Kansas refraction profiles

by
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Abstract

Historically, refraction surveys have been conducted in hopes of mapping distinct layers within the earth. Refraction is a useful tool provided its limitations and the assumption that layers increase in seismic velocity with increasing depth are kept in mind. A traditional reversed-refraction profile was conducted along a 500-km (300-mi)-long east-west line extending from Concordia, Kansas, to Agate, Colorado. Analysis of the data showed an average crustal velocity of 6.1 km/sec (3.7 mi/sec) and an average upper-mantle P_n phase velocity of 8.29 km/sec (4.97 mi/sec) with a Moho depth calculated to be 36 km (23 mi) on the eastern end and 46 km (29 mi) on the western end. Some evidence suggests velocities as high as 7.2 km/sec (4.3 mi/sec) in the crust at various locations along the survey line. The strong east-west regional gravity gradient of -0.275 mgal/km supports the seismically drawn conclusion of a thinning of crust in north-central Kansas. In order to supplement the data from this refraction survey, we took advantage of the Kansas earthquake seismograph network. A crustal study using earthquakes as energy sources and a regional earthquake network as seismometer locations resulted in a crustal-velocity model that will improve determination of local earthquake locations. A large anomalous body in the upper mantle/lower crust, assumed to be related to the Precambrian-aged Midcontinent Geophysical Anomaly (MGA), resulted in early P-wave arrivals from refracted energy from the Moho recorded at Concordia, Salina, Tuttle Creek, and Milford. An omnidirectional positive P residual zone near El Dorado may be related to the Wichita geomagnetic low. Some evidence suggests the presence of a lower velocity material on the western and eastern flanks of the MGA, possibly representing the Rice Formation. A P_n velocity of 8.25 km/sec±0.15 km/sec (4.95 mi/sec±0.09 mi/sec) with the crust thinning from west to east and an apparent thinning from the north and from the south was determined from the 16 regional earthquakes studied. Crustal thickness from central Kansas through western Missouri seems to be relatively consistent.

Introduction

For regional geologic, geophysical, and tectonic studies, having some idea of gross characteristics of the deep crust is useful. Knowledge of the deep crust and upper mantle necessarily comes from indirect observations. Since the deepest borehole in Kansas extends only to about 3.5 km (2.1 mi), geophysical observations made at or near the earth's surface must be used to infer geologic conditions at greater depths.

Methods used to model deep crustal structure in Kansas include seismic reflection (Brown et al., 1983; Serpa et al., 1984; Serpa et al., this volume), seismic refraction (Steeples, 1976), gravity and magnetics (Yarger 1983; Yarger et al., 1980), and studies using earthquake waves (Hahn, 1980; Hahn and Steeples, 1980; Steeples, 1982; Lui, 1980; Miller, 1983). More direct information was obtained by examination of deep-crustal and upper-mantle xenoliths in kimberlites from Riley County by Brookins and Meyer (1974).

This paper is a presentation of more complete results of work by Steeples (1976) and an abbreviation of an M.S. thesis by Miller (1983). This paper is likely to be most useful to those interested in earthquakes and tectonics of the midcontinent and to those interested in existing and planned deep seismic-reflection surveys in Kansas.

Geologic setting

The area under investigation is contained within the Central Stable Region (Snyder, 1968, fig. 1). The generalized continental crust is made up of a silicic upper layer covering a mafic lower layer (Pakiser and Zeitz, 1965). The stable environment of the midcontinent shows a relatively thick crust (approximately 50 km [30 mi]) with lower crustal P-wave velocities about 6.7 km/sec (4 mi/sec) generally representative of a gabbro. The upper mantle rocks show P-wave velocities in excess of 8.0 km/sec (4.8 mi/sec) with relatively high density (Pakiser, 1963).

The region under study contains many major upper crustal structural features (fig. 1), whose origins are not clearly understood. It has been proposed that during the late Precambrian, an extensive series of geologic events occurred

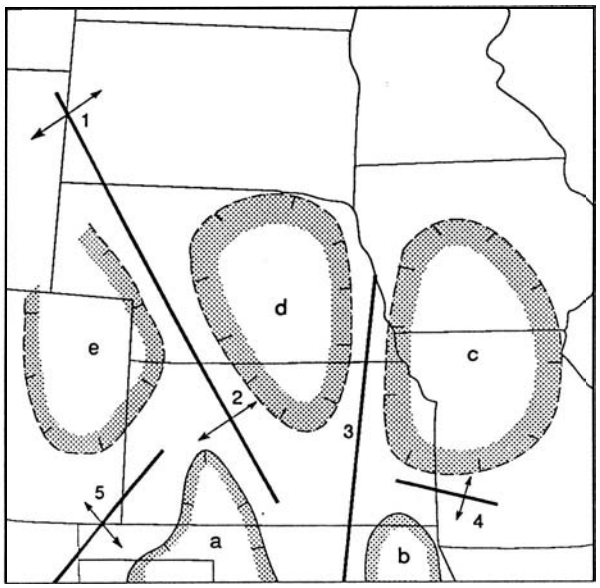


FIGURE 1—MAJOR STRUCTURAL FEATURES WITHIN THE STUDY AREA AS DESCRIBED BY SNYDER (1968).

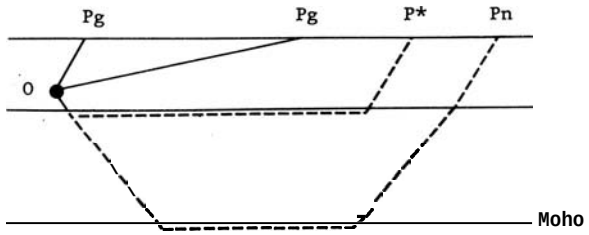


FIGURE 2—PATHS OF MAJOR COMPRESSIONAL WAVES THROUGH THE CRUST.

in the midcontinent (Snyder, 1968). First, long periods of igneous activity over the entire midcontinent resulted in the formation of the Keweenawan basin. This was followed by more igneous activity and the development of a major lineament (Midcontinent Geophysical Anomaly) dividing the midcontinent. Subsidence, uplift, and erosion continued throughout the Paleozoic with the gradual formation of the basins and arches.

This sequence of events was, in part, responsible for the major structural features of the Precambrian basement distinguished by drill data (Cole, 1976). The Precambrian surface in northern Kansas is composed of a wide variety of igneous and metamorphic rocks, primarily of granitic affinity, except near the center of the MGA as discussed below (Bickford et al., 1979).

The MGA is the major Precambrian-aged feature from a geophysical viewpoint in the Central Stable Region (CSR). We prefer the MGA terminology as opposed to the equivalent Central North American rift system (CNARS) or Midcontinent rift because the name MGA is descriptive and not interpretive. The MGA extends laterally from Lake Superior (where the rocks that cause it crop out) through Kansas and possibly into Oklahoma (Yarger, 1983). It represents a 60-mgal gravity high flanked by 100-mgal gravity lows dissected in several places by possible transform faults (Chase and Gilmer, 1973). In Kansas, the MGA is a structure with suggested upper-mantle presence in the form of low P-wave velocity (Hahn, 1980). The magnetic influence of the MGA is seen quite clearly as highs of nearly 1,000 gammas and flanking lows of several hundred gammas (Yarger, 1983). A potential-fields model of the MGA by Yarger (1983) correlates a high density and highly magnetic rock body with bottom-hole basalt samples from the central part of the MGA (Bickford et al., 1979).

The Nemaha Ridge is another of the major structural discontinuities in the study area. Seismic-reflection evidence suggests major uplift during late Mississippian time which produced the Nemaha Ridge, forming the boundary between the Salina and Forest City basins in Kansas (Steeples, 1982, also this volume). Aside from the MGA and Nemaha Ridge, the Ozark uplift, Sioux uplift, Central Kansas uplift, Forest City basin, Anadarko basin, Denver basin, and Salina basin are structural features having some acoustical significance to this study.

Investigation of deep crustal structure in the study area was undertaken by the Consortium for Continental Reflection Profiling (COCORP) in 1979 with preliminary interpretation done by Brown et al. (1983). The seismic lines were chosen in order to sample subsurface points related to significant geophysical or geological features in Kansas. The structural features of particular interest to the investigators were the Nemaha Ridge, the MGA, and the Humboldt fault. Deep crustal structure also was a primary target of the study.

The COCORP investigation in 1979 suggests the structure below 12 km (7 mi) is characterized by discontinuous, heterogeneous dipping layers with varying thicknesses (Serpa et al., 1984; Brown et al., 1983). In the small area being studied by COCORP, deep-reflection techniques show regions with possible large granitic intrusions and acoustically transparent zones. Below these granitic intrusions are acoustically significant layers with complex reflection patterns, many of which are assumed to be from outside the plane of the survey. The granitic-to-gabbroic rock interface may be of sufficient acoustic impedance contrast to cause the relatively high reflectivity observed throughout this zone of apparent complex structure.

The COCORP investigation in Kansas has been unable to show a vivid continuous reflector from the 11 to 13 sec (34–38 km [20–23 mi]) zone which is the expected arrival time for the crust/mantle interface (Moho) reflec-

tion throughout the survey area in Kansas. However, the character of energy returning from below the assumed Moho reflector zone has subtle irregularities and trace-appearance differences. The assumption is that these character changes are in some way related to the Moho (Serpa et al., 1984).

Reversed-refraction studies

The U.S. Geological Survey (USGS) shot reversed seismic-refraction profiles between Agate, Colorado, and Concordia, Kansas, in 1965 (fig. 3). This section of the paper reports the authors' interpretation of the record sections from those profiles. The uninterpreted record sections are shown by Warren (1975). The USGS seismic-refraction recording system is described by Warrick et al. (1961), and field procedures are outlined by Jackson et al. (1963).

Several refraction surveys have been performed previously in central North America. Jackson et al. (1963) calculated crustal thickness of 48 km (29 mi) and upper-mantle P-wave velocity of about 8.0 km/sec (4.8 mi/sec) for an unreversed profile from eastern Colorado to southwestern Nebraska. Stewart and Pakiser (1962) calculated a crustal thickness of 51 km (31 mi) in eastern New Mexico from an unreversed profile obtained from the Gnome explosion. Tryggvason and Quails (1967) determined a crustal thickness of 51 km (31 mi) and an upper mantle-velocity of 8.32 km/sec (4.99 milsec) for a reversed northeast-southwest refraction profile centered near Oklahoma City. Stewart (1968) calculated an upper-mantle P-wave velocity of 8.0 km/sec (4.8 mi/sec) and a crustal thickness of 40 km (24 mi) for a reversed east-west profile in northern Missouri.

Little work has previously been done in Kansas on either crustal thickness or deep-crustal and upper-mantle velocities. Ewing and Press (1959) estimated the crust to be 35–41 km (21–25 mi) thick from phase velocity of Rayleigh waves. Steinhart and Woollard (1961) estimated from gravity data and the assumption of isostasy at 96 km (58 mi) that the crust thickens to about 48 km (29 mi) at the west end from about 39 km (23 mi) at the east end of the present profile. Herrin and Taggart (1962) showed a P velocity of 8.1 km/sec to 8.3 km/sec (4.9–5 milsec), with the higher velocity occurring near the east end of the present profile.

Data

Shot information is reproduced from Warren (1975) in table 1. Figs. 4-6 show typical records reduced for a velocity of 6.0 km/sec (3.6 rnilsec). Reduced simply means that a rotation of the distance-versus-time coordinate system was made such that waves with a velocity of 6.0 km/sec (3.6 mi/sec) arrive parallel to the distance axis. This allows quick analysis of the data and allows more data to be plotted in a small space. Note that the first arrivals have a velocity of approximately 6 km/sec (3.6 milsec) out to a distance of about 180 km (108 mi). Fig. 5 shows that close-in arrivals indicate at least two different sedimentary velocities above the 6-km/sec (3.6-mi/sec) layer. Fig. 6 shows arrivals beyond 200 km (120 mi) that have phase velocities in excess of 8.0 km/sec (4.8 rnilsec). The linear least-squares fits of lines to the first arrivals beyond 200 km (120 mi) are shown below:

Eastward	Westward
$T = 10.59 + \frac{x}{8.43}$	$T = 8.25 + \frac{x}{8.16}$

Interpretation

The following assumptions are made in the interpretation:

1) The distinct seismic velocities are associated with gently dipping layers separated by discontinuities.
2) The layer(s) beneath the granitic (5.99–6.15 km/sec [3.59–3.69 milsec]) material are laterally homogeneous.
3) The measured first arrivals are the true first arrivals.
4) The compressional wave velocities increase with depth.

Several of the record sections show two distinct arrivals out to a distance of about 15 km (9 mi). The earlier set of arrivals commonly exhibits a velocity of 3.0 to 3.9 km/sec (1.8–2.3 milsec) and are here interpreted to represent the arrivals through the Ogallala Formation, a Tertiary molasse deposit. The second set of arrivals has a phase velocity of 4.4–4.5 km/sec (2.6–2.7 mi/sec) and is considered to be refracted arrivals with the velocity of the Paleozoic sediments except on the Stockton profile. The 4.4-kmlsec (2.6-milsec) arrival is first and direct at Stockton because the Ogallala Formation has been eroded away at the shotpoint.

The thickness calculated for the sediments in Colorado (Hale East) is 2.6 km (1.6 mi) compared to a known thickness of 2.5 km (1.5 mi) from drill evidence. At Stockton the sedimentary thickness is calculated to be 2.0 km (1.2 mi) compared to a known sedimentary thickness of 1.3 km (0.8 mi). The discrepancy is due to an increase in sedimentary thickness westward from the Stockton shotpoint to the recording sites and because the Ogallala Formation is present at the recording sites but absent at the shotpoint. These two effects cause a larger intercept time and, hence, an erroneously large calculated sedimentary thickness.

The arrivals commonly assigned to granites (5.9-6.15 km/sec [3.5–3.7 rnilsec]) arrive first from a distance of about 5-6 km (3–3.6 mi) to a distance of more than 180 km (108 mi; see figs. 6a and 6b). A somewhat gradual increase in the velocity of the granitic layer seems to occur with depth,

TABLE 1—SHOT INFORMATION FROM WARREN (1975).

Name and nmber	May 1965 date	Time, Mountain Daylight	Size, lbs	Latitude N.	Longitude W.
Agate 1	11	07:00:00.27	20,000	39°33.95'	103°44.86'
Hale 1	6	07:00:00.27	150	39°34.41'	102°07.31'
Hale 2	6	11:30:00.53	500		
Brewster 1	6	06:30:00.34	500	39°35.03'	101°25.61'
Brewster 2	6	12:00:00.36	150		
Brewster 3	7	07:00:00.35	150		
Brewster 4	7	11:30:00.35	500		
Brewster 5	9	20:50:00.16	1,000		
Brewster 6	10	07:20:00.30	1,500		
Brewster 7	10	12:00:00.29	2,500		
Brewster 8	10	17:00:0031	4,000		
Selden 1	7	06:30:00.45	500	39°34.40'	100°41.90'
Selden 2	7	12:00:00.44	150		
Selden 3	9	20:00:00.44	150		
Selden 4	10	07:40:00.57	500		
Lenora 1	6	06:30:00.45	4,000	39°34.79'	100°04.45'
Lenora 2	6	11:00:00.09	2,500		
Lenora 3	7	06:00:01.04	1,500		
Lenora 4	7	11:00:01.09	1,000		
Lenora 5	9	20:30:00.41	500		
Lenora 6	10	08:00:00.36	150		
Lenora 7	10	12:40:00.41	150		
Lenora 8	10	17:20:00.34	500		
Stockton 1	10	12:20:00.46	510	39°34.97'	99°20.82'
Stockton 2	10	17:40:00.56	150		
Concordia 1	12	05:00:00.63	20,000	39°31.22'	97°54.67'

so a single least-squares line was fit to the data providing an average velocity. The thickness of the granitic layer is highly dependent upon whether the interpretation includes a deeper mafic (6.7–7.2 km/sec [4.0–4.3 milsec]) layer. The evidence for this layer is not overwhelming or necessarily unique, though it has been used by most investigators in the midcontinent (Stewart, 1968; Tryggvason and Quails, 1967; Jackson et al., 1963). While clear evidence of these intermediate crustal velocities exists, clear evidence for a continuous layer or layers does not. COCORP studies in the northeastern part of Kansas suggest these higher velocities are caused by mafic plutons intruded into the lower crust (Brown et al., 1983; Serpa et al., 1984). If interpretation is done using a deep crustal layer of 6.7 to 7.2-km/sec (4.0–4.3-mi/sec) velocity, the effect is to make the crust seem about 5 km (3 mi) thicker.

Some evidence of a Moho reflection in fig. 6b is found at a distance of 130 to 155 km (78–93 mi). This event, if it is a reflection, suggests an average crustal velocity of 6.2 km/sec (3.7 milsec). This interpretation implies a Moho depth of 40 km (24 mi) near the middle of the refraction line.

Fig. 6c shows the P-arrival data from a distance of 230 km to 410 km (138–246 mi) westward from the Concordia shotpoint. The least-squares fit of 8.16 km/sec (4.89 mi/sec) is shown on the figure to highlight the first arrivals. Fig. 6d similarly displays the data from 220 km to 430 km (132 mi to 258 mi) eastward from the Agate shotpoint, with the first arrival phase-velocity of 8.43 km/sec (5.06 mi/sec) highlighted.

Discussion

Fig. 7 shows a cross sectional model of the refraction data. The resulting crustal thickness in eastern Colorado (46 km [28 mi]) is a couple of kilometers thinner than previously calculated for an unreversed profile by Jackson et al. (1963). The calculation of a 36-km (22-mi)-thick crust in north-central Kansas at the west edge of the MGA is thinner than might have been expected on the basis of previously mentioned results in Missouri and Oklahoma. This number is reasonable, however, as discussed later.

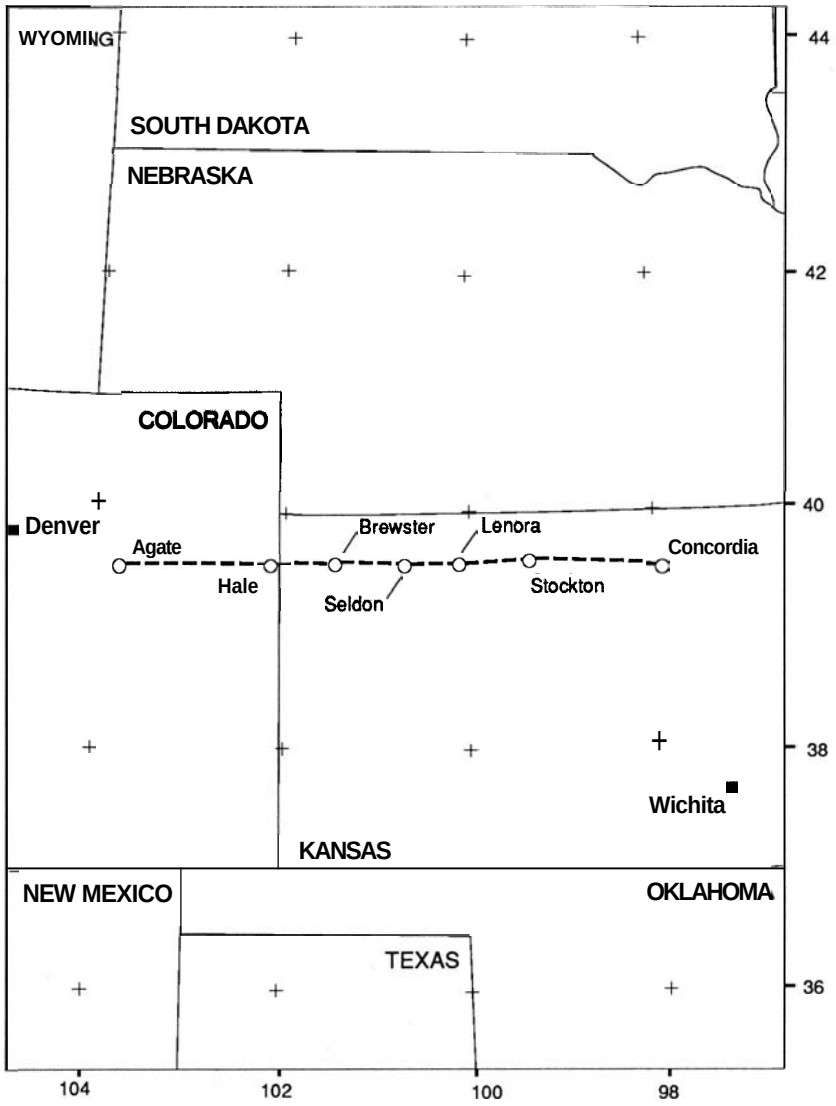


FIGURE 3—LOCATION OF LINE OF SHOTS. Open circles show shotpoints; recordings were made along dashed line.

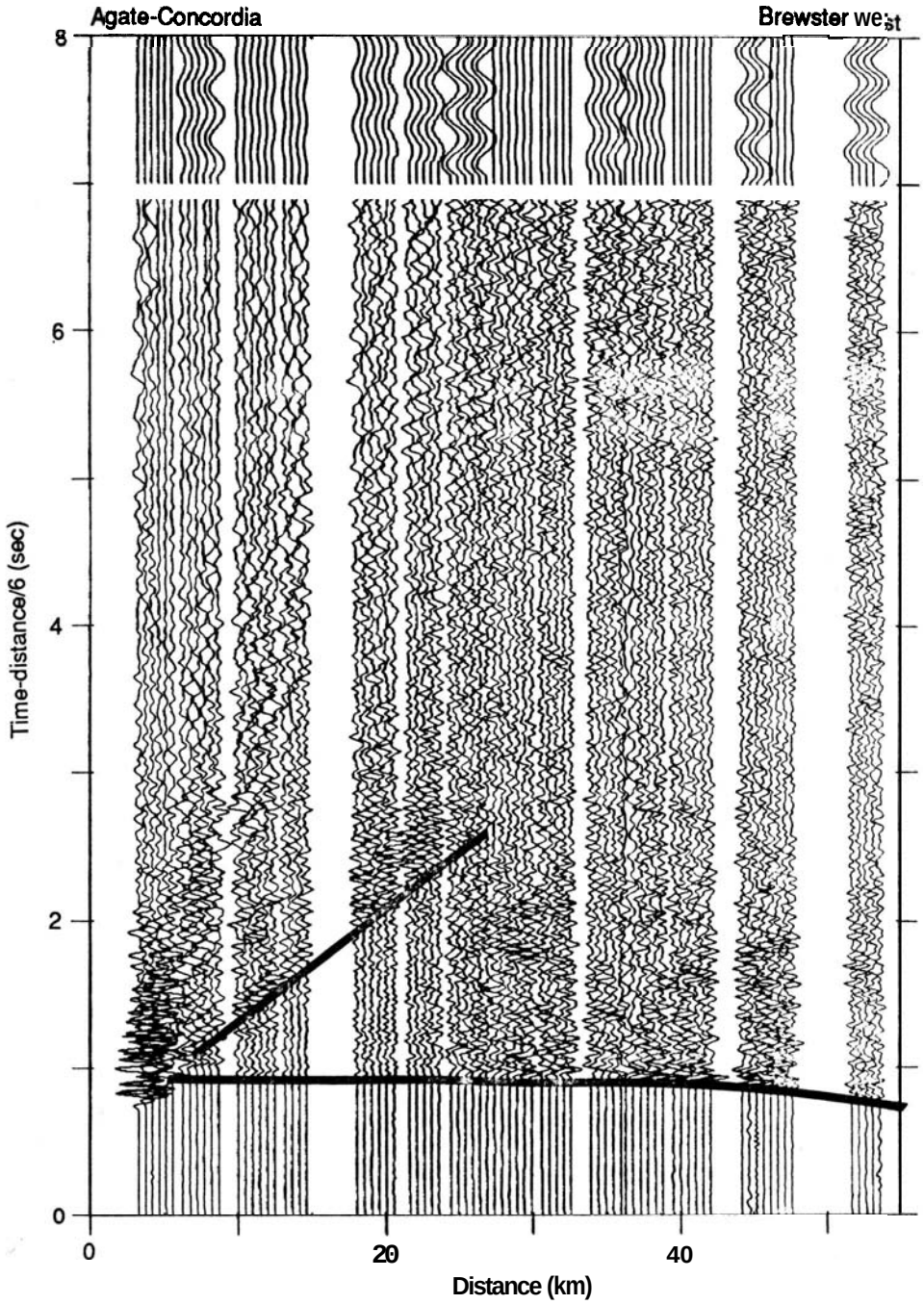


FIGURE 4—A TYPICAL REDUCED RECORD SECTION (time vs. distance curve). The heavy dark line shows interpreted direct wave (sedimentary veneer) and first defraction.

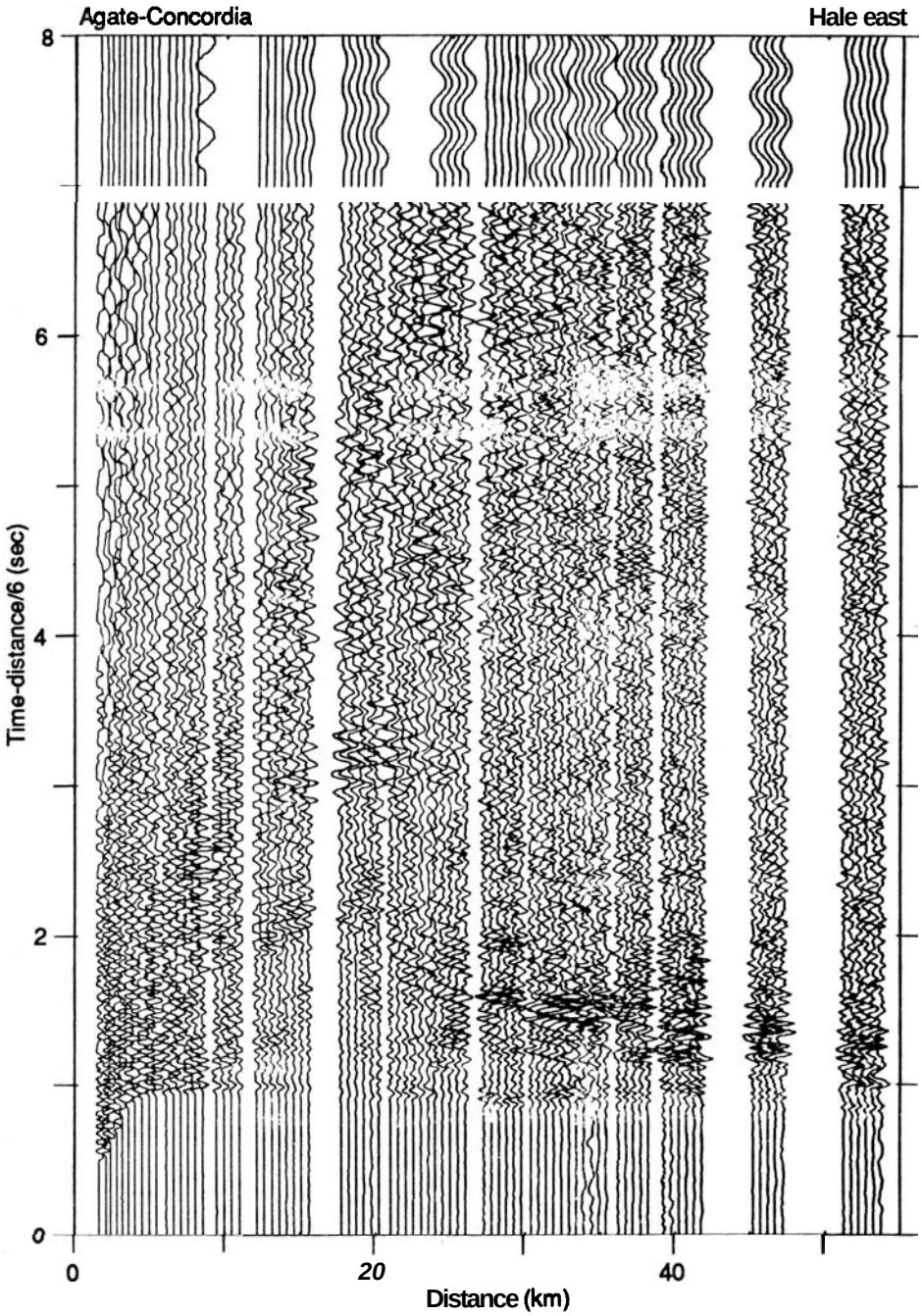


FIGURE 5—ON THE VERY NEAR TRACES, TWO DISTINCTLY DIFFERENT SEDIMENTARY VELOCITIES CAN BE INTERPRETED.

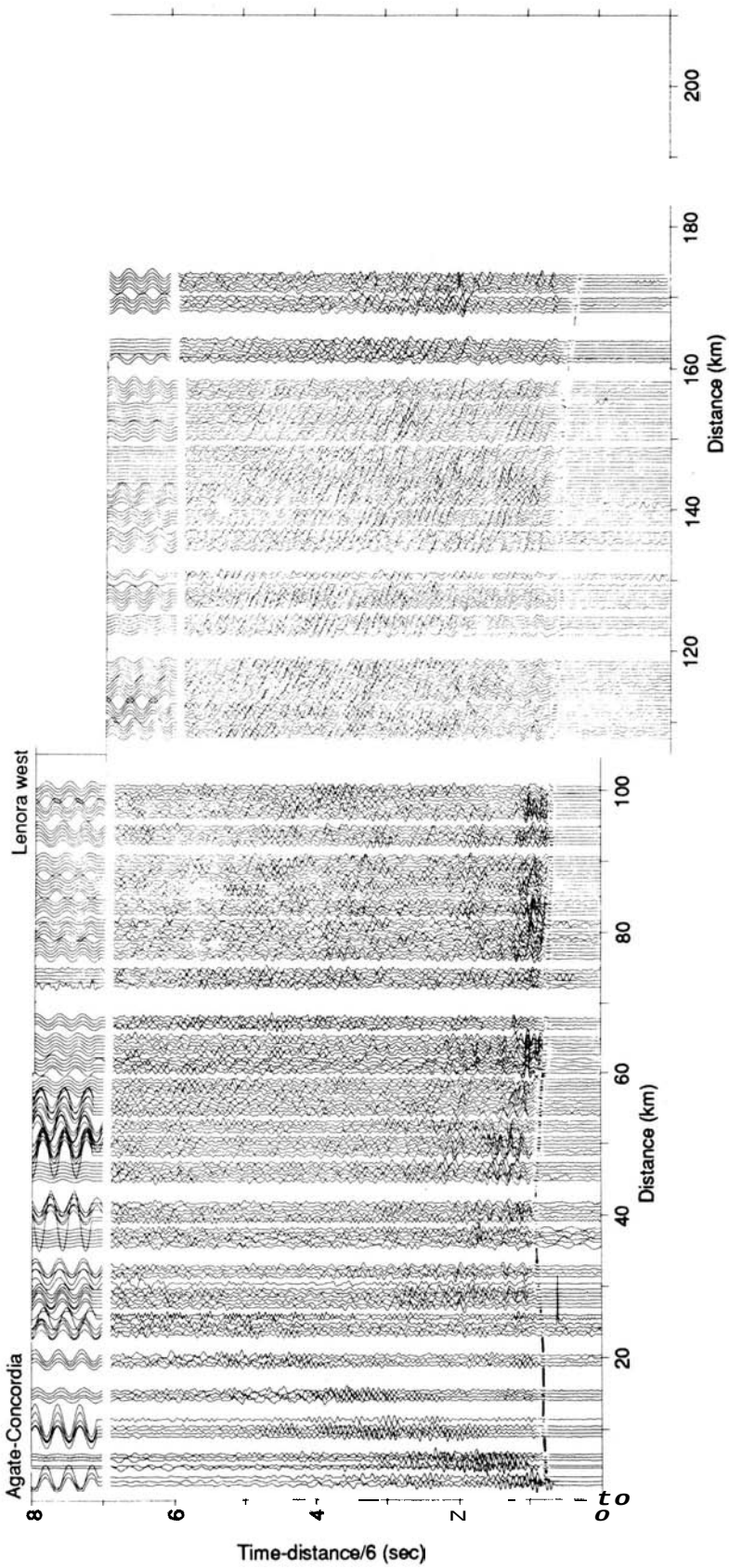


FIGURE 6—A) ARRIVALS TO DISTANCES OF 175 KM (105 MI) SHOT WESTWARD FROM LENORA, KANSAS.

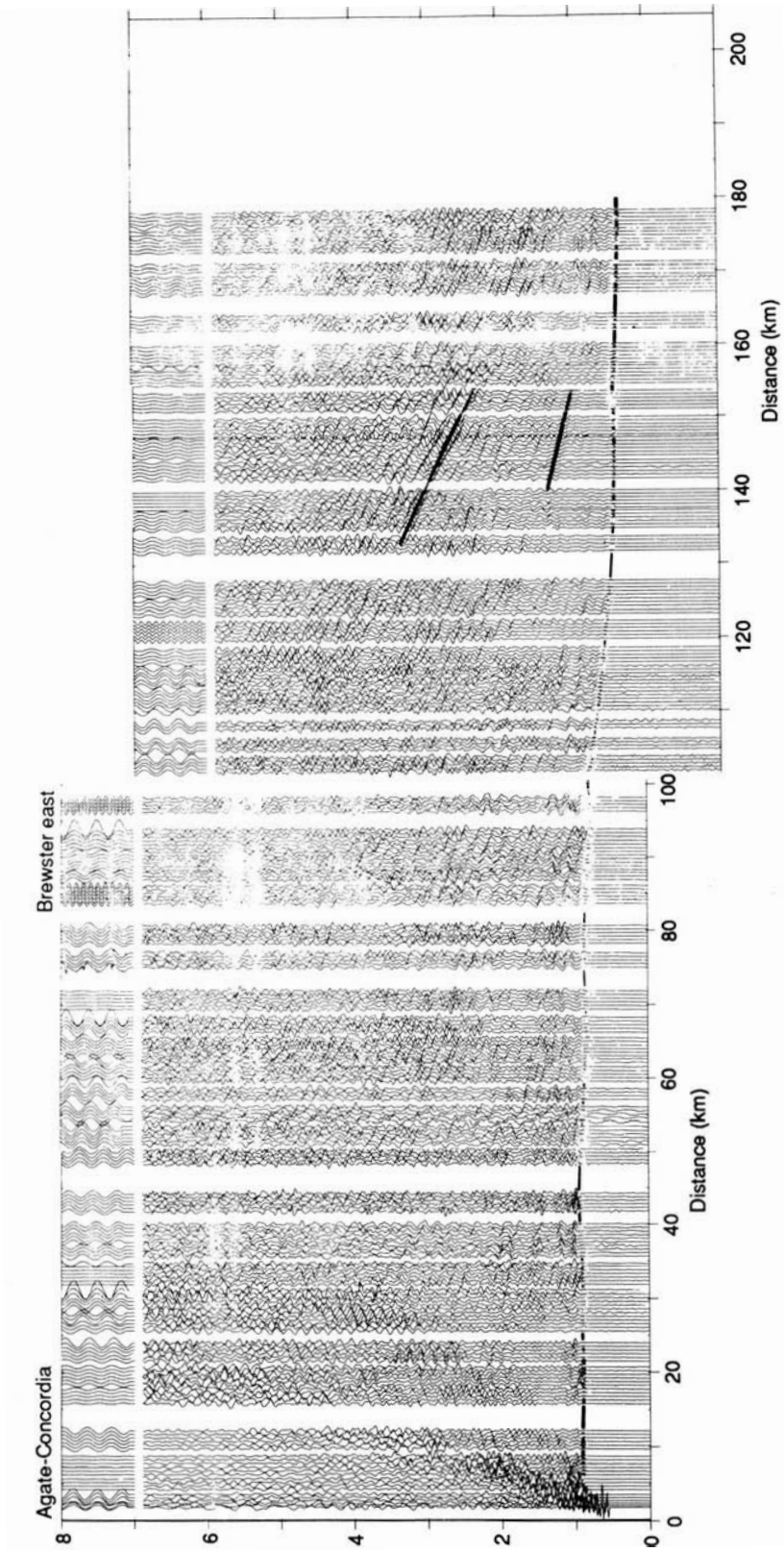
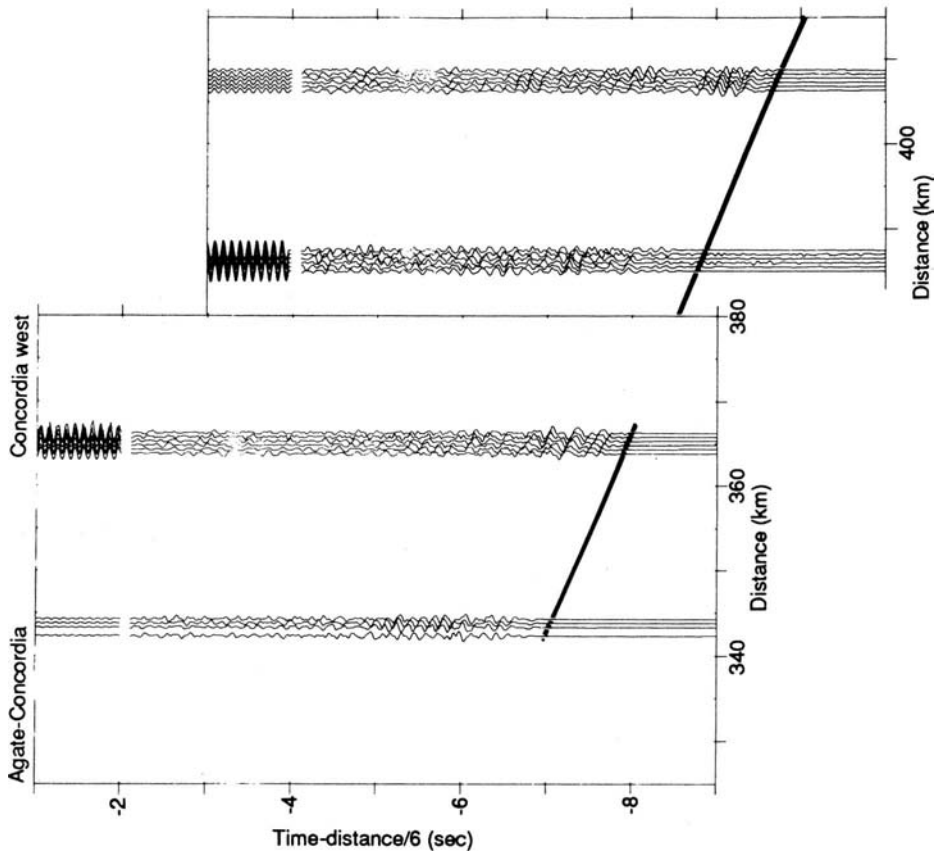


FIGURE 6—B) ARRIVALS TO DISTANCES OF 180 KM (108 MI) SHOT EASTWARD FROM BREWSTER, KANSAS. POSSIBLE MOHO REFLECTIONS ARE HIGHLIGHTED BY DARK LINES BETWEEN 132 AND 153 KM DISTANCE.



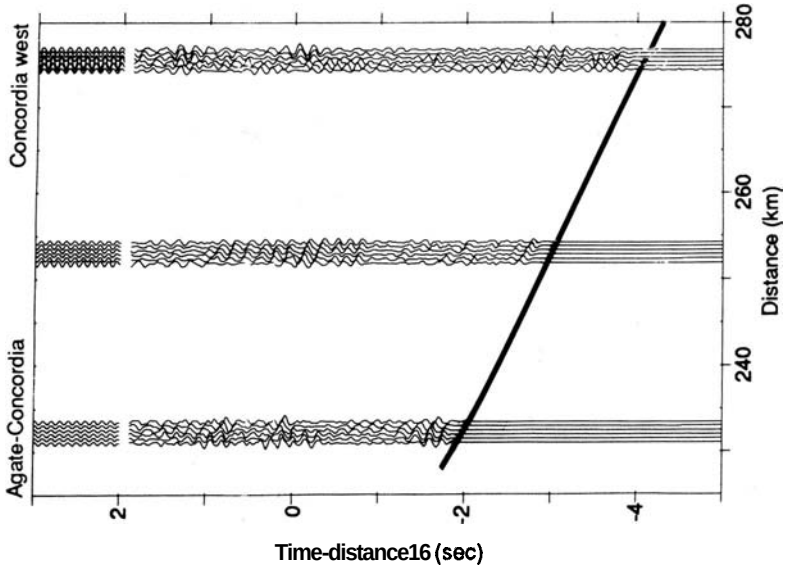
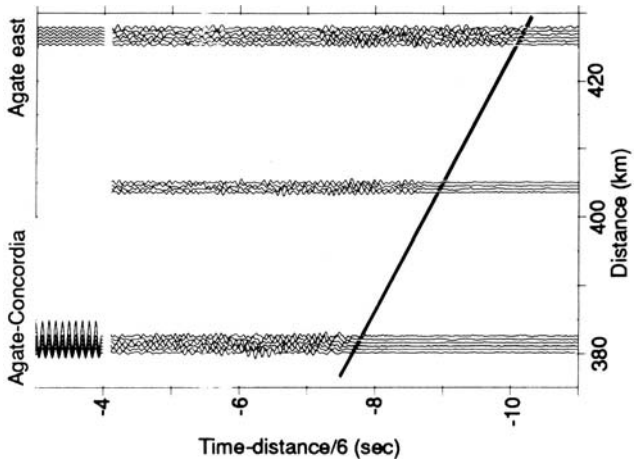


FIGURE 6—c) THE REDUCED TIME-DISTANCE CURVE HERE SHOWS ARRIVALS FROM RECEIVERS GREATER THAN 200 KM (120 MI) WESTWARD AWAY FROM THE SHOT NEAR CONCORDIA, KANSAS.



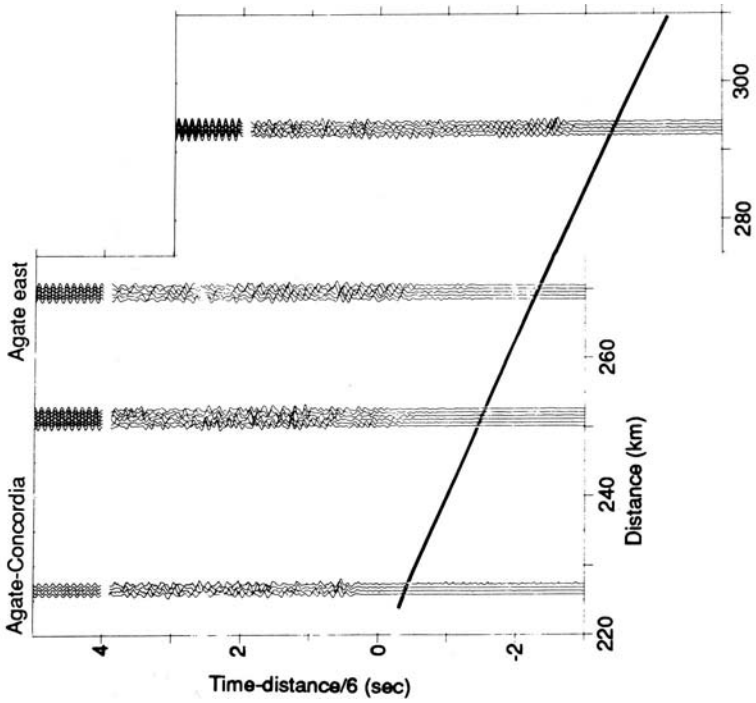


FIGURE 6—D) THE REDUCED TIME-DISTANCE CURVE HERE SHOWS ARRIVALS FROM RECEIVERS GREATER THAN 200 KM (120 MI) EASTWARD FROM THE SHOT NEAR AGATE, COLORADO.

The P_n velocity at the top of the upper mantle is high enough that it deserves special attention. The phase velocities of P_n are 8.16 km/sec (4.90 milsec) for the westward direction and 8.43 km/sec (5.06 milsec) for the eastward direction. This in itself shows that the crust must thicken to the west. When the problem is solved for the general dipping case, a true P_n velocity of 8.29 km/sec (4.97 milsec) results while the dip angle is 1.03' westward for the crust-mantle interface measured from the topographic surface. This high velocity is not caused by picking errors on the seismograms, since such errors usually involve late picks, resulting in lower apparent velocities.

The P_n velocity reported here is nearly equal to the highest P_n velocity reported in the United States (8.32 km/sec [4.99 milsec] in Oklahoma by Tryggvason and Quails, 1967) and the 8.43 km/sec (5.06 milsec) P_n phase velocity for the eastward profile is the highest reported. Some investigators have shown P_n phase velocities in excess of 8.4 km/sec (5.0 mi/sec) at distances exceeding 445 km (267 mi) in Canada (Mereu and Hunter, 1968; Eisler and Westphal, 1967), but data in the present work are not recorded beyond about 440 km (264 mi).

Gravity data

Fig. 8 shows regional gravity data excerpted from a USGS gravity map of the United States. Note the regional gravity increases steadily eastward. The gradient is almost exactly parallel to the seismic profile of this study which is one reason the line was shot at this location with the east-west orientation. A least-squares fit of a linear gradient to the observed gravity contours along the seismic profile shows the Bouguer anomaly to increase rather uniformly eastward by 0.27 mgal/km. Thus a direct comparison of the gravity model and the seismic data is relatively simple.

The gravity modeling was done using the approximation presented by Nettleton (1976, p. 201–203) as follows

$$Ag = 41.93\ LT \tan 9 \qquad (1)$$

where Ag is gravity difference in milligals, L is the distance in kilometers, 9 is the dip angle between points A and B, and T is the density contrast between layers. For θ of about 11° this approximation introduces error of only 3%, so the errors arising here from the approximations should be less than 0.3% because 9 is about 1° for the model discussed. Gravity fields may be calculated for individual dipping layers, then summed to produce the gravitational effects for each model.

Most of the observed 120-mgal change in the Bouguer anomaly must be due to deep structure, as only about 6 mgal could be due to changes within the sedimentary cover (Woollard, 1959). As mentioned earlier, Steinhart and Woollard (1961) assumed isostatic compensation at 96 km (58 mi) depth to estimate crustal thinning of 10 km (6 mi) along this seismic profile. On the basis of Bouguer anomalies, they estimated a crustal thickness of 48 km (29 mi) at the west end and 39 km (23 mi) at the east end of the profile. That estimate is consistent with the combined gravity and seismic-refraction results from the present study, although our comparable numbers are 46 km (28 mi) and 36 km (22 mi), respectively. The density contrast between the crust and upper mantle needed to satisfy the 10 km (6 mi) of crustal thinning is 0.32 gm/cm^3. This is a reasonable number—possibly a bit lower than might be expected.

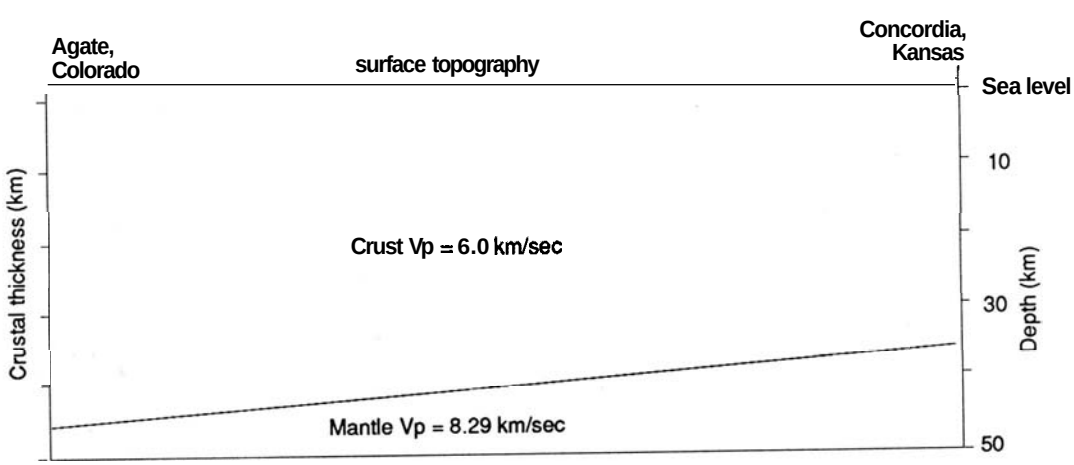


FIGURE 7—CROSS SECTION SHOWING CRUSTAL THICKNESS, AVERAGE VELOCITY OF THE CRUST, AND VELOCITY OF THE UPPER MANTLE. This cross section directly correlates to the regional gravity-gradient plot of fig. 8.

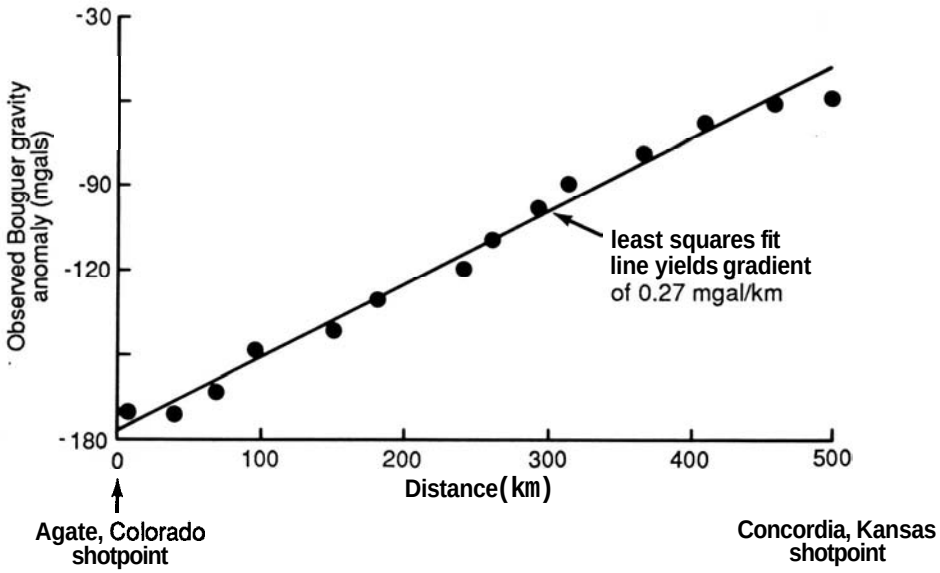


FIGURE 8—THE REGIONAL GRAVITY GRADIENT WAS MEASURED ALONG THE PROFILE FROM CONCORDIA, KANSAS, TO AGATE, COLORADO. The gravity data agree quite well with the reversed-refraction profile-derived cross section.

Regional earthquake travel times

In the previous section of this paper, reversed-refraction profiles with a maximum source separation of 500 km (Concordia, Kansas, to Agate, Colorado) were used to develop crustal models for western Kansas. In this section of the paper, analysis of compressional wave arrivals from regional (epicentral distance greater than 250 krn [150 mi]) earthquake seismograms is used to study the crust and upper mantle in eastern Kansas.

The use of earthquakes as the energy source for refraction profiles has not been extensively investigated. Earthquakes in the midcontinent typically occur several kilometers beneath the earth's surface, whereas explosive sources necessarily occur very near the surface. Much of the travel-time variation from regional distances can occur within the upper few kilometers of the crust.

Receiver network

The earthquake network in Kansas began operation at a high level of sensitivity in December 1978. The nine-station network was designed and installed to locate microearthquakes along potentially active geologic features in eastern and central Kansas (fig. 9). The network possesses the sensitivity necessary to locate events with duration magnitude as low as 1.5 in the eastern half of Kansas (Sheehan and Steeples, 1983). Funding for the stations was provided by the United States Nuclear Regulatory Commission (NRC) and the Kansas City District, U.S. Army Corps of Engineers.

Residuals

As a head wave propagates across the network, it interacts with anomalous velocity zones that originate and/or terminate within the network. If heterogeneities of this sort exist, energy propagation across the network will be nonlinear in time. The size of the time variation (At) is dependent on the size and orientation of the anomalous (At) zone and the velocity contrast between zones (V_2, V,)

$$\Delta t = \frac{\Delta h\,(V_2 - V_1)}{(V_2 \cdot V_1)} \qquad (2)$$

P_n arrivals, by definition, penetrate all layers of the crust. Any major anomalous zone between the crust/mantle interface and the ground surface (within a volume defined by the areal extent of the network and the critical angle of the crust/mantle interface) should result in a deviation in the actual from the expected arrival time of a head wave through a homogeneous medium. The residual values are directly related to the deviation in actual arrival time from expected.

The calculation of residuals is implemented with one assumption. The least-squares determined slope of the line intersecting the first-break times at the proper source/receiver offset is the true apparent velocity for the layers. Then with the use of the equation for a line, residuals are calculated by the following equation:

$$\textit{Residual} = \textit{Arrival Time} - \frac{\textit{Epidistance} - \textit{Intercept}}{\textit{Velocity}} \qquad (3)$$

From the travel-time residuals (i.e. the amount of time a phase arrival deviates from the straight-line approximation of the apparent velocity of propagation), a characteristic directional delay or early arrival for each station can be identified. Certain stations show consistently late or early arrivals, depending on crust and upper-mantle anomalies. Using a statistical approach of determining directional delays enables interpretation of velocity anomalies and apparent unconformities within the subsurface of the sampled area.

Data analysis

Picks of impulsive first-break energy (P_n) inherently carry more integrity than picks of energy arrivals later in the wave train (Pakiser and Steinhart, 1964). Therefore, all analysis of earthquake energy on seismograms will be restricted to the first arrivals. Events chosen for this study had sufficient energy to arrive at a majority of the receiver stations as an impulsive first arrival. For most stations, this would amount to an instantaneous change in amplitude of approximately 20 dB. Events 8, 9, 10, 11, and 12 were located by the Oklahoma Geophysical Observatory and have general progression patterns from south-southwest into the network with epicentral distances ranging from 365 km (219 mi) for the closest station to about 900 km (540 mi) for the most distant station with discernible phase arrivals. The five south events are plotted with a vertical tick indicating time and amplitude of each interpreted phase arrival (fig. 10). Events 8 and 9, originating within Oklahoma, show very similar wave-train character (fig. 11). An apparent velocity of 8.35 km/sec±0.15 km/sec (5.01 mi/sec±0.09 mi/sec) for the crust/mantle interface recorded by these two events is within the experimental error of previously obtained values (Tryggvason and Quails, 1967). Events 10, 11, and 12 originated from the Texas panhandle region. Event 12 has features that closely resemble the two Oklahoma events (8 and 9). The slight variation in values might be associated with subtle regional changes in crustal thickness and upper crustal heterogeneities. Comparison of events 8, 9, and 12 gives a model with a large subsurface sampling area. This model, however, would not show the proposed Moho low in central Oklahoma (Warren and Healy, 1973).

The upper crustal structure between the eastern events and the recording network is very complex (Stewart, 1968). Events 2, 3, and 4 were used as indicative of the subsurface to the east (fig. 12). These events show good P_n velocity agreement within themselves, with an average of 8.21 km/sec ±0.15 km/sec (4.93 mi/sec±0.09 mi/sec). They also all have epicenters in the Mississippi embayment. Earthquakes that originate from outside a concentrated grouping in northeastern Arkansas lack regional consistency and are used predominantly for station residual determinations.

Event 6, originating in Kentucky, contained sufficient energy to saturate the analog recorders. All significant information beyond excellent first-break energy was clipped. Event 6 has an epicentral distance greater than any other studied here. After rejection of LAK and EDK from the data set due to poor correlation to the least-squares line, an apparent P_n velocity of 8.87 km/sec H.15 km/sec, was obtained at offsets from about 1,000 to 1860 km from the source. This observation is discussed in more detail later in this paper.

Events 13, 14, and 15 are the northern data set (fig. 13). These three events have calculated epicenters with horizontal separation between events 13 and 15 of approximately 580 km (350 mi) and between events 13 and 14 of 150 km (90 mi). Event 15 has an emergent P_n arrival and therefore was not used to determine the P_n velocity from the north. Events 13 and 14 have apparent P_n velocities that show good consistency, very little deviation, and were calculated from picks of impulsive first breaks. For those reasons the apparent P_n velocity for the northern crust is formulated using events 13 and 14. The apparent P_n velocity from the north is 8.38 km/sec±0.15 km/sec (5.03 mi/sec±0.09 mi/sec).

As characteristic of many seismic-refraction surveys when applying statistical techniques, lack of sufficient data poses a sizeable difficulty in interpreting data on travel curves (Borcherdt and Healy, 1968). This problem was encountered in dealing with energy arrivals originating in the west. Only one event falls within the guidelines for data acceptance. From event 16, an apparent P_n velocity of 8.59 km/sec H.15 km/sec (5.15 mi/sec±0.09 mi/sec) was calculated.

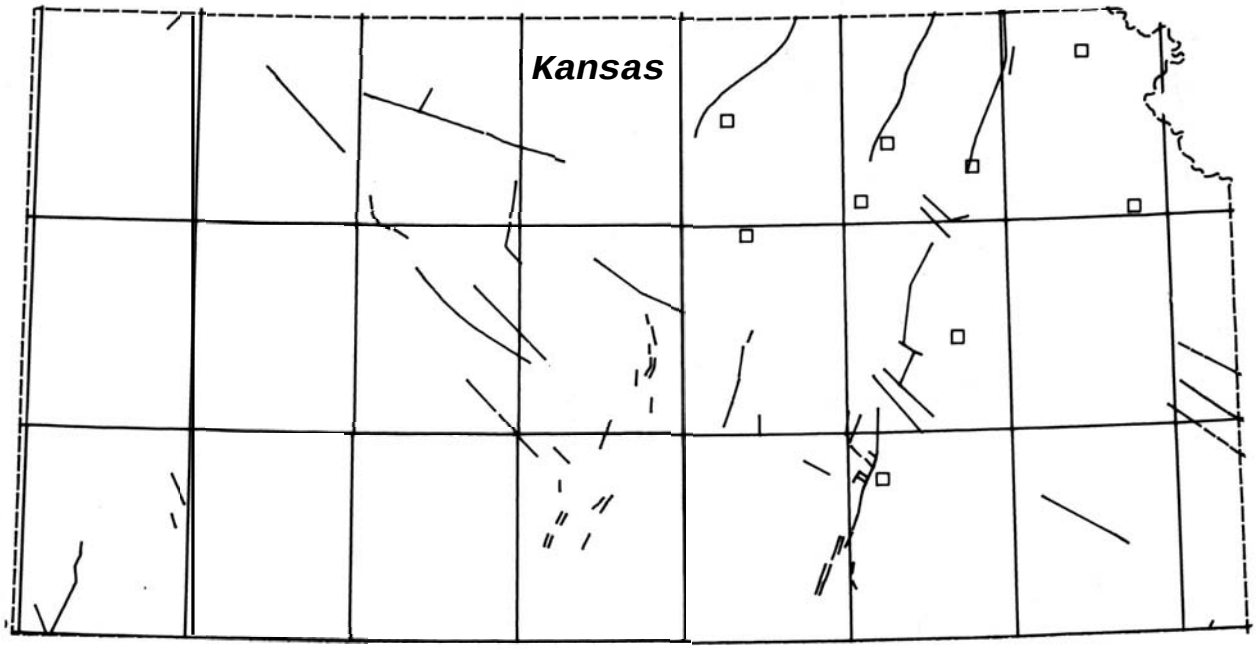


FIGURE 9—KANSAS GEOLOGICAL SURVEY'S EARTHQUAKE STATIONS IN CONJUNCTION WITH MAPPED FAULTS (□ = station locations; / = faults).

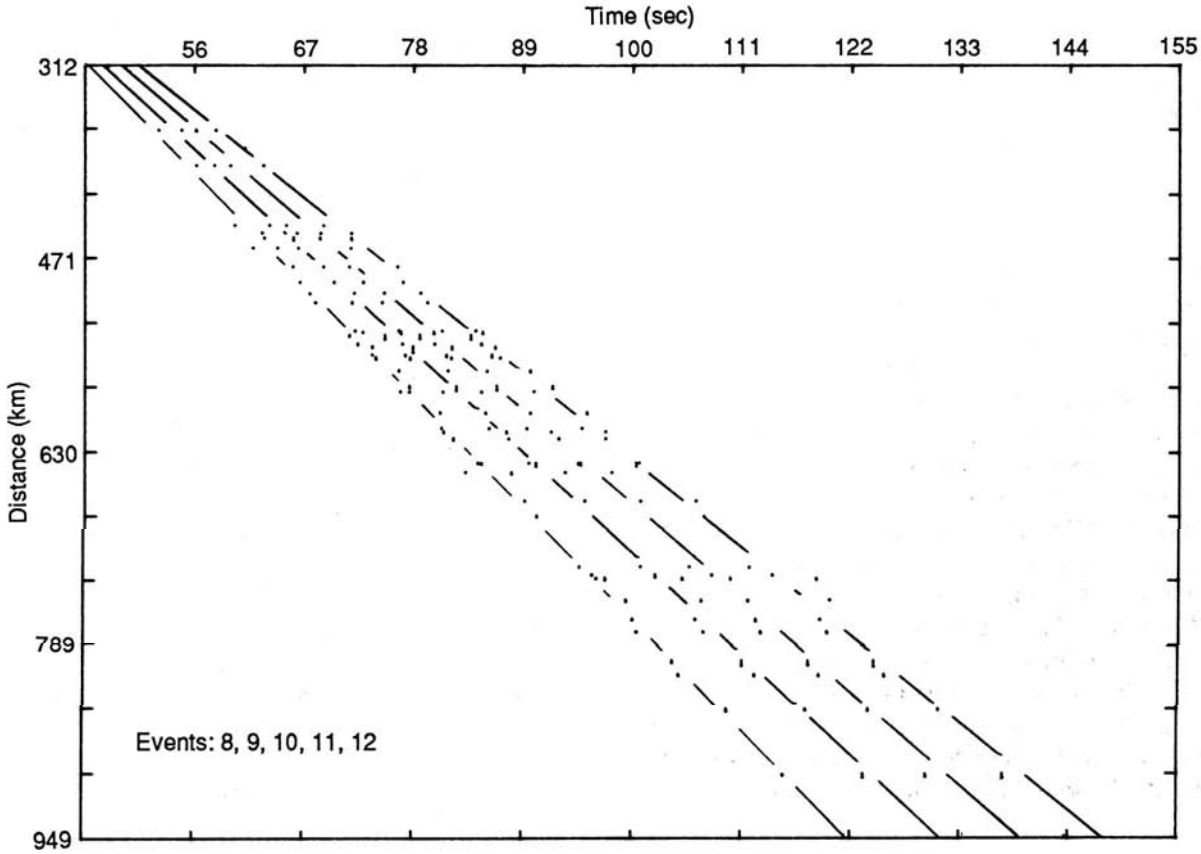


FIGURE 10—ALL EVENTS FROM THE SOUTH USED IN THE FORMULATION OF THE RESIDUAL MODEL.

Discussion

Using earthquakes as the energy source for refraction surveys restricts much of the control over the data set that is present on conventional refraction profiles. Since the stations are concentrated in eastern Kansas in a somewhat circular array (fig. 9), refraction-type studies, using the stations as receivers, lack the two-dimensional precision of conventional refraction studies. However, by using a great circle technique of station alignment, regional features should have only minor distortion (assuming proper sampling). Thus, by comparing several good-quality events arriving at the network from the same general direction, certain acoustic properties and apparent structural anomalies of the crust and upper mantle can be identified.

An anomalous high-velocity crust in the northwestem portion of the network can be interpreted from P_n residuals (fig. 14). The presence of such a velocity anomaly has been suggested in previous studies (Lui, 1980; Serpa et al., 1984). Good correlation can be found between the potential-fields model and the seismic-velocity delay model (Yarger, 1983). Transition of a P-wave from a granitic crust, which is commonly accepted as the predominant crustal rock type in the area, into a basaltic environment could give velocity residuals similar to those observed here. The size of the residual, of course, also is dependent upon height of the anomalous rock column. Exact determination of the anomalous body's boundaries using the existing station array is not possible. Boundaries can be estimated using the stations within the anomalous material as control points for a minimum affected area. In general, the anomalous high-velocity area includes CNK, SNK, MLK, and TCK (fig. 14). Also, localized patches of anomalous crust are found in the central part of the station array many times only observed on one or two stations.

Consistently late arrivals at EDK from regional events could be related to a geophysical anomaly previously detected and its borders defined by an aeromagnetic study of Kansas (Yarger, 1983). The anomaly is represented by a low of 600 gammas centered near EDK. A continental collision zone has been given as a possible explanation for the low (Yarger, 1983). Rock types at the Precambrian surface in the area lack any evidence for the presence of this aeromagnetic low (Bickford et al., 1979). However, some type of plate-convergence process, which could have been operational during mid-Proterozoic time, has been suggested (Van Schmus and Bickford, 1983). Local-event residuals do not suggest the presence of low-velocity material in the upper part of the crust. The arrival-time patterns observed therefore are consistent with both the shallow Precambrian geology of Bickford et al. (1979) and Van Schmus and Bickford, (1983) and the hypothesized collision zone of Yarger (1983).

Examination of first-arrival P-wave residuals from event 6 (from Kentucky) reveals possible evidence for a low-velocity upper mantle near the MGA. An apparent P-wave velocity of 8.87 km/sec±0.15 km/sec (5.32 mi/sec±0.09 mi/sec) was calculated for a 1,000 km to 1,860 km (600-1,116 mi) source-to-receiver offset. Previous studies in the mid-continent have documented a 328-km (197-mi)-deep velocity discontinuity with a 9.27 km/sec±0.05 km/sec (5.56 mi/sec±0.09 mi/sec) velocity at a source/receiver offset of 1,700 km (1,020 mi) (Masse, 1973). Assuming the documented velocity is representative of the midcontinent at the indicated offsets, the slower velocity observed here at similar offsets could be the result of a low-velocity upper mantle beneath the network. The observed velocity and suggestion of a lower-velocity upper mantle is consistent with observations and conclusions drawn from teleseismic P-wave arrivals by Hahn (1980). Only event 6 had sufficient source/receiver offset for the first arrivals to be refractions from a sub-Moho velocity discontinuity.

Within the crust, complex reflection patterns from COCORP data have been hypothesized to be the result of either gneissic banding, interlayering of granite and restites relating to anatexis, or mafic intrusions (Serpa et al., 1984). Complex structure results in multiple interactions with anomalous velocity zones, complicating the interpretation. Refracted waves originating from within this complicated structure would lack good continuity and arrival consistency from one surface location to the next. Therefore, interpretation of intermediate-layer depths and velocities from secondary compressional-wave phase arrivals is speculative at best. This applies to both the reversed-refraction data and the earthquake-arrival data.

Exact determination of a crustal model using the earthquake data set has not proved to be totally successful. However, from analysis of P_n energy incident on the network from different directions, specific characteristics of the crust and upper mantle can be identified. Certain characteristics (amplitude of intermediate arrivals, frequency of impulsive head waves, etc.) of the P-wave arrivals studied showed dependence on local geology and direction from which the energy originated. Assuming gross consistency in characteristic velocities, approximate trends in layer dip and directional-station delays can be determined. From the four apparent P_n wave velocities determined for each direction data set, the Moho dips away from the study area to the west, north, and south, and seems to level off to the east. The velocity of the upper mantle is approximately 8.25 km/sec ±0.15 km/sec (4.95 mi/sec±0.09 mi/sec). Station delays could indicate anomalous velocity zones or crustal heterogeneities possibly representative of previous tectonic activity or vulcanism.

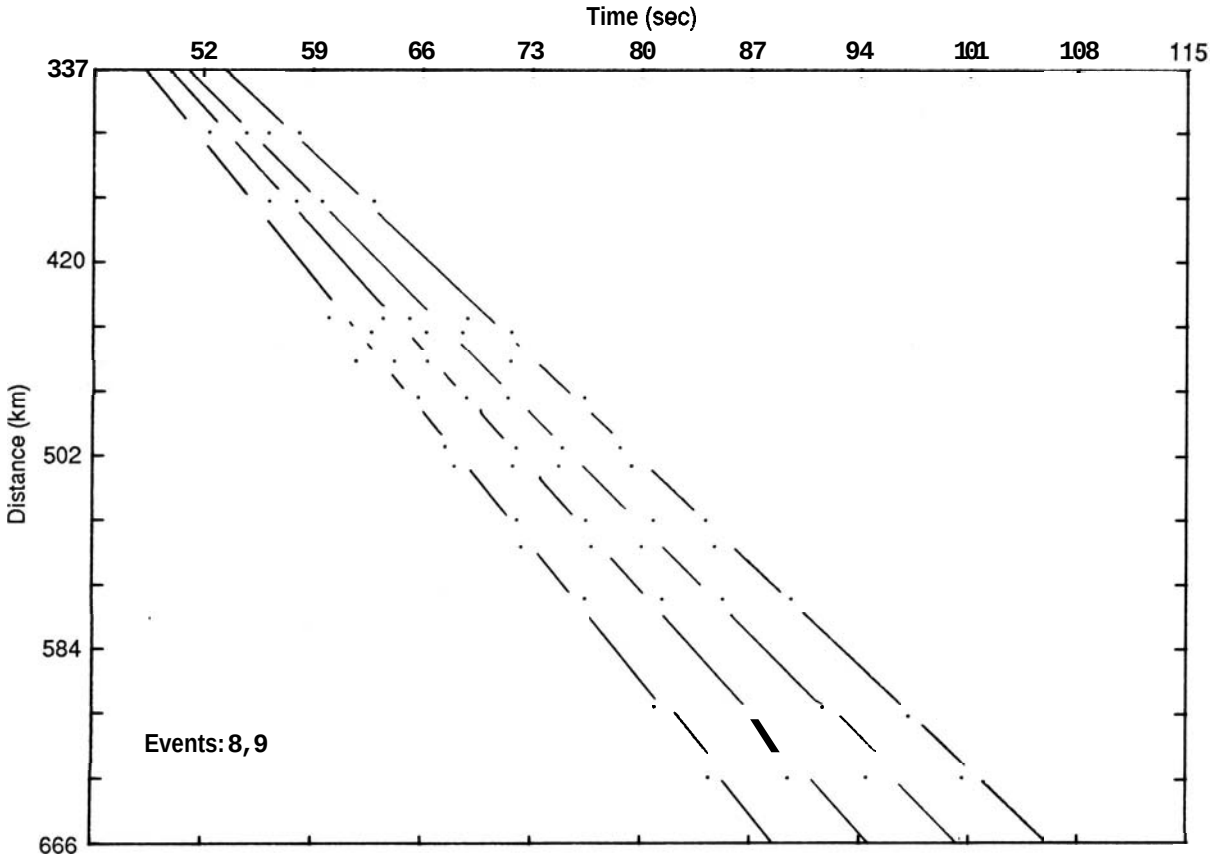


FIGURE 11—EVENTS FROM THE SOUTH USED TO DETERMINE CORRELATION OF PHASE.

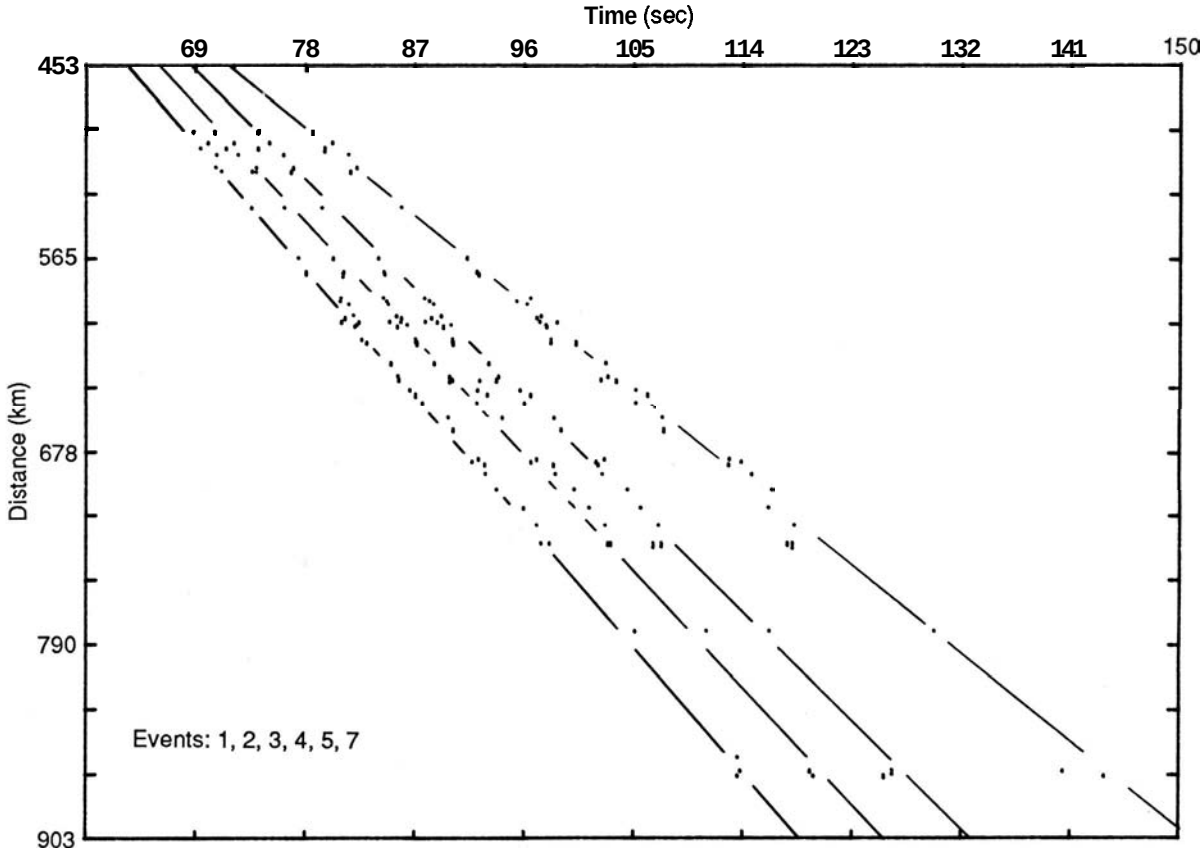


FIGURE 12—ALL EVENTS FROM THE EAST USED IN THE FORMULATION OF THE RESIDUAL MODEL.

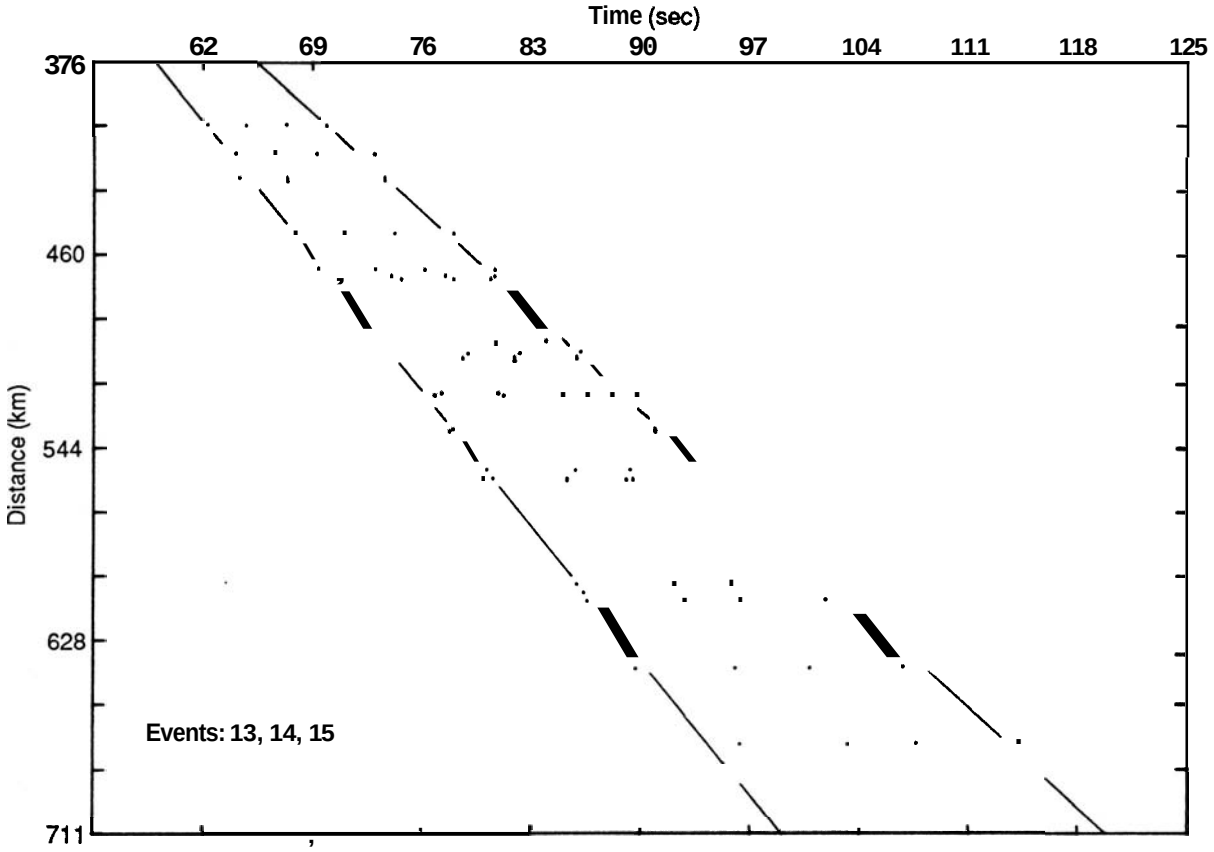


FIGURE 13—ALL EVENTS FROM THE NORTH USED IN THE FORMULATION OF THE RESIDUAL MODEL.

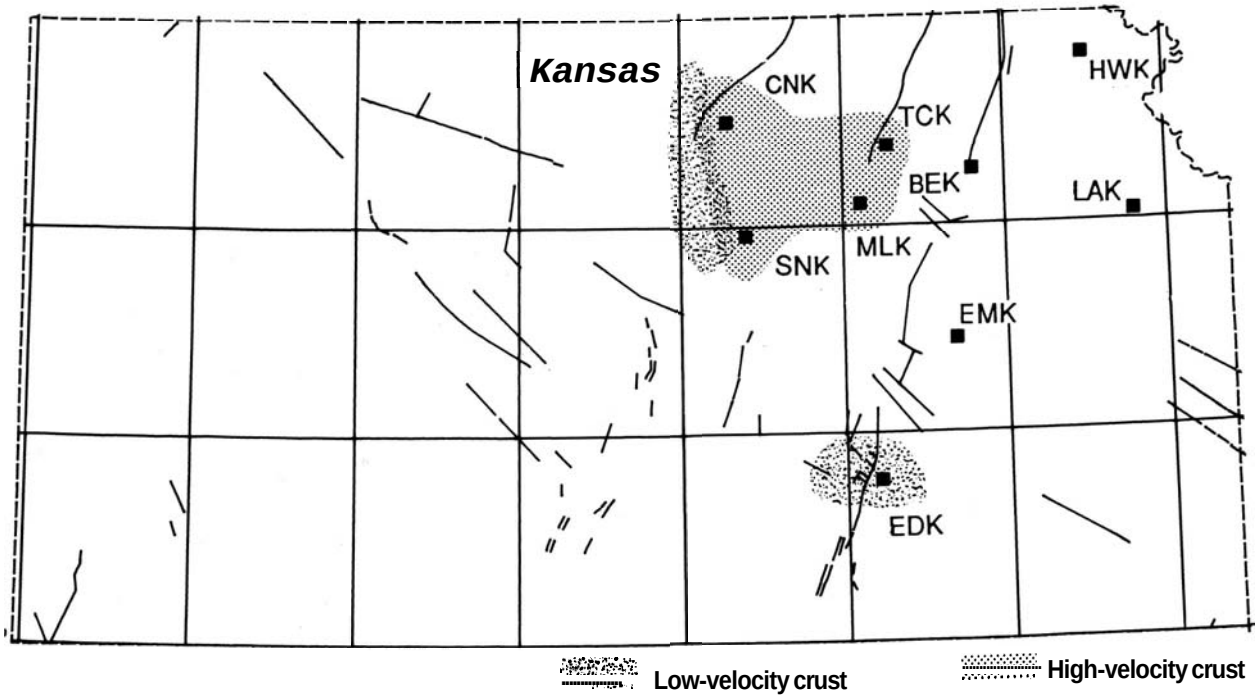


FIGURE 14—P_N RESIDUALS INTERPRETATION.

Conclusion

Reversed-refraction profiles show that the crust-mantle interface has an apparent dip of 1° westward along the profile in fig. 3. The linear gravity gradient shown in fig. 8 is virtually parallel to the refraction profile, suggesting that the 1° dip value is probably true dip. Crustal thinning of about 10 km (6 mi) occurs between eastern Colorado and north-central Kansas. Density contrast of 0.32 gm/cm^3 at the Moho simultaneously fits the gravity data and the seismic-refraction data.

Earthquake P_n wave residuals were used to postulate the presence of several anomalous velocity areas (fig. 14). Earthquake stations CNK, BEK, TCK, and MLK clearly define an anomalous zone in the crust with a high-velocity center flanked by low-velocity borders on the east and west (high and low velocities are qualitative with respect to the average crustal velocities over the entire network). This strengthens the results of Lui (1980) and Serpa et al. (1984) that a high-velocity crust exists beneath the MGA. Consistent delays in the P_n wave at EDK may result from the same anomalous zone that causes a strong aeromagnetic low centered near EDK previously identified by Yarger (1983).

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