred within the fault radius of the main event.

The focal mechanism was computed by using first motions and the ratios of P- and S-wave amplitude. The 10 km depth was used to compute take-off angles, although varying the depth from 5 to 40 km had little effect on the focal mechanism solution. The largest gap in station coverage was 58 degrees (see Figure 5). The focal mechanism solution computed from either first motions or amplitude ratios requires strike-slip faulting for the main event. The domains of valid P, T, and B axes (Guinn and Long, 1977; Zelt and Long, 1984) shown in Figure 5 permit either a north-south or east-west trending vertical fault. The north-trending plane would have right-lateral slip. The strike-slip motion along either north-south or east-west planes is consistent with focal mechanisms reported by Johnston *et al.* (1985) for the seismicity in

southeastern Tennessee. In contrast, the aftershock indicated a normal focal mechanism with planes striking N 85° W and dipping 30 degrees.

Spectral Data

The north Georgia earthquake triggered an event recorder on a three component station, CBT, located 100 km northeast of the epicenter. The shear wave saturated on two channels and was not used in the spectral studies (see Figure 6a). The first 2.5 seconds of the P wave were used to determine the spectra. Spectral analyses of the remainder of the P phase show an increase in 4 to 8 Hz energy which was attributed to frequency selective trapping of energy in the surface sediments. The apparent angle of incidence at the surface, 47.1 degrees, was computed from the covariance matrix of the first 2.5 seconds of the three-component data. This angle is consistent with a surface layer velocity greater than 4.26 km/s. The P-wave trace for spectral analysis was then generated by using the three-component data to compute the particle motion in the apparent direction of propagation (Figure 6b). The spectra of the P wave shows a corner at 2 Hz and a high frequency slope proportional to w^{-2}. Following the formulation of Brune (1970) as used in Marion and Long (1980), we computed a seismic moment of 3.5×10^{23}, a fault radius of 1.14 km and a stress drop of 75 bars. In computing moment, the geometrical spreading was assumed to be spherical, whereas a gradient might be closer to cylindrical spreading and could reduce by 50 percent the moment. The moment implies a magnitude of 3.3, which is consistent with the $m_b = 3.5 \pm 0.1$ estimated from duration. The relation used to compute m_b is from Bollinger et al. (1984) and has been calibrated for the southeastern United States.

Summary and Conclusions

The m_b = 3.5 northwestern Georgia earthquake of October 9, 1984, was typical of events which occur about once every year in the Southern Appalachian Seismic Zone in southeastern Tennessee. Its depth of focus places it in the crust below the thrusted Paleozoic sediments, and hence the north-south trend of the focal mechanism is consistent with north-trending structures of the crust to the north and not the northeast-trending Paleozoic thrusts of the sediments.

Acknowledgments

The cooperation of the Tennessee Valley Authority, the Tennessee Earthquake Information Center, and the Virginia Tech Seismic Observatory and other net operators in providing data is greatly appreciated. Field monitoring was assisted by Robert Duckworth and Mitch Craig.

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List of Figures

Table Ia. Historical Events of Northwestern Georgia Occurring in the Area Bounded by 34.5°N, 35.1°N, 85°W, 86°W.

Date	Time	Inten.	Lat.	Long.	Location
1902 May 29	02:30	V	35.1	85.3	Chattanooga, Tennessee
1902 Oct 18	17:00	V	34.47	84.58	Tennessee-Georgia Region
1909 Oct 8	5:	IV-V	34.9	85.0	800 sq mi
1913 Mar 13		IV	34.5	85	Calhoun, Gorden Co. in storm
1927 Jun 16	7:00	V	34.7	86	2500 NE Alabama
1927 Oct 8		V	35	85.3	
1940 Oct 19	00:55	IV	34.7	85.1	
1941 Sep 8	9:45	IV	35	85.3	
1947 Dec 28	00:05	IV	35	85.3	
1964 Feb 18	4:31	V	34.66	85.39	Alabama-Georgia Border

Table Ib. Seismicity in Northwestern Georgia from 1978 through 1984.

Yr/Mo/Da	UTC	Latitude	Longitude	Depth	Mag	#Sta	ELoc	EZ
78/06/16	20:40	34.770	85.040	---	2.3	---	---	---
79/08/14	08:21	34.663	85.318	6.0F	1.6	---	---	---
81/09/04	17:21	34.634	85.166	2.6	3.0	116	0.5	1.1
81/09/28	18:03	34.573	85.435	8.6	2.1	29	0.6	1.1
81/12/13	09:42	34.985	85.143	4.0	0.6	19	1.0	3.2
81/12/23	16:10	34.825	85.813	6.3	1.6	46	1.8	1.4
82/02/23	09:19	34.575	85.445	6.5	2.3	51	0.7	1.5
82/05/12	01:21	34.896	85.020	10.0	2.6	31	0.3	0.6
82/05/20	07:12	35.039	85.148	10.7	1.3	20	0.6	0.6
82/05/26	07:42	34.990	85.265	18.5	2.0	8	0.5	0.5
82/11/23	04:51	35.068	85.446	0.0F	2.0	78	2.9	---
83/01/08	22:30	34.915	85.526	0.0F	2.3	63	2.2	---
83/01/31	23:04	34.962	85.512	0.0F	2.1	68	4.6	---
83/02/11	01:15	35.038	85.006	13	0.7	54	2.4	2.5
83/09/16	09:47	34.811	85.021	21.2	1.5	35	0.8	1.7
83/10/03	02:51	34.810	85.056	6.2	1.2	32	1.5	2.7
83/10/13	10:56	34.879	85.156	17.4	2.0	21	0.5	0.9
84/02/27	17:08	34.662	85.390	10.9	2.1	35	0.3	0.4
84/04/16	07:40	34.696	85.062	8.9	1.5	41	0.6	0.9
84/04/23	06:08	34.857	85.156	19.9	1.2	22	0.8	1.1

Table Ic. Aftershocks of the October 9, 1984, Georgia Earthquake.

Yr/Mo/Da	UTC	Latitude	Longitude	Depth	Dur(s)	#Sta	EH	EZ
84/10/11	01:47:42	----	----	---	10	2	---	---
84/10/11	04:43:47	34.7891	85.1974	3.7	--	4	1.7	3.8
84/10/13	19:30:59.91	34.7488	85.2033	8.0	12	7	1.0	1.4
84/10/15	16:56:52.02	34.7538	85.1754	11.63	80	4	0.5	0.7
84/10/19	05:56	----	----	---	--	1	---	---
84/10/20	11:04	----	----	---	--	1	---	---
84/10/21	02:20	----	----	---	--	1	---	---
84/10/29	08:59	----	----	---	--	1	---	---

Table II. Recording Locations for the Aftershock Survey.

Station		Lat (N)	Long (W)	Elev. (ft)	Gain (dB)
HOGT	Hogg Farm	34.7948	85.2402	890	90
GOGT	Goodman	34.7743	85.2123	1000	72
RPGT	Round Pond	34.7308	85.2275	1050	84
MLGT	Maple Lake	34.7105	85.2592	1070	78
TRGT	Taylor Ridge	34.7233	85.1740	1000	84
GFGT	Gracy Farm	34.7943	85.1755	880	90
ST1		34.7227	85.1431		
ST2		34.7264	85.1653		
ST3		34.7011	85.1819		
ST4		34.7489	85.1708		

Table III. Arrival Time Data for the October 9, 1984, Earthquake.

Sta	Phase	Time	Phase	Time	Dur (s)	Mag	Takeoff Angle	SV/P	Dist (km)	Azim (deg)
CBT	PD	11:54:45.20	-	----------	305	3.63	93.22	---	113	38
CH6	PC	11:55:09.00	S	11:55:39.5	225	3.25	87.94	3.2	260	111
MLA	PD	----------	S	11:55:56.0	277	3.51	86.55	---	322	226
OCA	PD	11:54:45.20	S	11:54:59.6	286	3.55	93.22	---	115	263
BKA	PD	----------	S	11:55:45.5	290	3.57	87.34	2.3	288	245
TSA	PD	11:54:04.9	-	----------	281	3.53	-----	---	---	---
HVA	PD	11:54:53.40	-	11:55:13.2	275	3.50	92.09	1.8	165	242
TDA	PC	11:54:51.9	-	----------	315	3.67	92.34	---	153	211
HGA	PC	11:54:40.0	-	----------	260	3.43	94.72	---	80	229
LMTN	PC	11:54:59.20	S	11:55:24.33	---	----	-----	---	207	51
ONTN	PC	11:54:59.57	S	11:55:24.1	---	----	89.85	---	205	19
SSKY	PC	11:55:03.79	-	----------	---	----	-----	---	235	347
COTN	PC	11:54:58.22	-	----------	---	----	-----	---	201	306
ETT	PD	----------	S	11:54:53.4	---	----	93.95	---	93	45
MSAL	PC	11:54:48.75	-	----------	---	----	92.64	---	137	275
BHT	--	----------	S	11:55:01.8	---	----	-----	---	125	10
SWTN	PC	11:54:43.57	-	----------	---	----	93.64	---	103	308
BBI	PD	----------	-	11:55:02.4	---	----	92.9	1.5	127	83
GBTN	PD	11:54:48.93	-	11:55:04.4	---	----	92.64	1.7	136	41
DRT	PD	11:54:51.73	-	11:55:09.75	---	----	92.34	---	154	31
TKL	PD	11:54:53.28	-	11:55:11.8	---	----	92.22	1.9	165	51
CSPT	--	11:54:53.90	-	----------	---	----	-----	---	169	47

(Continued on next page)

Table III. (Continued)

Sta	Phase	Time	Phase	Time	Dur (s)	Mag	Takeoff Angle	SV/P	Dist (km)	Azim (deg)
CPO	PC	11:54:43.00	-	----------	---	----	93.64	---	102	340
PWLA	PC	11:55:07.84	-	----------	---	----	92.6	1.6	265	276
LCAL	PD	11:54:35.07	-	----------	---	----	98.55	---	47	240
RICH	PC	11:55:06.4	-	----------	---	----	92.22	1.8	253	58
ATL	PC	----------	-	----------	---	----	91.95	---	164	147
BRCC	PD	----------	-	----------	---	----	91.22	1.8	285	71
HPK	P	11:54:55.8	S	11:55:16.7	---	----	91.95	2.7	182	40
BCKY	PC	11:55:09.22	-	----------	---	----	-----	---	---	---

Other arrival data can be found in the Southeast Net Operator's Bulletin.

$\bar{m}_b$(Dur) = 3.5 $\pm$ 0.1

34.7752 $\pm$ 0.8 km

85.1929 $\pm$ 0.9 km

14.93 $\pm$10.5 km

T_o = 11:54:26.25

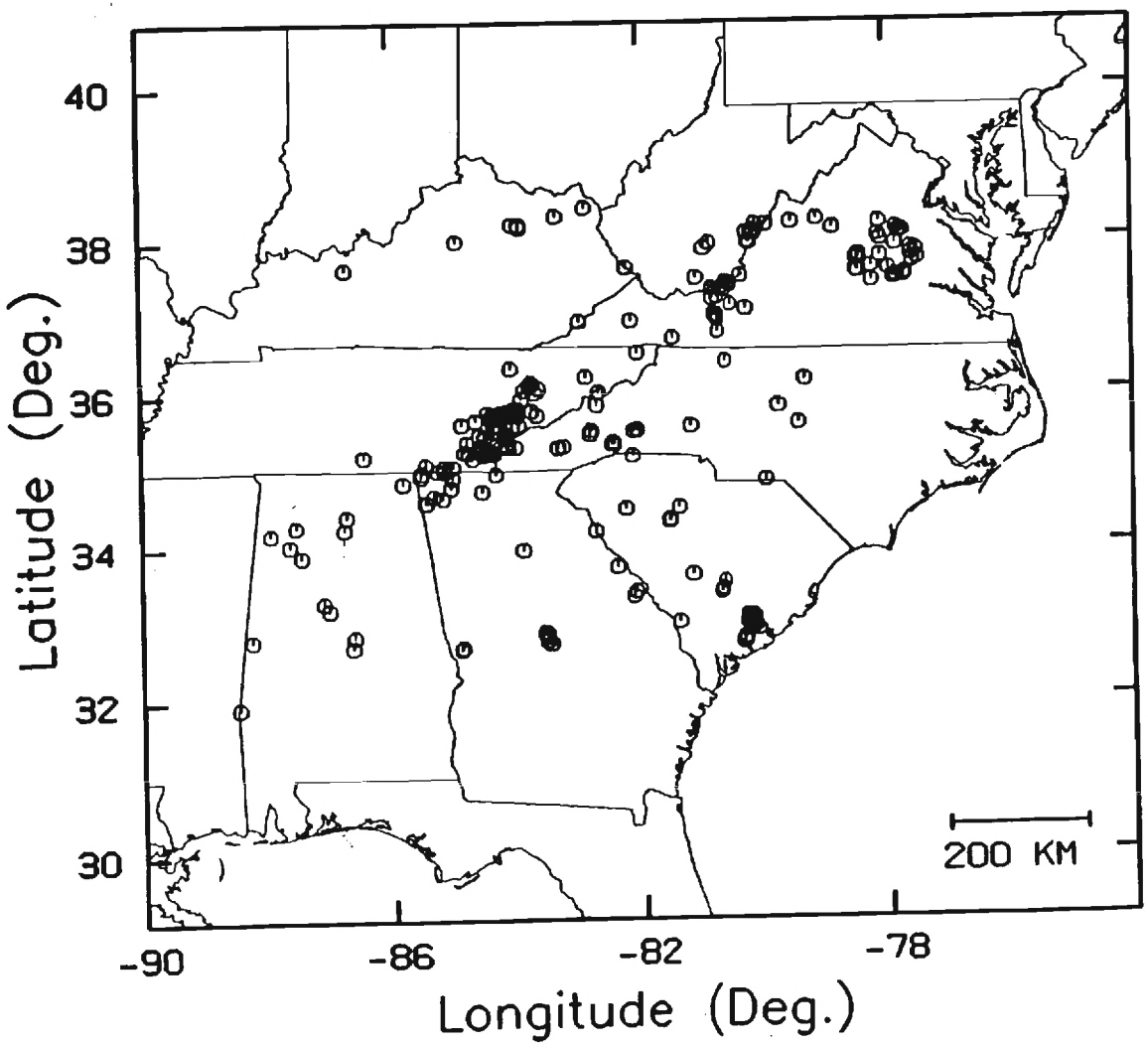


Figure 1.

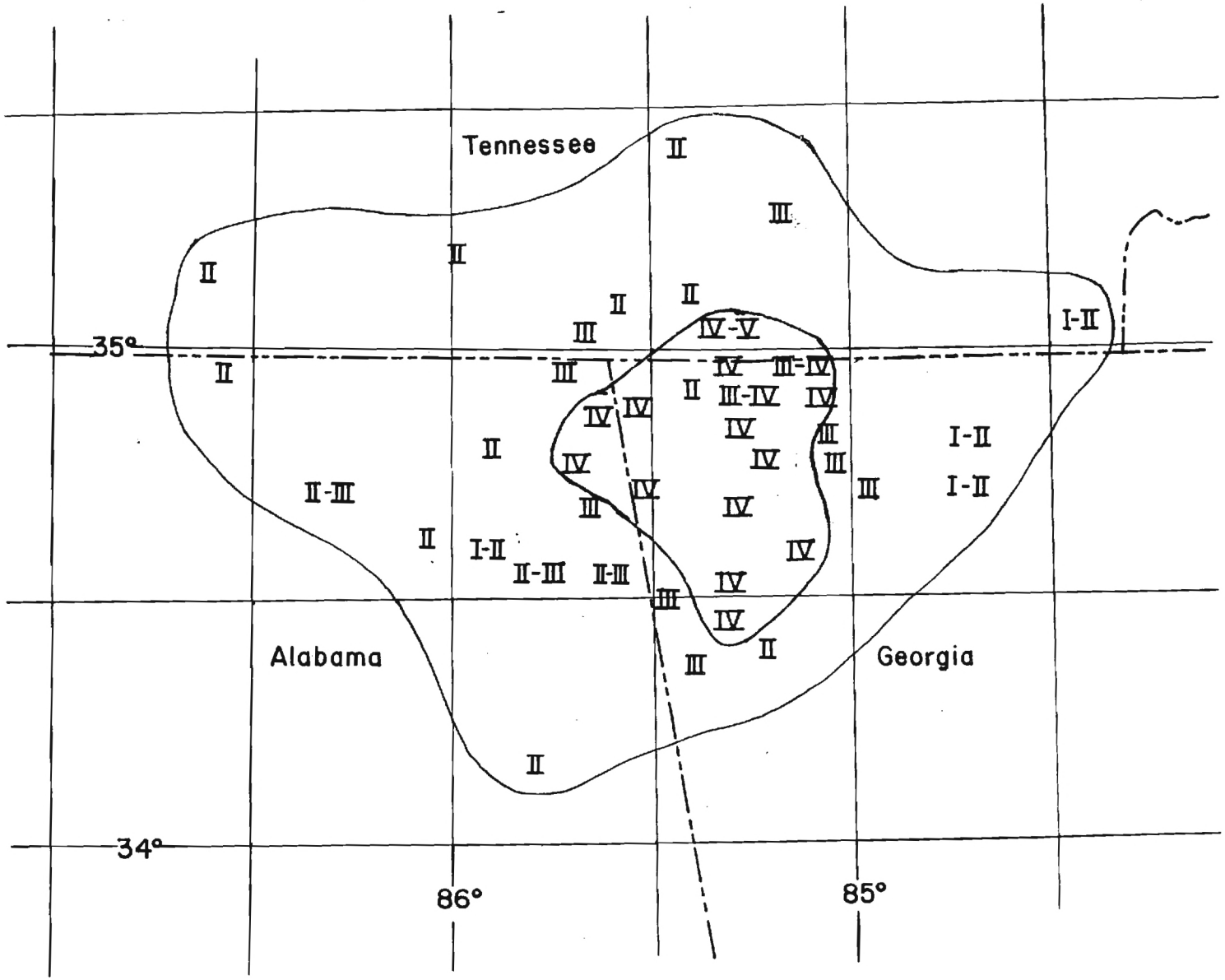

Tennessee
Alabama
Georgia
35°
34°
86°
85°
II
III
IV
IV-V
III-IV
I-II
II-III

Figure 2.

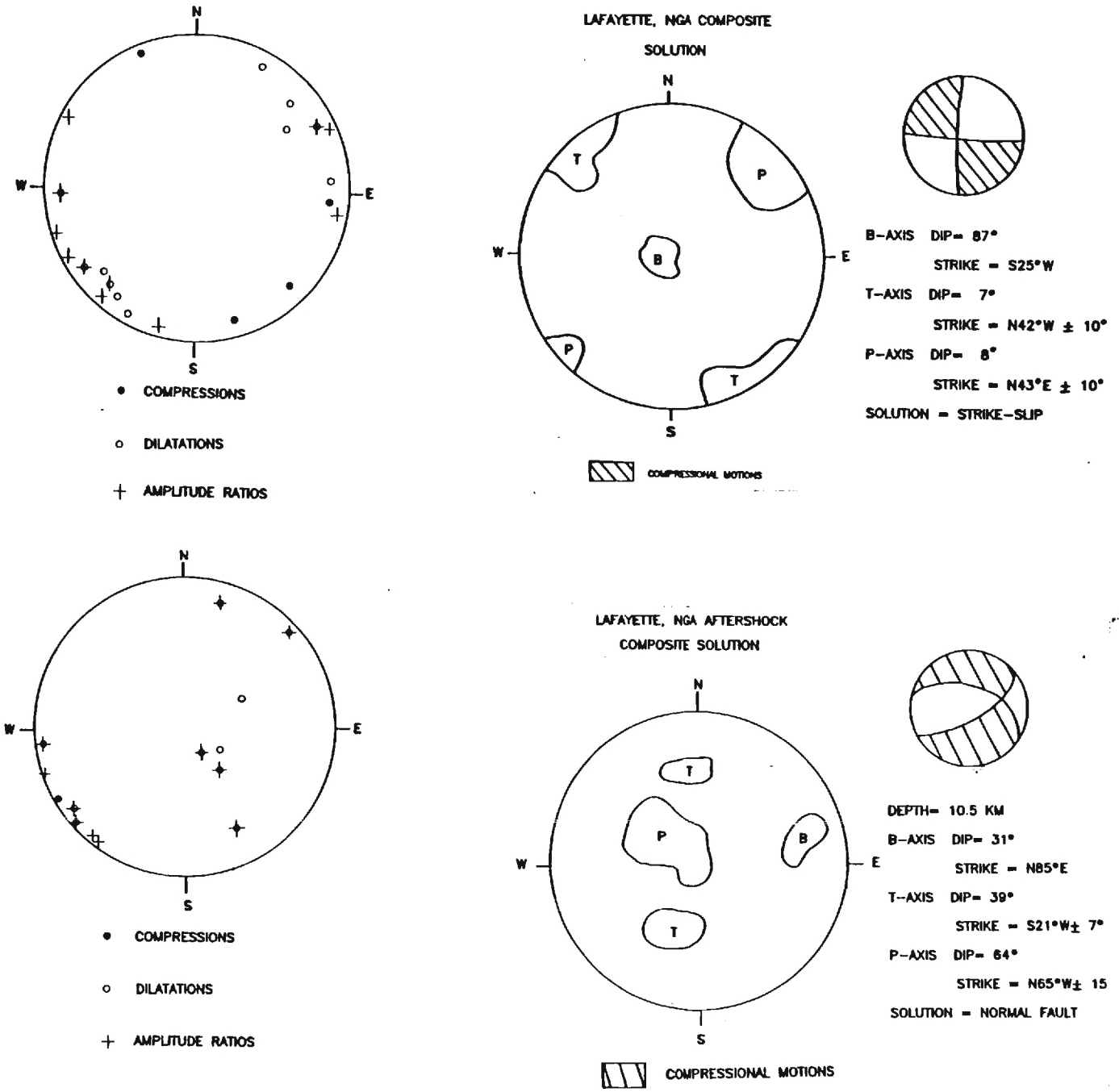

N
W
E
S
COMPRESSIONS
DILATATIONS
AMPLITUDE RATIOS
LAFAYETTE, NGA COMPOSITE
SOLUTION
T
P
B
COMPRESSIONAL MOTIONS
B-AXIS DIP= 87°
STRIKE = S25°W
T-AXIS DIP= 7°
STRIKE = N42°W ± 10°
P-AXIS DIP= 8°
STRIKE = N43°E ± 10°
SOLUTION = STRIKE-SLIP
LAFAYETTE, NGA AFTERSHOCK
COMPOSITE SOLUTION
DEPTH= 10.5 KM
B-AXIS DIP= 31°
STRIKE = N85°E
T-AXIS DIP= 39°
STRIKE = S21°W± 7°
P-AXIS DIP= 64°
STRIKE = N65°W± 15
SOLUTION = NORMAL FAULT

Figure 3.

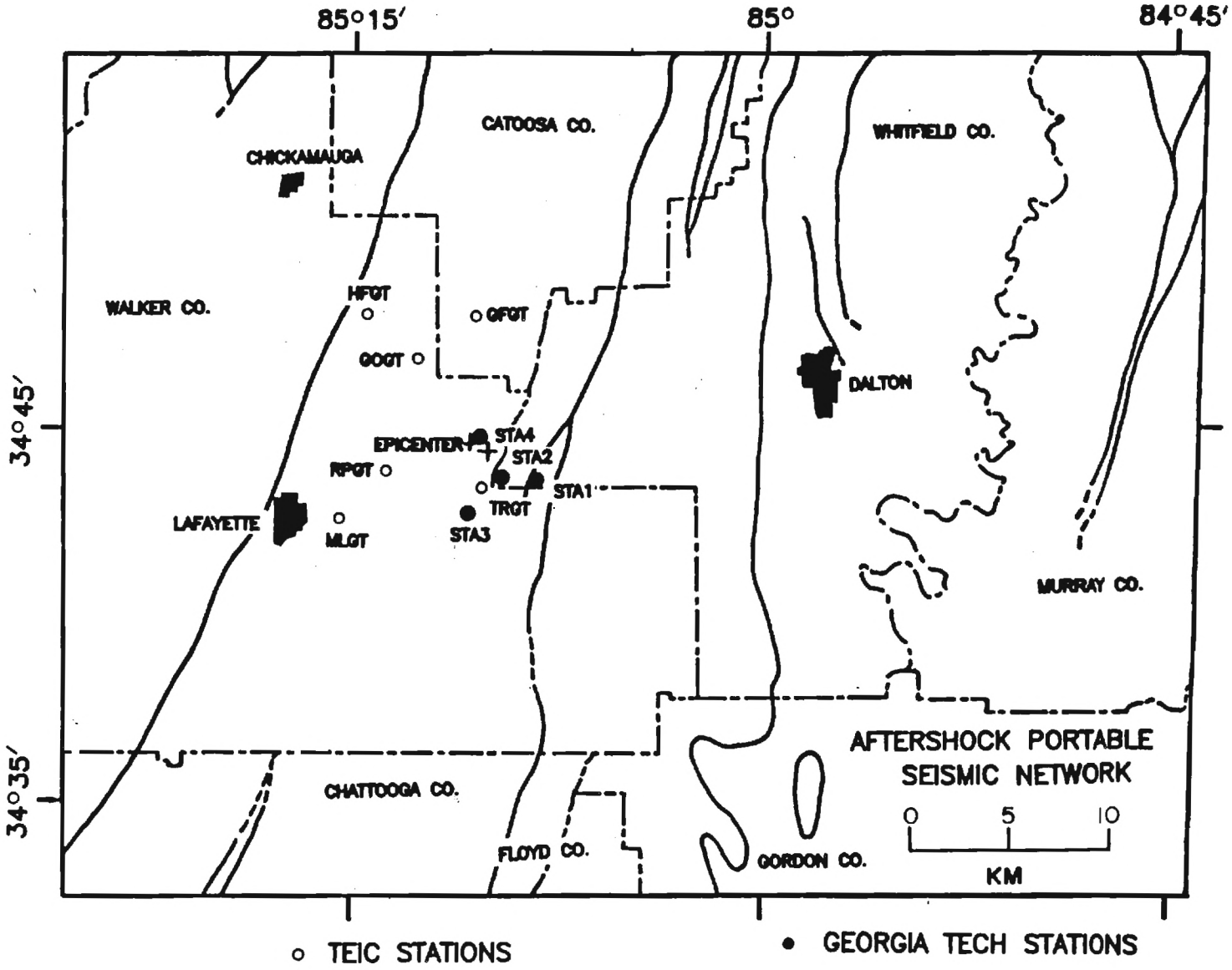

85°15′
85°
84°45′
34°45′
34°35′
CATOOSA CO.
WHITFIELD CO.
CHICKAMAUGA
WALKER CO.
HFGT
GFGT
GOGT
DALTON
EPICENTER
STA4
STA2
RPGT
STA1
TRGT
LAFAYETTE
MLGT
STA3
MURRAY CO.
AFTERSHOCK PORTABLE
SEISMIC NETWORK
0 5 10
KM
CHATTOOGA CO.
FLOYD CO.
GORDON CO.
TEIC STATIONS
GEORGIA TECH STATIONS

Figure 4.

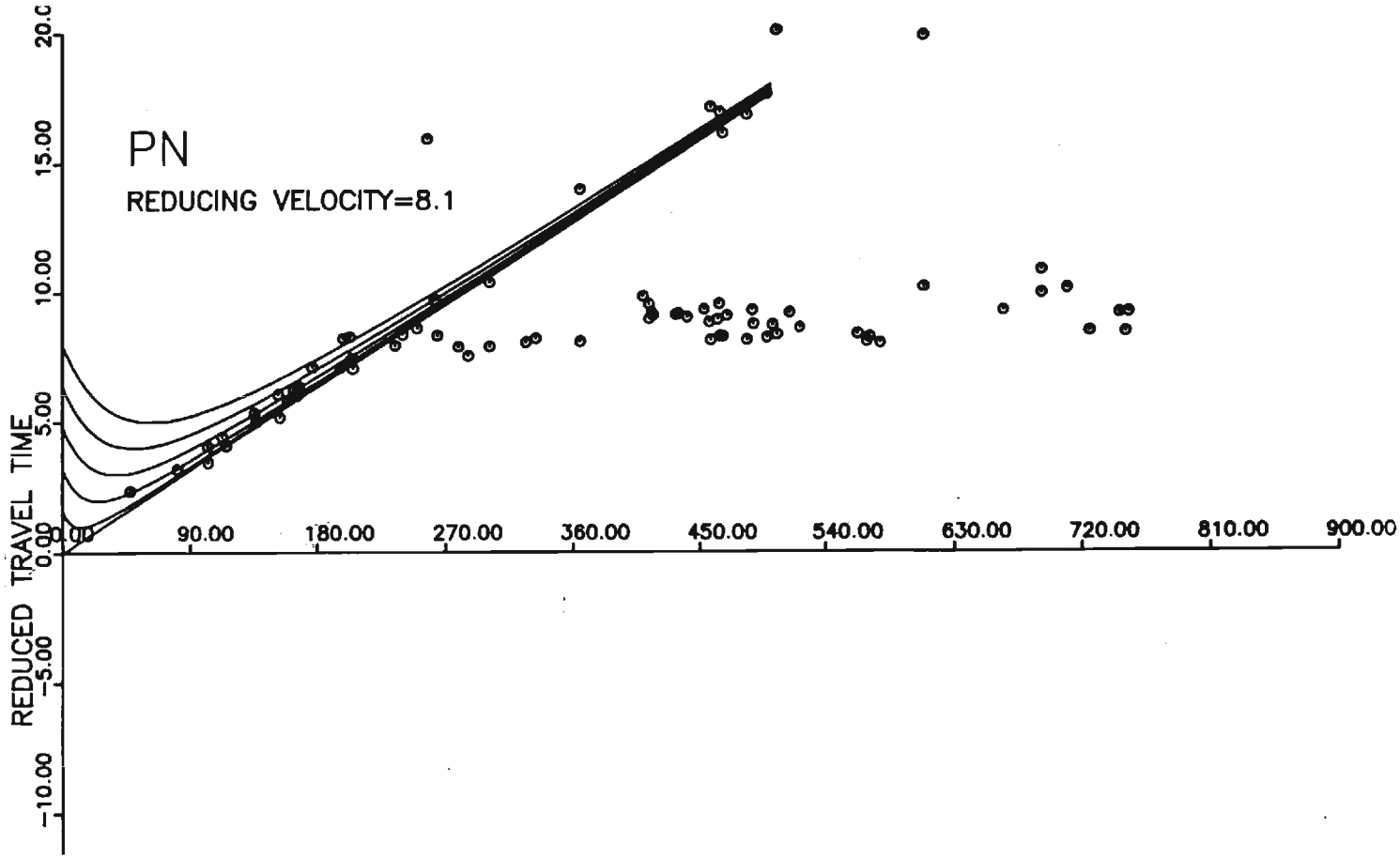
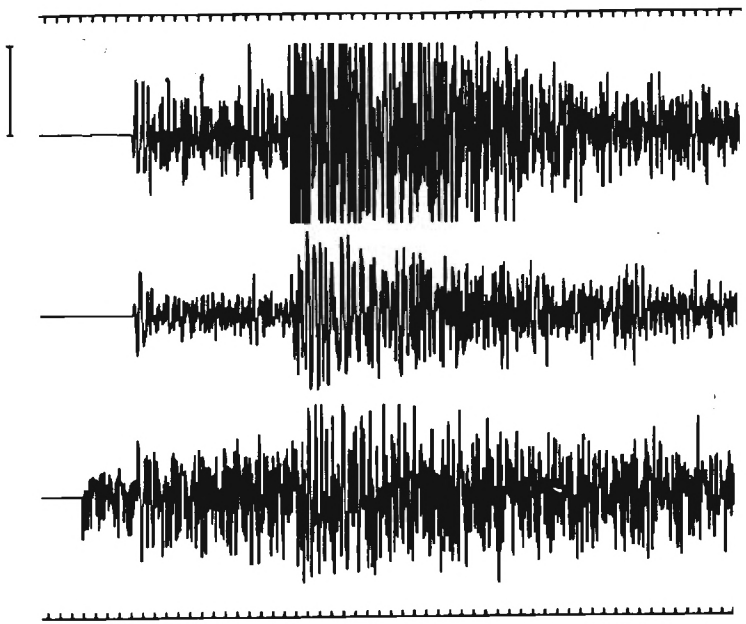

PN
REDUCING VELOCITY=8.1
REDUCED TRAVEL TIME
20.0
15.00
10.00
5.00
0.00
-5.00
-10.00
0.00
90.00
180.00
270.00
360.00
450.00
540.00
630.00
720.00
810.00
900.00

Figure 5.

CBT 3-COMPONENT STATION DIGITAL RECORDS



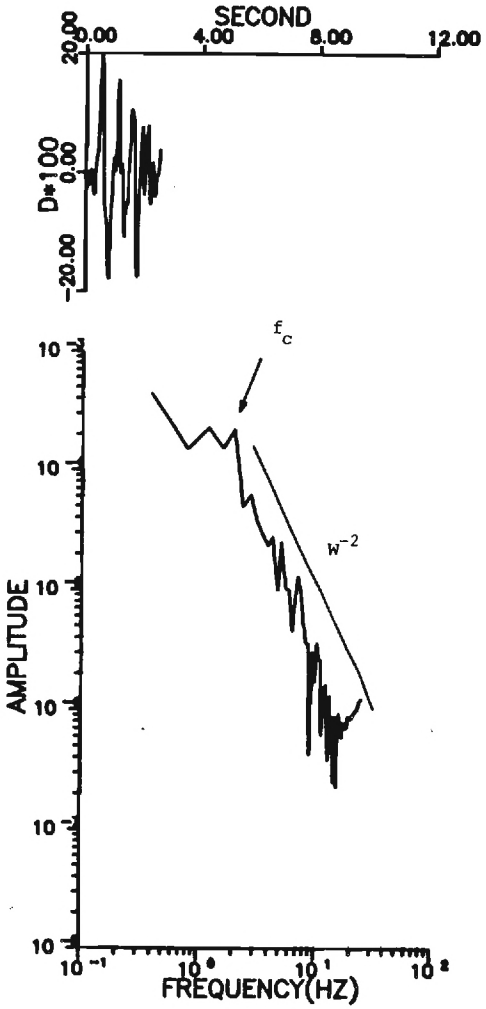


Figure 6.