The New York - Alabama Magnetic Lineament: its reflection character and relationship to the Grenville Front

by

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Dissertation submitted to the Faculty of the

Virginia Polytechnic Institute and State University

in partial fulfillment of the requirements for the degree of

Doctor of Philosophy in Geophysics

in the

Department of Geological Sciences

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July 1995

Blacksburg, Virginia

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(ABSTRACT)

The source of the New York - Alabama Magnetic Lineament (NYAML) is revealed on newly reprocessed seismic reflection data, and the Grenville Front Tectonic Zone (GFTZ) is imaged beneath it in eastern Tennessee. Industry data, correlated to lower crustal depths, image a wedge-shaped block beneath the shelf strata of the Cumberland Plateau and Valley and Ridge provinces of eastern Tennessee. Two dimensional gravity and magnetic modeling corroborate the interpretation that the contrast in density and magnetic susceptibility between the wedge and the adjacent crust produces the Lineament. The boundary across which the contrast is generated dips approximately 30^{o} northwest.

East-dipping reflections imaged below 7 seconds can be extended northwest to the surface where they align with the position of the Grenville Front. The reflections are interpreted as evidence of deformation related to the GFTZ in Canada. The mid-crustal band of reflectivity visible on most of the reflection profiles lies above the east-dipping reflections and is interpreted to delineate the eastern margin of the GFTZ.

The crust southeast of the wedge-shaped block exhibits high reflectivity with well-developed west-dipping events. The west-dipping events might correlate to those reported in Ohio on COCORP data, suggesting that they are pervasive in the basement throughout the eastern United States. The fabric is interpreted to have formed during the continent-continent collision of the Grenville Orogeny. The absence of west-dipping reflections within the wedge suggests that the wedge is younger than the

development of the fabrics recorded by the reflections. Vertical dike swarms are interpreted to intrude the west-dipping fabric. The swarms model as felsic and appear on migrated data to be older than uplift, erosion, and deposition of the shelf strata.

Crustal thickness estimates by previous authors of over 45 km are corroborated with interpreted images of the Moho on two deeper reflection profiles. The thick crust might be the locus of anatectic melting following Grenville collision. The emplacement of granitic or granodioritic magmas provides an explanation for the density, magnetic susceptibility, and difference in reflectivity of the wedge-shaped block.

The New York - Alabama Magnetic Lineament diverges from the location of the Grenville Front north of the study area. The position of the NYAML can be interpreted to represent the axis of anatectic melting following collision, and indicates that the thickest part of the crust formed farther east of the Front in Canada than in Tennessee.

Pseudomagnetic field investigations permit the distinction between the source of the New York - Alabama Magnetic Lineament and adjacent high susceptibility sources to the northwest. The sources to the northwest appear from the modeling to be mafic intrusions that might be related to the Norris Lake peridotite.

Earthquake locations in the Eastern Tennessee Seismic Zone (ETSZ) are aligned along the southwest edge of the gradient of the NYAML, and fall within the crust characterized by strong west-dipping reflections. Because the contact between the wedge and the region of west-dipping reflections is dipping to the northwest, the relationship between the NYAML and the ETSZ is not clear. More accurate hypocenter locations are necessary to clarify whether the earthquakes are restricted to the region of the crust typified by west dip. If not, the relationship between the earthquakes and the NYAML might be coincidental. A velocity model that considers the dipping boundaries in these reflection data should result in hypocenter locations that can constrain the relationship.

Acknowledgements

The pursuit of a Ph.D. does not occur in a vacuum of support. The success I have or will achieve is due in a great measure to the encouragement and confidence of my husband, Lynn Partington, who enthusiastically began to inquire whether my dissertation was complete two weeks after I began the program. My mother, Audre Hopkins, and brother, Blake Hopkins, are essential models for the person I am today, as were my maternal grandparents, Thomas and Lorraine Dewey. Without their influence, I could not have traveled this path. My child, Kim Chinle Hopkins-Partington, and stepkids, Mike P. and Michelle P. Marker, were extraordinarily flexible and cooperative during my tenure at Virginia Tech, and kept crises to a minimum. There exist no adequate words to express my appreciation to these family members.

I owe a debt of gratitude to my University of Utah advisor, Ronald Bruhn, and to my co-students Terry Pavlis, Laura Serpa, Sharon Alley, Larry Guth, and Susan Beck, for building an earth scientist out of a zoologist and manual laborer. Utah provided an excellent foundation in the geological and geophysical sciences.

No mentor could have been more enthusiastic about my thoughts and opinions than my advisor, John Costain. He and my committee provided me a superb, broad spectrum education that will serve me well in academic or industry pursuits. Martin Chapman, Matt Sibol, Chris Powell, and Gil Bollinger provided generous support for the earthquake work. Accolades are due my fellow graduate students at Tech. I profited from illuminating discussions with many geologists and geophysicists, reaffirming what we all know: students learn as much from each other as they do from formal classes. I owe special gratitude for input from the people I dragged into my office to show off my data, including Krishna Sinha, Lynn Glover, Rick Law, Janet Schweitzer, Dave Valentino, Mike Pope, Aus Al-Tawil, Chris Fedo, Cole Davison, Will Orndorff, Pat Jenks, and Bill Domoracki. Additional thanks are due to Arthur Snoke, Rick Law, Lynn Glover, Bob Bodnar, Bob Tracy, Barbara Bekken, Bill Henika, and Paul Ribbe for extended educational discussions which occasionally bore on the earth sciences.

My years at Tech would have been impoverished without my house-mates. Llyn Sharp, Charles Oakes, Ron Sheets, and Sam Peavy formed a priceless support group during trying times. Friends Barb Munn and Steve Miller kept me properly caffeinated in our last year together, and buddy Ron Wirgart went far beyond the call of duty on presentation slides.

Mildred Memitt, unofficial committee member, provided countless hours of instruction on the use of the DISCO seismic data processing system, and Bob Montgomery and John Wonderley kept the VAX and PC computer systems alive. The office staff, Linda Bland, Mary McMurray, Kathy Shelor, and Carolyn Williams, deserves accolades for wonders worked, and Karen Hunt continues to perform miracles for all of us. Mark Fortney deserves thanks for his invaluable help with the defense slides.

Funding and data for this research were provided by ARCO, Amoco, Chevron, Mobil, Texaco, and the Society of Exploration Geophysicists.

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Chapter 1: Introduction

The source of the New York-Alabama Magnetic Lineament (NYAML) has been the topic of speculation since the Lineament was discovered (King and Zietz, 1978). This paper is concerned with the integration of newly reprocessed reflection seismic data with potential field data and models across the Lineament in eastern Tennessee to determine the nature of the source (Figure 1.1). In addition, this study investigates the relationships of features observed on the reflection data to geophysical phenomena described in Canada in the vicinity of the Grenville Front and to those reported within the United States near the NYAML. The paper proposes models for the development of the Lineament and speculates about its relationship to the Grenville Front.

Throughout this manuscript, the term Grenville basement will follow the recommendation of Moore (1986) and refer to rocks deformed during the Grenville orogenic event of 1.1 Ga. References to the Grenville Front will relate to the northwestern or western limit of deformation associated with orogenesis whether or not the deformed units have affinity with the North American craton. This convention is chosen because the lithologies of the autochthonous basement beneath eastern Tennessee are not constrained by outcrop, are poorly constrained by drill holes or xenoliths, and the nearest exposures of Precambrian crystalline rocks are allochthonous and have been transported to the northwest significant distances (Hatcher, 1984). Hence, the relationship between the autochthonous rocks and the ancestral North American craton cannot be unequivocally determined. The Grenville basement beneath eastern Tennessee will be referred to here as autochthonous, with the recognition that it has been termed parautochthonous by Canadian workers (e.g. Moore, 1986) interested in distinquishing between the Grenville Province and the Archean - Proterozoic North American Craton.

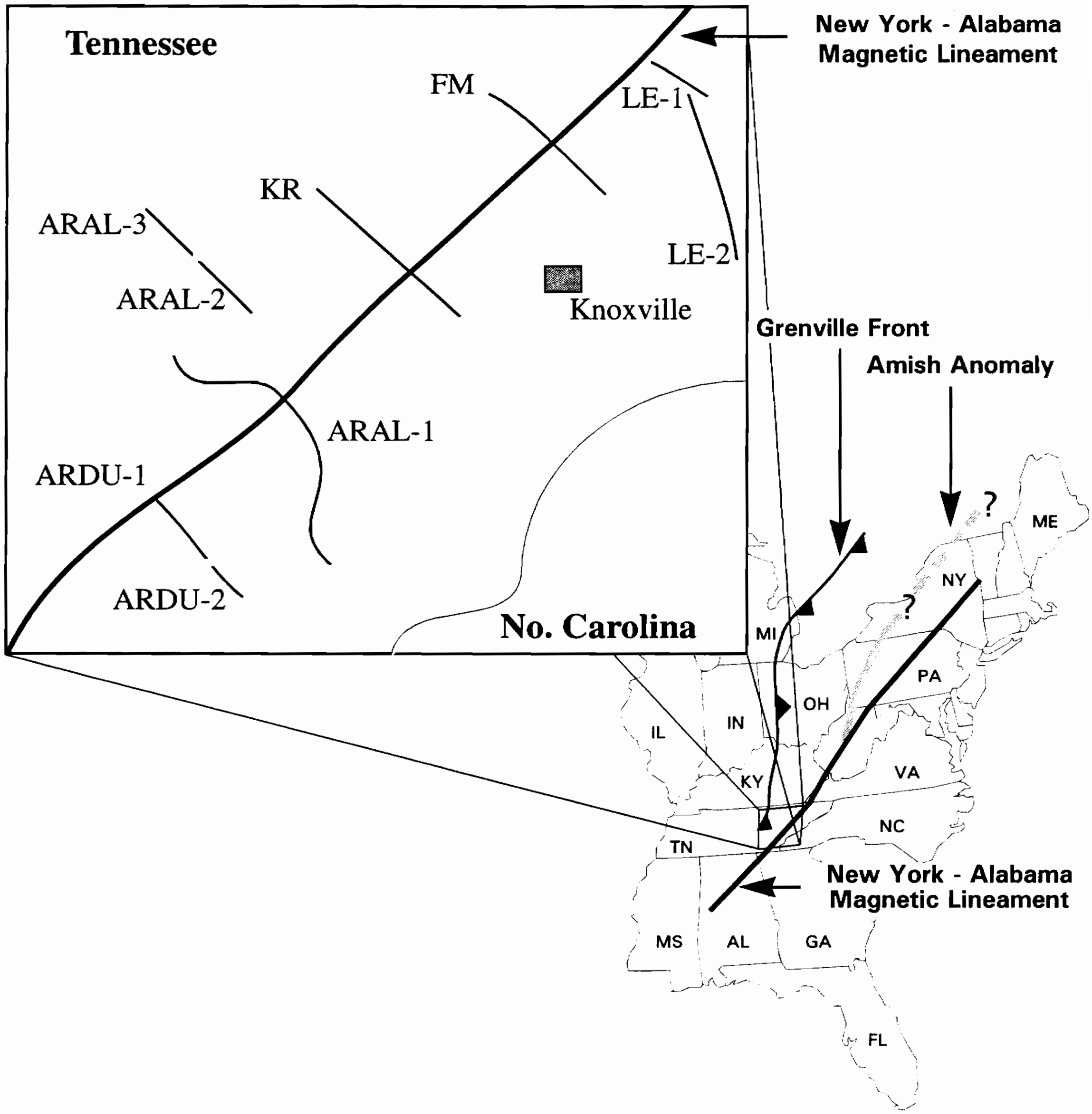


Figure 1.1. Index Map: King and Zietz (1978) interpreted the New York - Alabama Magnetic Lineament (black line) as extending into the central part of New York. The Amish Anomaly (gray line) is shown as an alternate location. The Grenville Front (black line with teeth) is interpreted in the United States on the basis of potential field data and wells. The insert of the study area in eastern Tennessee shows the reflection data reprocessed for this investigation. The data were chosen for their perpendicular orientation across the Lineament and for their length, and image important subsurface structures that lead to new interpretations for the crystalline basement in eastern Tennessee.

Magnetic and gravity signatures

The New York-Alabama Magnetic Lineament shown on Figure 1.2 was recognized by King and Zietz (1978) from aeromagnetic maps over the eastern United States. They described the feature as extending northeast for 1600 km from the Mississippi embayment in Alabama to the Green Mountains of New York. King and Zietz indicated that the Lineament bears a strong correlation with the gravity data in the eastern United States. They pointed out that the NYAML lies to the west of the gravity low (Figure 1.3) that dominates eastern North Carolina and Virginia, and that it separates northeasterly trending gravity ridges southeast of the anomaly from more northerly trending gravity ridges to the northwest.

The New York - Alabama Magnetic Lineament is characterized by a steep gradient with relief of as much as 3200 nanoteslas in eastern Tennessee. The linearity of the feature prompted King and Zietz to propose that the signature recorded the presence of a major strike slip feature associated with continental collision. On the basis of the wavelengths of the features, Culotta et al. (1990) suggested that the New York-Alabama Magnetic Lineament could be extended along the Amish anomaly in western New York instead of the more easterly position chosen by King and Zietz. The westerly location of the Amish anomaly aligns the Lineament with the Central Metasedimentary Belt of the Grenville Province in Canada, west of the Adirondack Mountains of New York. The King and Zietz location places it along the southeastern border of the Adirondack Mountains.

A string of small lows extends along the New York-Alabama Magnetic Lineament in eastern Tennessee northeast of the regional gravity low in North Carolina (Figure 1.3). West of the gradient forming the NYAML, the gravity field is characterized by a pronounced gravity ridge that trends north-south into central Kentucky. Termed the East-continent Gravity High by Bryan (1975), the ridge corresponds to an area of high magnetic values.

Figure 1.2. Magnetic Map of eastern North America: Portion of the Magnetic Anomaly Map of North America showing the signatures of the New York - Alabama Magnetic Lineament, the Mid-continent rift system signature, and the Grenville Front. Magnetic highs are shown in red and lows are shown in blue. The Amish Anomaly is shown in western New York. (D.N.A.G. Project map, Geological Society of America, 1987).

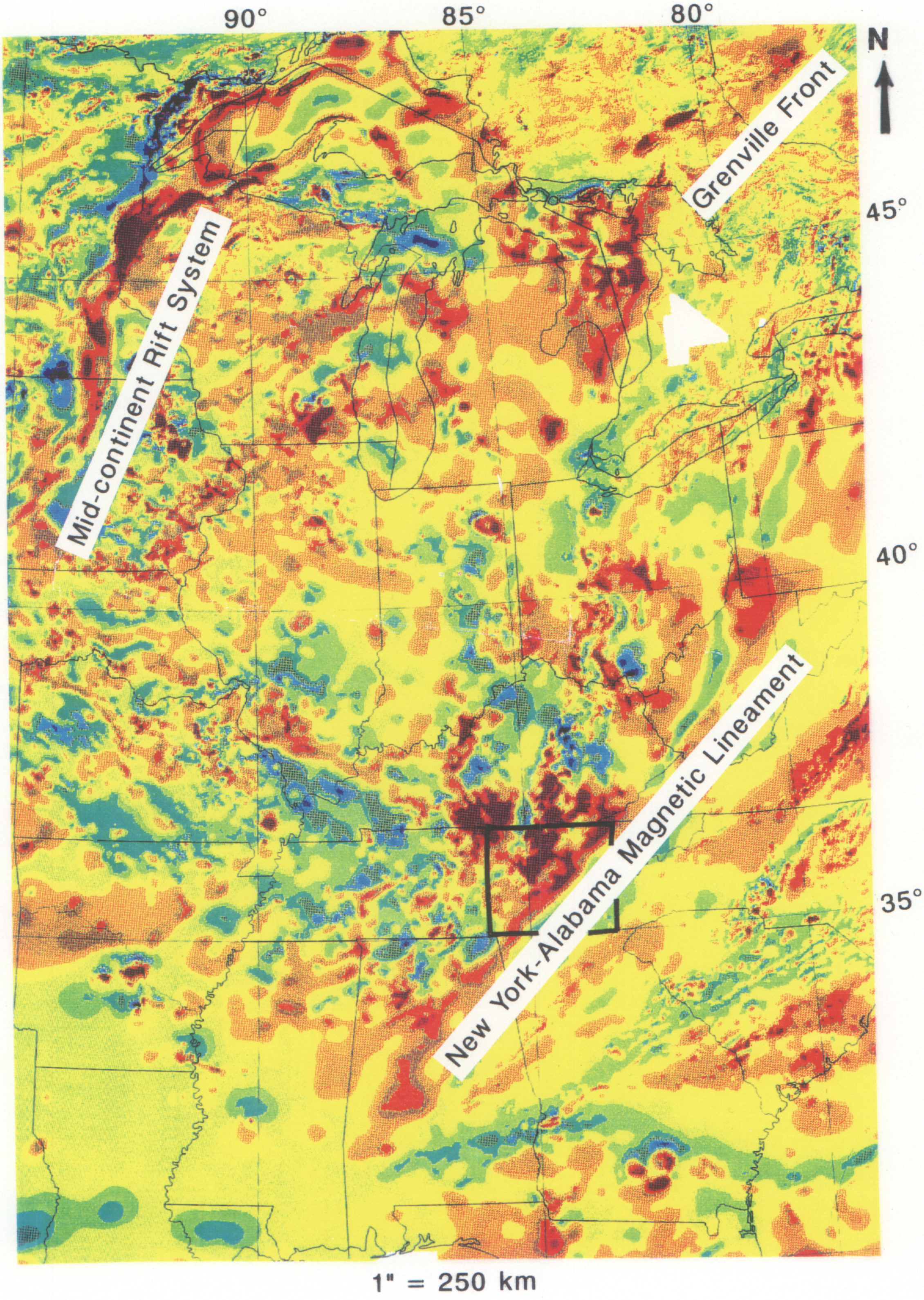
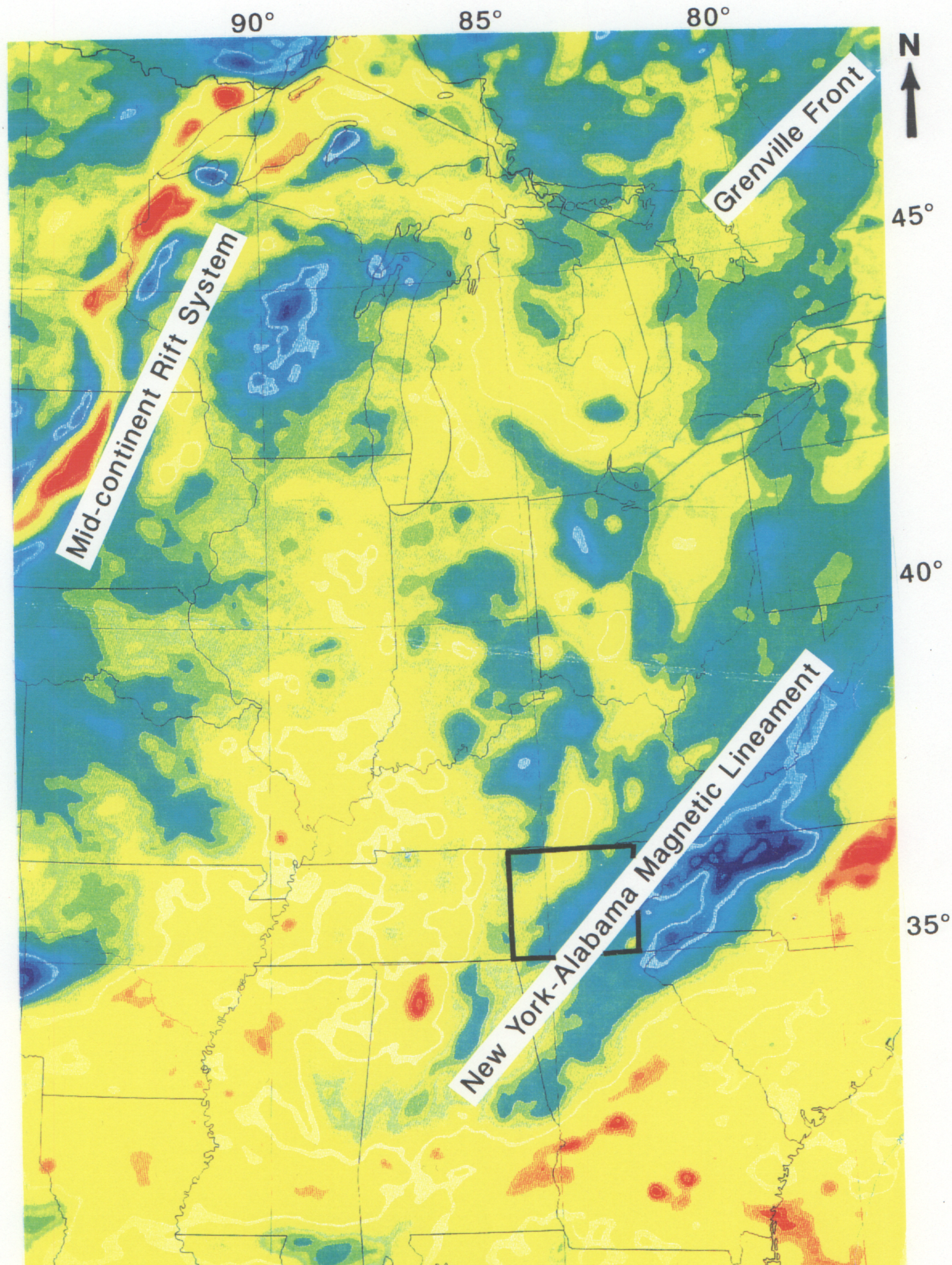

90°
85°
80°
N
Grenville Front
45°
Mid-continent Rift System
40°
New York-Alabama Magnetic Lineament
35°
1" = 250 km

Figure 1.3. Gravity Map of eastern North America: Portion of the Gravity Anomaly Map of North America showing the locations of the New York - Alabama Magnetic Lineament, the Grenville Front, and the Mid-continent Rift System. Gravity highs are shown in red and lows are shown in blue. The position of the Amish Anomaly is shown in western New York. (D.N.A.G. Project map, Geological Society of America, 1987).


90°
85°
80°
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45°
40°
35°
Grenville Front
Mid-continent Rift System
New York-Alabama Magnetic Lineament
1" = 250 km

The region northwest of the New York - Alabama Magnetic Lineament in the southern Appalachians is typified by a group of short wavelength anomalies extending from western Ohio southward through central Kentucky and into northern Tennessee. King and Zietz (1978) described the anomalies as trending more or less north-south. In east-central Kentucky, these relatively discrete anomalies are accompanied by flanking lows along the north. Such pairing suggests that the anomalies are produced by features that might be relatively small, isolated bodies in comparison with the deeper source of the New York-Alabama Magnetic Lineament.

Possible sources

Several late Paleozoic or Mesozoic peridotite and kimberlite dikes have been reported in scattered locations from Tennessee to New York (Figure 1.4). Reports of mafic igneous bodies at the surface throughout the eastern United States have been summarized by Jachens et al. (1989). Parrish and Lavin (1982) discuss surface exposures of kimberlites in Pennsylvania. The Norris Lake kimberlite in eastern Tennessee is described by Johnson (1961), Meyer (1976), and Zartman et al. (1967) and is proposed to be Mississippian-Permian in age, while the intrusions in Pennsylvania are younger than the Alleghanian orogeny (Parrish and Lavin, 1982; Dennison, 1983) The presence of mafic intrusions suggests a possible source for the positive gravity and high magnetic susceptibility observed in the potential field maps northwest of the gradient of the NYAML in eastern Tennessee.

Xenoliths reported from a teschenite dike in Virginia (Johnson et al., 1971) were described as probably of basement origin. Xenoliths present in the dike were reported to be of compositions consistent with the presence of gneissic or granitic basement. Although the dike is located hundreds of kilometers northeast of the study area, it is approximately along the strike of Alleghanian structures and could contain samples representative of the Tennessee autochthon.

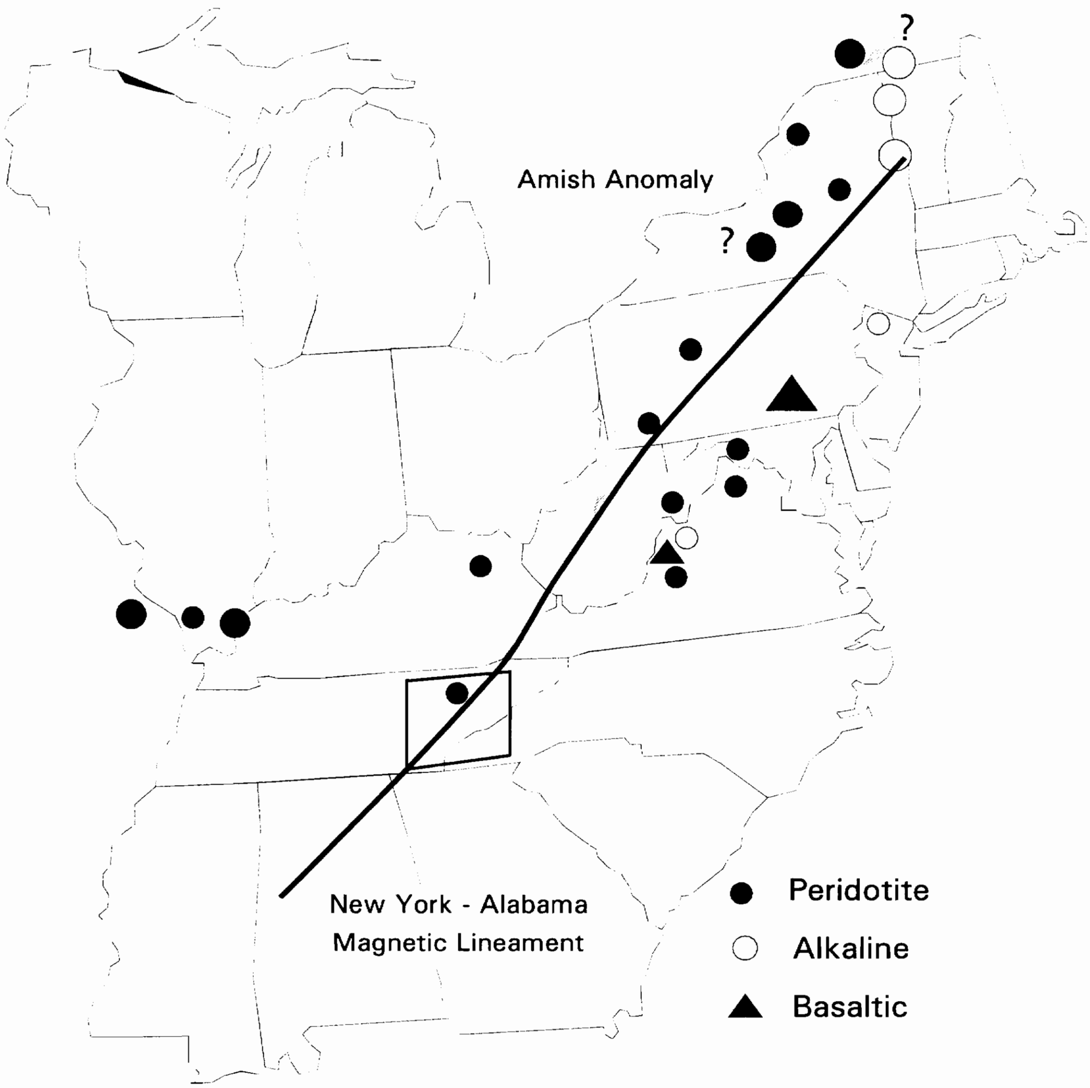


Figure 1.4. Dikes: Dikes mapped at surface in the vicinity of the New York - Alabama Magnetic Lineament suggest the presence of sub-surface candidates for localized magnetic and gravity sources. The dikes range in composition from peridotitic to alkalic. The dike in eastern Tennessee is located at Norris Lake. Larger symbols indicate multiple dikes.

A recent evaluation of drill cuttings in the Omaha oil field in southern Illinois confirmed the presence of magnetically susceptible alnöite intrusive sills totalling 72 m in thickness (Sparlin and Lewis, 1994). The sills are described by Sparlin and Lewis as mantle-derived ultramafic intrusions with modal analyses indicating 9% primary magnetite by volume.The authors assigned an intrusive age of late Paleozoic (260 m.a.) to the ultramafic sills based on K-Ar dating done by previous workers (e.g. Zartman et al., 1967; Bickerman et al., 1982; Lewis and Mitchel, 1987) of exposures of igneous rocks in Illinois, Kentucky, and Missouri.

The region characterized by discrete magnetic anomalies widens to the northeast into central Ohio to western Lake Erie. North of the lake the Grenville Front is drawn by Green et al. (1988) on the basis of reflection seismic data through the easternmost part of Lake Huron, through Georgian Bay, and into Canada where it correlates with a magnetic gradient trending northeast - southwest. In contrast to the magnetic signature in the U.S., where higher magnetic values fall southeast of the Grenville Front, in Canada the high values are located to the northwest of the Front. The specific relationship between the anomalies and exposed lithologies has not been determined.

A gravity low is coincident with the position of the Grenville Front magnetic high in Georgian Bay, Lake Huron. The overall trend of gravity ridges is northeast and parallel to the ridges on the magnetic map, but the gravity data does not delineate the Grenville Front in Canada as well as does the magnetic data.

Grenville basement exposures in Canada and the eastern United States

The distribution of Grenville inliers in the eastern United States is shown in Figure 1.5. Although the relationship of the crystalline basement in eastern Tennessee to nearby allochthonous exposures of Grenville rocks is speculative, structures and metamorphic conditions might be representative of conditions in the basement of eastern Tennessee.

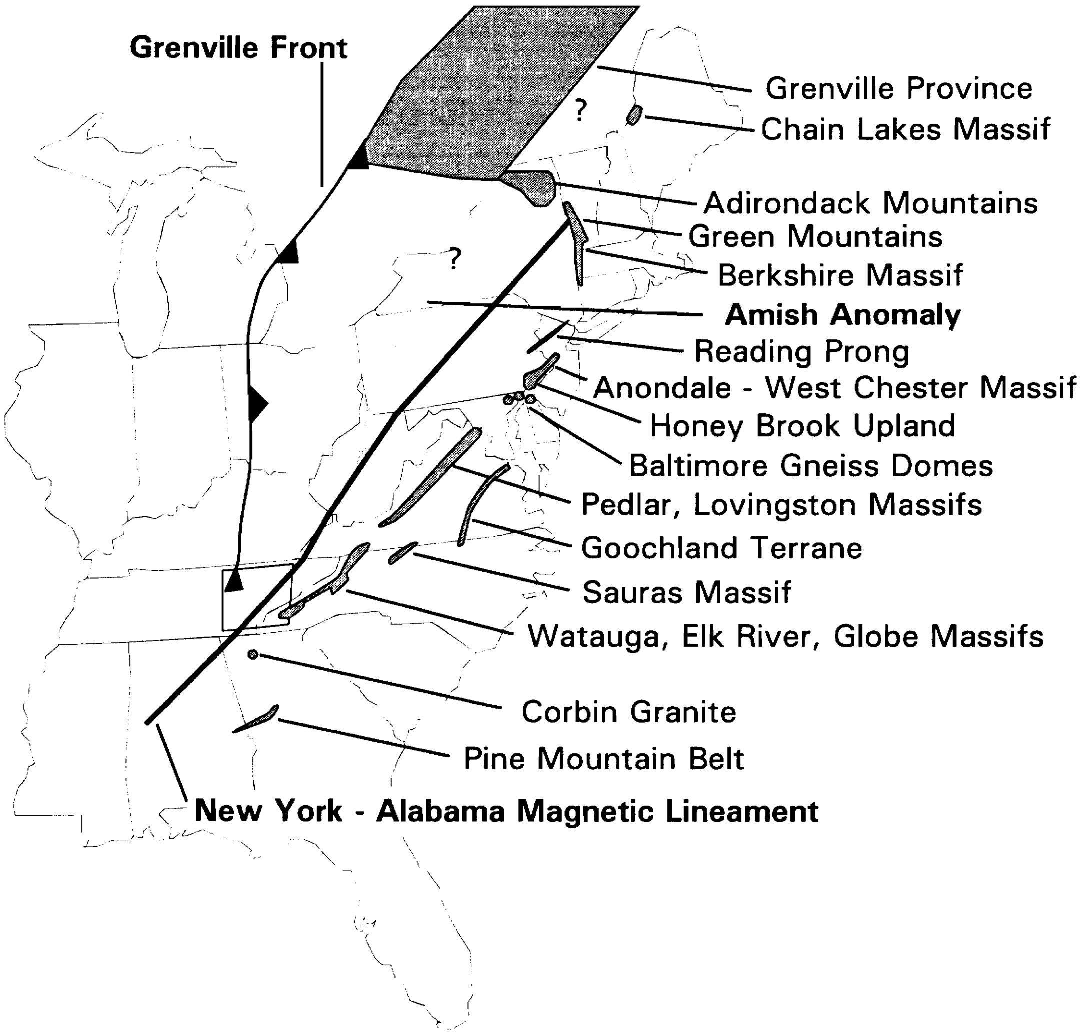


Figure 1.5. Grenville inliers: Outcrops of Grenville basement are shown with the New York - Alabama Magnetic Lineament, Amish Anomaly, and the Grenville Front. The exposures lie mainly to the southeast of the Lineament south of New York, but all lie to the southeast if the northerly trend of the Lineament is chosen along the Amish anomaly. The box in eastern Tennessee delineates the study area.

Grenville outcrops south of New York fall to the east of the New York - Alabama Magnetic Lineament (black line), while the Adirondack Mountains, the Green Mountains, and the Chain Lakes Massif fall to the west. Placing the location of the NYAML along the Amish Anomaly (gray line) as proposed by Culotta et al. (1990), suggests that all of the inliers in the United States lie to the east of the Lineament. If the NYAML indicates the presence of a significant crustal boundary, then the positions of the outcrops are significant in determining the tectonic setting for the source of the Lineament.

The Grenville terrane in Canada

An examination of the seismic reflection expression of sub-provinces within the Grenville province (Figure 1.6) aids in the interpretation of the eastern Tennessee data. The Grenville province in Canada consists of the Grenville Front, the Grenville Front Tectonic Zone, the Central Metasedimentary Belt, the Central Gneiss Belt, and the Central Granulite Terrane. The units east of the Grenville Front Tectonic Zone apparently have no affinity with the North American craton and have been interpreted to be accreted microterranes containing evidence of older deformation episodes (Moore, 1986). Reflection seismic data (Figure 1.7) were obtained in Lake Huron by the of GLIMPCE (Great Lakes International Multidisciplinary Program on Crustal Evolution) consortium. The data reveal the Grenville Front Tectonic Zone (GFTZ) (Green et al., 1988), which is characterized by strong east-dipping reflections that extend from the surface to at least 9 seconds (30 km). The reflections have been interpreted by Green et al. as originating from a region of intense ductile strain that penetrates the entire thickness of the crust. Green et al. suggested that the zone correlates with thick mylonites mapped within the exposed GFTZ and that reflectivity is associated with velocity contrasts at highly strained contacts.

The boundary between the Central Gneiss Belt and the Central Metasedimentary Belt has been imaged on reflection data by Milkereit et al.

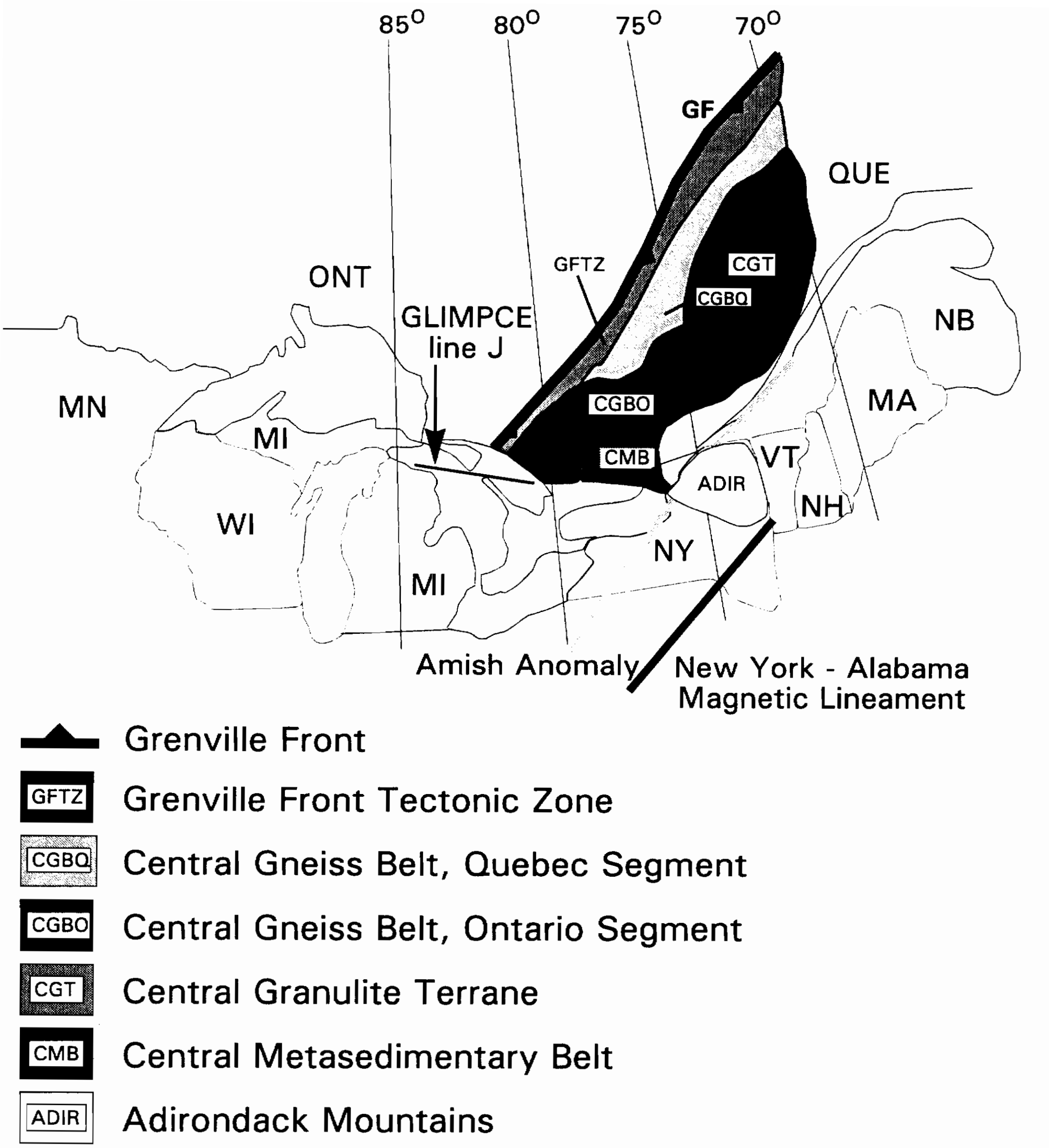


Figure 1.6. Grenville Province in Canada: The Grenville Front (GF) bounds the region to the west and forms the western margin of the Grenville Front Tectonic Zone. Tectonic units shown are those of interest to this study. The New York - Alabama Magnetic Lineament (black line) falls along the southeastern margin of the Adirondack Mountains. The Amish Anomaly (gray line) is an alternative location for the Lineament and can be extended along the western edge of the Adirondack Mountains. The location of GLIMPCE reflection line J is shown in Lake Huron. Modified from Moore (1986)

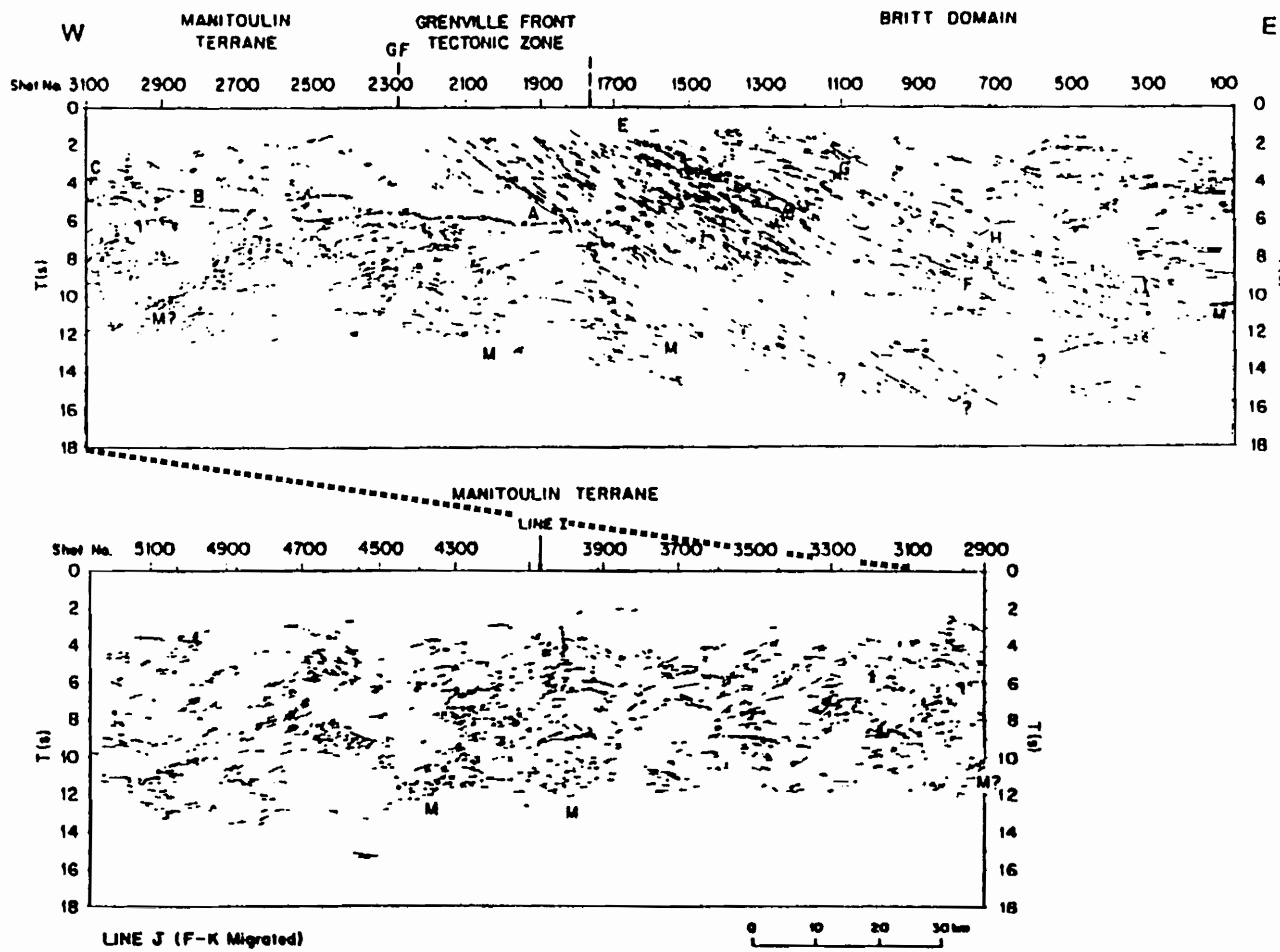


Figure 1.7. GLIMPCE line J: The strong east-dipping reflections delineate the subsurface expression of the Grenville Front Tectonic Zone. The western margin of the zone forms the Grenville Front. The reflections are associated with mylonite zones and extend to at least 9 seconds. The dotted line indicates the position at which the line sections connect. (from Green et al., 1988).

(1992), who interpreted the contact as a terrane boundary. They based the interpretation on the presence of east-dipping reflections that extend to middle crustal depths and dip 15° to 30°. The reflections were interpreted to arise from a mylonite zone separating the two belts. Exposures of the structurally lower and older Central Gneiss Belt contain evidence of polycyclic metamorphism predating the Grenville orogenic episode by >1.0 Ga. (Davidson, 1986) while the Central Metasedimentary Belt contains metasupracrustal and plutonic rocks younger than 1.3 Ga. (van Breemen, et al., 1986). Thus, the latter zone appears to have been affected only by Grenville orogenesis.

The Central Metasedimentary Belt was emplaced along a series of ductile shear zones from southeast to northwest to its present position against the Central Gneiss Belt (Davidson, 1986).

Adirondack Mountains of New York

The Adirondack Mountains of New York provide information about the character of the Grenville basement in the United States. The Adirondacks have been affected by at least 5 phases of folding, as well as low angle ductile faults including the Carthage - Colton mylonite zone reported to have been deformed by the F3 stage of folding (McLelland and Isachsen, 1986). The northwest dipping mylonite zone extends for 110 km and ranges from a few meters in width to over 5 km. The zone has been described as separating the lithologically distinct highlands comprising the bulk of the Adirondack Mountains from the lowlands to the northwest. McLelland and Isachsen did not indicate whether or how much movement might be associated with the mylonite zone, but indicated that it represents a major Grenville structure.

Southeastern United States exposures

The composition and metamorphic grade of the thick crust beneath eastern Tennessee is poorly constrained, but is assumed here to be of Grenville age and metamorphosed to at least amphibolite and probably granulite grade based on exposures of Grenville rocks to the east and northwest of the study area. Grenville basement appears throughout the eastern United States in thrust sheets within the Blue Ridge province in Virginia and North Carolina. The exposed Grenville rocks nearest the study area, including the Elk River, Globe, and Watauga Massifs near the Tennessee - North Carolina border at Roan Mountain, contain mineral assemblages and relict textures indicative of Grenville-aged metamorphic grade ranging from amphibolite to granulite grade (Bartholomew and Lewis, 1984). The massifs were described as layered gneisses into which Grenville - aged intrusive complexes were emplaced. Rankin (1975) reported that the complexes in the vicinity of eastern Tennessee were intruded again in a younger late Precambrian episode unrelated to Grenville orogenesis.

Outcrops of crystalline basement nearest to the study area include the Carver's Gap gneiss and the Cloudland gneiss of the Elk River Massif on the Haysville thrust sheet. Bartholomew and Lewis (1984) proposed a minimum displacement of 50 km for the Elk River Massif containing the gneisses on the Linville Falls fault. The Carvers Gap gneiss and the Cloudland gneiss were metamorphosed to granulite facies in the Precambrian at P-T conditions of 755^{o} to 845^{o} C and 6 to 8 km according to Monrad and Gulley (1983). Monrad and Gulley determined the whole rock Rb-Sr age of the Carvers Gap Gneiss to be about 1.815 Ga., and interpreted its formation as a mid-Proterozoic orogenic event related to either metamorphism of the continental crust or to formation of crustally derived igneous material. The younger Cloudland gneiss was dated at 0.807 Ga. and, according to Monrad and Gulley, could reflect isotopic homogenization during the later stages of Grenville metamorphism. Wilcox and Poldervaart (1958) indicated that both gneisses are intruded by diabase dikes of the Bakersville gabbro. Rankin et al. (1973) and Rankin (1975) linked these dikes with the Crossnore Plutonic-Volcanic group and suggested that they were intruded during opening of the Iapetus Ocean during the late Precambrian.

The Elk River Massif has probably been transported farther to the northwest than the nearby Watauga and Globe Massifs (Bartholomew and Lewis, 1984). Therefore the compositions of the latter two massifs might be a better indication of the composition of the crystalline basement beneath eastern Tennessee than the rocks in the Elk River Massif. According to Bartholomew and Lewis (1984), the Watauga Massif contains country rocks composed of layered mesogneisses and meso-granulite gneisses intruded by Grenville-aged granitoids, biotite granitoids, and biotite dioritoids, all of which have been intruded during the late Precambrian by granitoids and dioritoids. The Globe Massif includes layered mesogneiss country rock intruded by dioritoids, porphyritic dioritoids, and granitoids, all of which have been intruded by younger late Precambrian granitoids (summarized in Bartholomew and Lewis, 1984).

To the northeast of eastern Tennessee and along strike of the study area, Sinha and Bartholomew (1984) described the Pedlar Massif as composed of high grade metamorphic gneiss (the Lady's Slipper Granulite Gneiss) intruded by charnockitic rocks (the Pedlar River Charnockite Suite). They used U/Pb dating to determine the intrusive age of about 1.075 Ga. for the charnockite into 1.13 Ga. gneiss. Bartholomew and Lewis (1984) described the Lady's Slipper Granulite Gneiss as a deep granulite grade rock. The Lovingston Massif, at granulite to amphibolite facies, is slightly lower in metamorphic grade than the Pedlar Massif, while several rocks of Grenville age exposed south of the Pedlar and Lovingston blocks are at a minimum amphibolite metamorphic grade.

Based on the exposures of basement in the southeast United States, and particularly on the compositions of the structurally nearest Globe, Watauga, and Pedlar Massifs, the rocks in the basement of eastern Tennessee are interpreted to be of at least amphibolite metamorphic grade. The basement most likely contains gneisses intruded by Grenville-aged granitoids and dioritoids comparable to those found in the exposed massifs. Therefore, interpretations of the lithologies and metamorphic grades beneath the Valley and Ridge province of eastern Tennessee will be consistent with the compositions and grades in the Watauga, Globe, and Pedlar Massifs.

Structure of the crust

Crustal velocities and structures have been interpreted by a number of authors for the eastern United States (Figure 1.8).

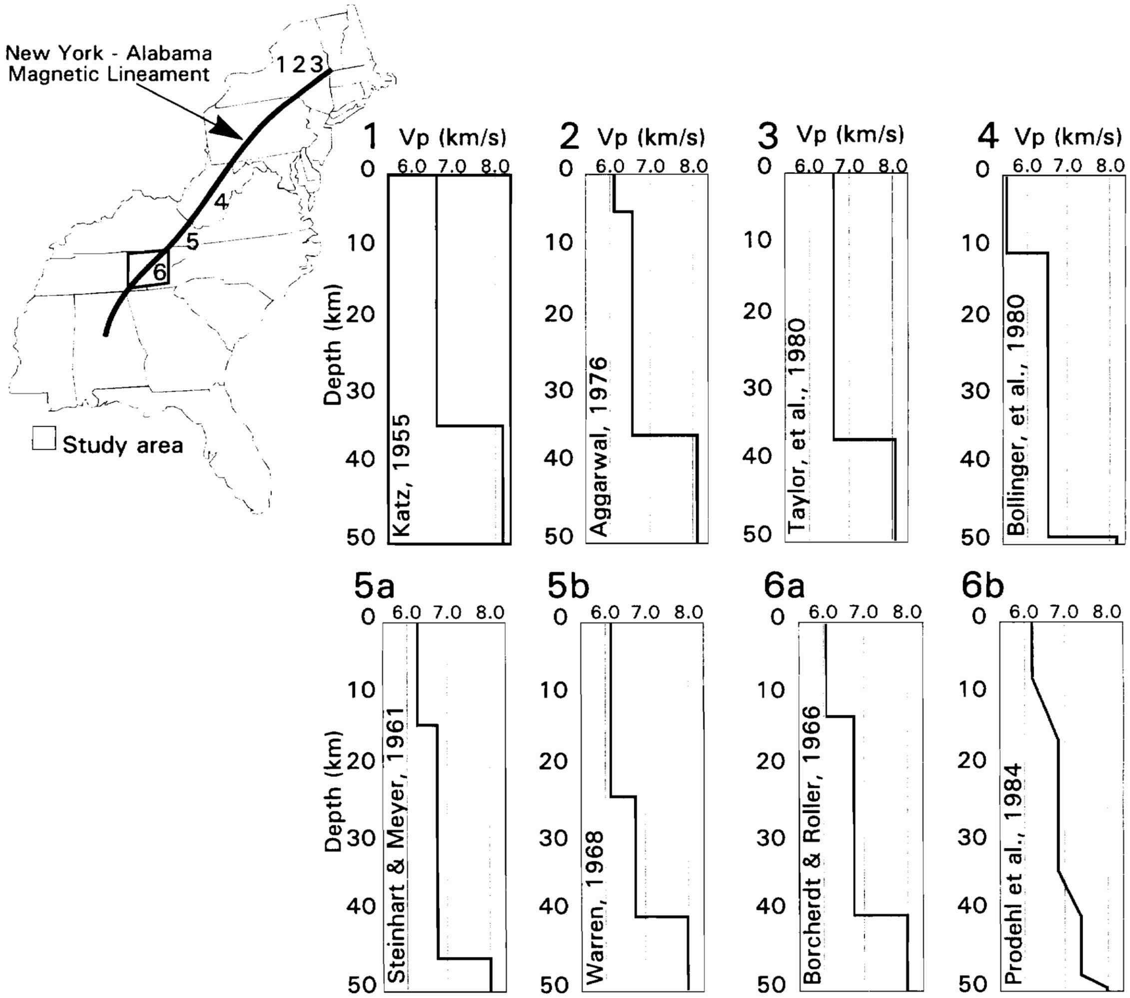


Figure 1.8. Velocity models: Reversed refraction profiles yield P-wave velocity models in the eastern United States and indicate the presence of thick crust in the southern Appalachians including eastern Tennessee. Modified from Taylor, 1989.

Prodehl et al. (1984) reinterpreted reversed seismic refraction data originally analyzed by Borcherdt and Roller (1966) and determined that the crust is up to 47 km thick in eastern Tennessee. The models corroborate the presence of thick crust in the southeastern Appalachians (Taylor, 1989) (Figure 1.9) including eastern Tennessee. The Prodehl model indicates the presence of gradational velocity changes between the upper and middle crust (7 to 13 km), between the middle to lower crust (32 to 40 km), and between the lower crust and mantle (47 to 49 km). The upper crustal velocities are fairly high at 6.1 km/sec for a 7 km surface layer in eastern Tennessee (Prodehl et al., 1984) when compared with those determined approximately along strike in western Virginia by Chapman (1979) and Bollinger et al. (1980), where they have been estimated at 5.6 km/sec for a 10 km surface layer. The transition in velocities between eastern Tennessee and western Virginia might support the observation by Prodehl et al. that the principle lateral velocity contrasts in that study were observed perpendicular to the strike of the surface structures.

Taylor (1989) indicated that the velocities determined from the refraction work in eastern Tennessee match those obtained by Owens et al. (1984) from receiver functions. This thickness estimate complements the results of James et al. (1968), who suggested that the crust is approximately 46 km thick in the study area based on time-term analyses from the East Coast On-Shore Off-Shore Experiment (ECOOE) in 1965.

Position of the Grenville Front in the southeastern United States

The position of the Grenville Front in the southeastern United States is constrained by previous workers on well data, gravity and magnetic data, and reflection profiles (e.g. Lidiak et al., 1966; Van Schmus and Hinze, 1985; Lucius and Von Frese, 1988; Drahovzal et al., 1992). Extending the Grenville Front into eastern Tennessee (where few wells penetrate the crystalline basement) places its location between high magnetic plateaus at approximately

85° W longitude at the Kentucky - Tennessee border. This position indicates that the source of the New York-Alabama Magnetic Lineament falls within crystalline basement affected by Grenville deformation.

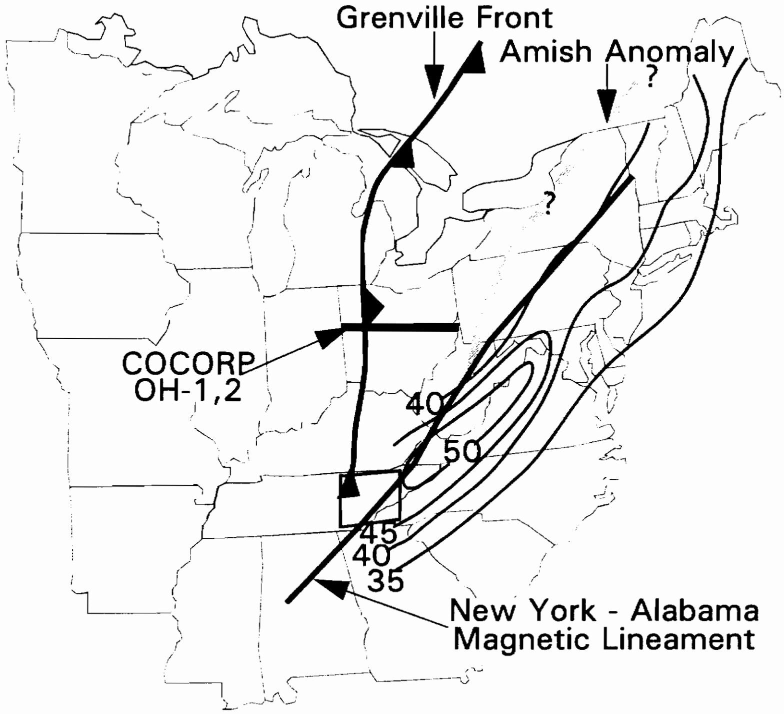


Figure 1.9. Crustal thickness: The thickness map of Taylor (1989) shows the deep crust interpreted for the southern Appalachian region. The New York - Alabama Magnetic Lineament trends subparallel to the axis of maximum crustal thickness. The box in eastern Tennessee outlines the study area. Contours are in kilometers. COCORP line OH-1,2 is shown as a dark line in central Ohio. (modified from Taylor, 1989).

Reflection data from Ohio (Pratt et al., 1989) are shown in Figure 1.10. These COCORP data indicate that west-dipping reflections pervade the crust in eastern Ohio. The authors assigned the west-dipping reflections to the Central Metasedimentary Belt (CMB) of the Grenville Province. Reflections dipping west at about the same angle are pronounced on reflection data in Tennessee (this paper). Possible relationships between the crust in eastern Tennessee and Ohio are discussed in the following chapters.

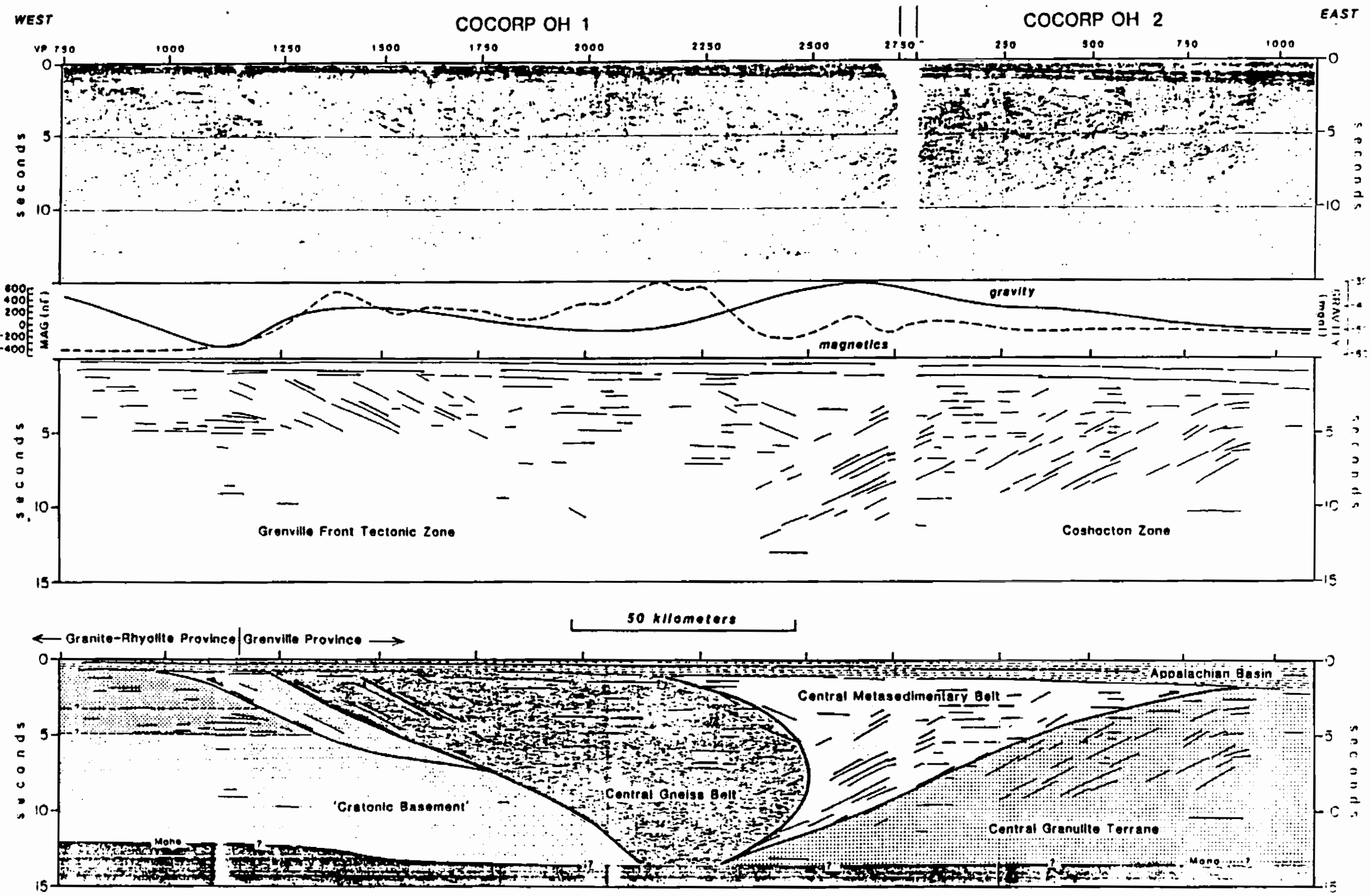


Figure 1.10. **COCORP line OH-1,2**: The reflection data in central Ohio obtained by COCORP indicate that the eastern part of the crust beneath 2 seconds is characterized by strong west-dipping reflections. The reflections were assigned to the Central Metasedimentary Belt and Central Granulite Terrane by Pratt et al. (1989). The west-dipping reflections are important in the correlation of terranes with those observed in eastern Tennessee (this study). (from Pratt et al., 1989).

Well data west and north of the study area indicate the presence of volcanics beneath the shelf strata (Figure 1.11). In Tennessee, King (1990) refers to three wells reportedly sampling rhyolites, diabases, and troctolites of Precambrian ages at depths ranging from 1420 to 2360 m below sea level. Wells in Kentucky and Ohio approximately along strike of the projected Grenville Front sampled rhyolite, basalt, troctolite, andesite, and lithic arenite of Precambrian ages at depths ranging from 995 m to 1705 m (Drahovzal et al., 1992). The volcanics described in Ohio and Kentucky by Drahovzal et al. were assigned to the Middle Run Formation and were interpreted by those authors to pre-date the Grenville orogeny.

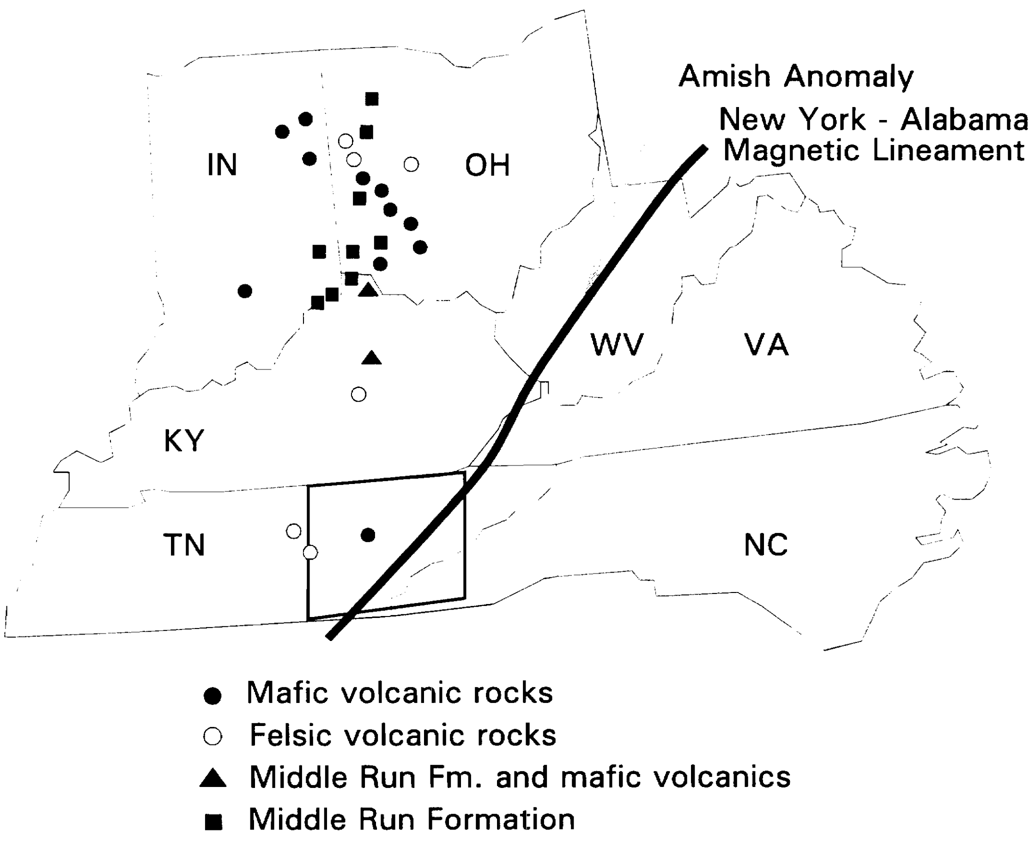


Figure 1.11. Wells into Precambrian rock: Although the wells sample Precambrian rocks, they do not extend deep enough to yield lithologic information about the crystalline basement in the study area (box). Well data from Drahovzal et al., 1992, and from King, 1990.

Summary

The New York - Alabama Magnetic Lineament is associated with gravity lows in the study area, and is flanked on the west by more isolated magnetic anomalies associated with the East Continent Gravity High. Northward, the Lineament extends into New York along the eastern flank of the Adirondack Mountains (King and Zietz, 1978).

Outcrops in the Grenville Province of Canada and in the Adirondack Mountains of New York indicate the presence of zones of intense ductile strain and thick mylonites. The Grenville Front Tectonic Zone, characterized by mylonite zones, is imaged on GLIMPCE reflection data as strong reflections extending from the surface to at least 9 seconds (27 km) and dipping 25° to 35°. Similar reflections have been reported along the contact between the Central Gneiss Belt and the Central Metasedimentary Belt on industry data. West dipping reflections have been reported on COCORP data in Ohio.

Thick crust is characteristic of eastern Tennessee, where crustal thicknesses of up to 47 km are indicated by refraction data. The axis of maximum thickness is subparallel to the New York - Alabama Magnetic Lineament throughout the eastern United States.

The nearby Watauga and Globe Massifs and the along-strike Pedlar Massif include gneisses, granitoids, and dioritoids at upper amphibolite to granulite grade. These outcrops are inferred to be representative of the grades and compositions of the crystalline basement in eastern Tennessee.

Chapter 2: Reflection Data

The available vibroseis seismic reflection data in the vicinity of the New York - Alabama Magnetic Lineament (NYAML) were acquired to image potential hydrocarbon reservoirs within the Valley and Ridge structural province (Figure 2.1). The longest lines and lines shot as a set were selected for this study. Those lines that extended perpendicular to and across the steep gradient of the NYAML were preferred. If the New York - Alabama Magnetic Lineament represents a crustal feature, then the lines chosen for this analysis are oriented most favorably for imaging boundaries in the crystalline rock that might delineate the source of the Lineament.

The nine separately acquired and processed seismic lines form six profiles. Profile ARAL (Figure 2.2) is composed of lines ARAL-1, ARAL-2, and ARAL-3 and is the longest profile at 76 km. To form the profile, line ARAL-1 was projected 10 km to the northeast along the strike of Valley and Ridge structures to connect to the southeastern end of ARAL-2. Lines ARAL-2 and ARAL-3 are in line and separated by 500 meters (a river) and required no significant projection. Because of its length and excellent quality, profile ARAL is featured throughout most of the discussion. Figure 2.3 illustrates a close-up view of ARAL-2 and ARAL-3. ARDU-1 and ARDU-2 were projected across a river to produce profile ARDU (Figure 2.4). Profile KR (Figure 2.5) consists of line KR-2, and line FM-1 forms profile FM (Figure 2.6). Lines LE-1 (Figure 2.7) and LE-2 (Figure 2.8) are treated as separate profiles because of their different orientations, where line LE-1 strikes sub-parallel to the profiles to the south (northwest to southeast) and line LE-2 strikes approximately north-south.

The acquisition and processing parameters used for these profiles are given in Appendices 1 and 2, respectively. The decision to use unmigrated data also is discussed in Appendix 2. In the conversion from seconds to

Figure 2.1. Index map of reflection data: The nine lines form 6 profiles that trend perpendicular to the strike of Valley and Ridge thrust faults on the Geologic Map of Tennessee (modified from Hardeman, 1966). The New York - Alabama Magnetic Lineament (NYAML) is indicated by the dark line drawn near the midpoint of the steep gradient (e.g. Figure 4.1). The NYAML strikes approximately parallel to the thrust faults. Therefore, the reflection data, collected perpendicular to the thrusts, is appropriate for imaging structures in the crust that might be responsible for the Lineament. Lines ARAL-1, ARAL-2, and ARAL-3 were composited to produce profile ARAL, with line ARAL-1 projected along the strike of the Valley and Ridge structures to connect to the southeastern end of line ARAL-2. Lines ARDU-1 and ARDU-2 form profile ARDU. The profiles northeast of ARAL are composed of single lines.

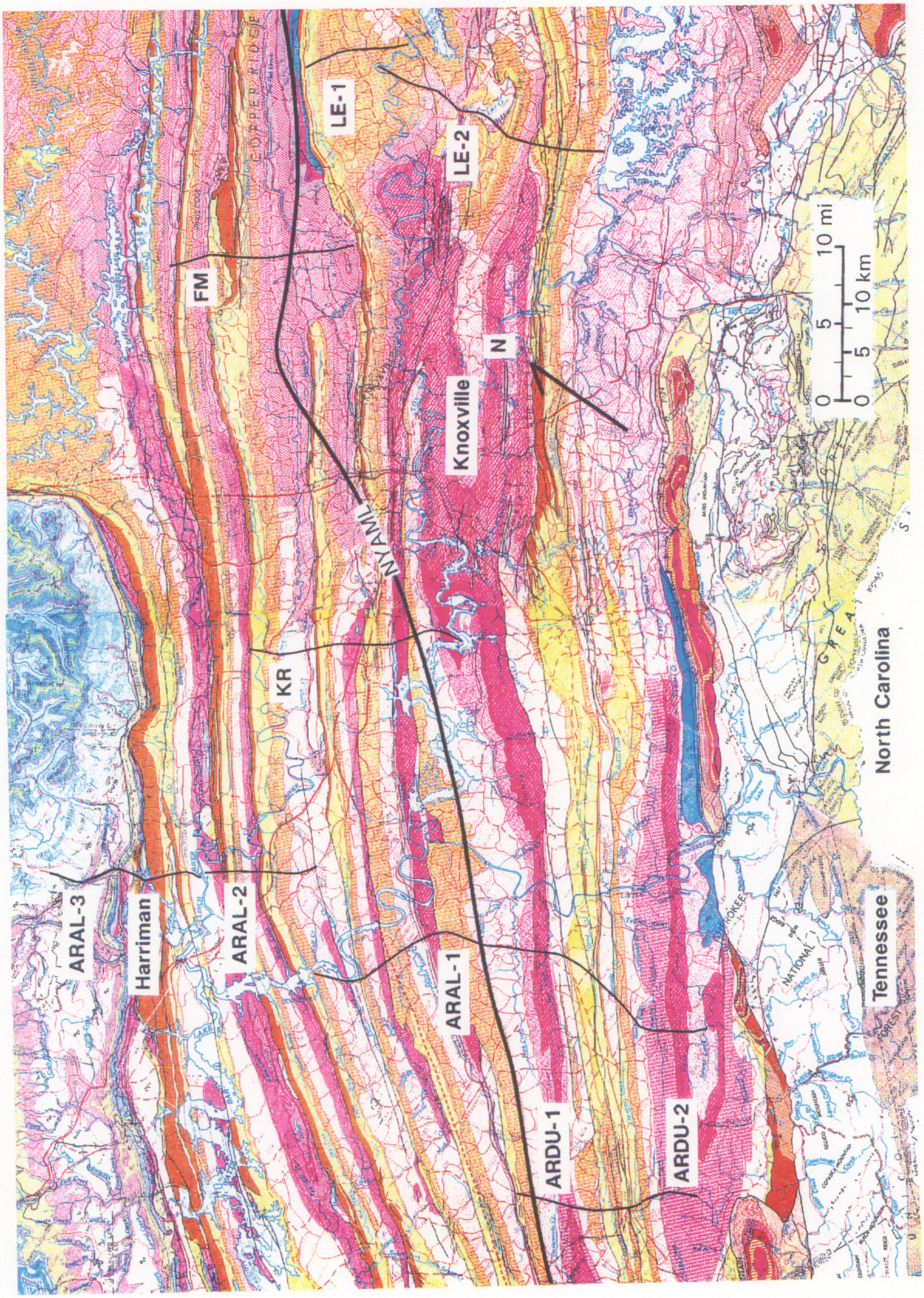

LE-1
LE-2
FM
N
Knoxville
NYAML
KR
ARAL-3
Harriman
ARAL-2
ARAL-1
ARDU-1
ARDU-2
0 5 10 mi
0 5 10 km
North Carolina
Tennessee

Figure 2.2a. ALD of reflection profile ARAL (uninterpreted): As the longest in the data set, profile ARAL is the best suited to image large bodies in the crust. Reflections from the shelf strata illuminate imbricate thrust sheets in the Valley and Ridge Province and horizontal layering in the Cumberland Plateau above 1.4 seconds (4.5 km). The Alleghanian structural front is visible at cmp 700 on line ARAL-3 near the northwestern end of the profile. A distinct wedge-shaped block is visible below the shelf strata extending across the northwestern third of the profile. Southeast of the wedge, west-dipping reflections dominate the data. The events appear to merge into a middle crustal band of high reflectivity extending across the profile at about 9 seconds. Beneath the band, east-dipping reflections on the northwestern end of the profile can be seen to about 11 seconds. The gravity (G) and magnetic (M) profiles above the reflection data are taken from the Bouguer anomaly map of Tennessee (Johnson and Stearns, 1967) and the Residual total intensity map of Tennessee (Johnson et al., 1979), respectively. The New York - Alabama Magnetic Lineament is formed by the steeply decreasing gradient near the center of the profile. The position of the NYAML shown on the index map in Figure 1.1 is centered on the gradient. ALD = Automatic Line Drawing.

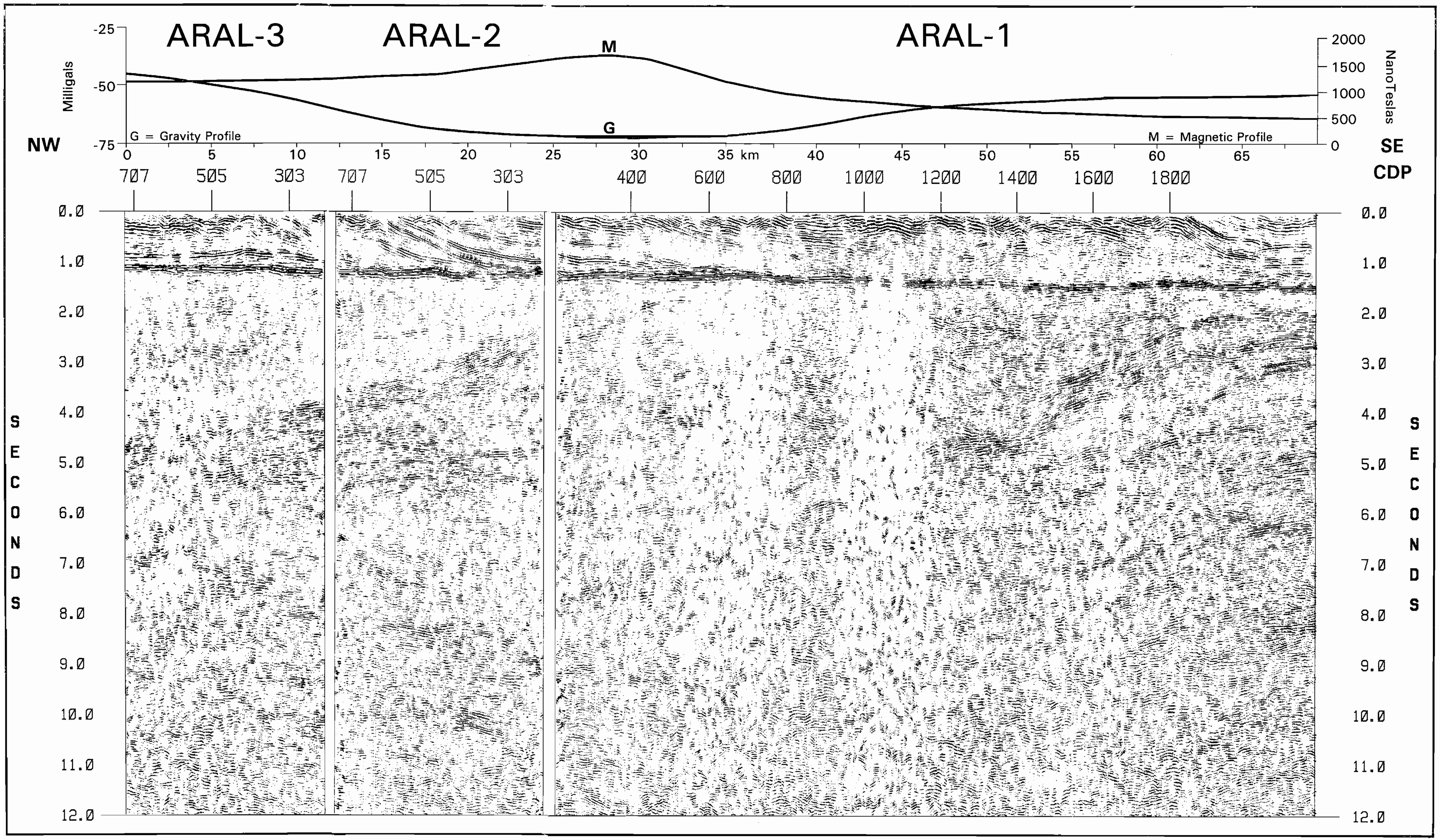

ARAL-3
ARAL-2
ARAL-1
M
G
Milligals
NanoTeslas
G = Gravity Profile
M = Magnetic Profile
NW
SE
CDP
km
SECONDS

Figure 2.2b. ALD of reflection profile ARAL (interpreted): Reflections from the shelf strata illuminate imbricate thrust sheets in the Valley and Ridge Province and horizontal layering in the Cumberland Plateau in the shallow part of the section above 1.4 seconds (4.5 km). The Alleghanian structural front is visible at cmp 700 on line ARAL-3 near the northwestern end of the profile. A distinct wedge-shaped block is visible below the shelf strata extending across the northwestern third of the profile. Southeast of the wedge, west-dipping reflections dominate the data. The events appear to merge into a middle crustal band of high reflectivity extending across the profile at about 9 seconds. Beneath the band, east-dipping reflections on the northwestern end of the line can be seen to about 11 seconds. The gravity (G) and magnetic (M) profiles above the reflection data are taken from the Bouguer anomaly map of Tennessee (Johnson and Stearns, 1967) and the Residual total intensity map of Tennessee (Johnson et al., 1979), respectively. The New York - Alabama Magnetic Lineament is formed by the steeply decreasing gradient near the center of the profile. The position of the NYAML shown on the index map in Figure 1.1 is centered on the gradient. ALD = Automatic Line Drawing.

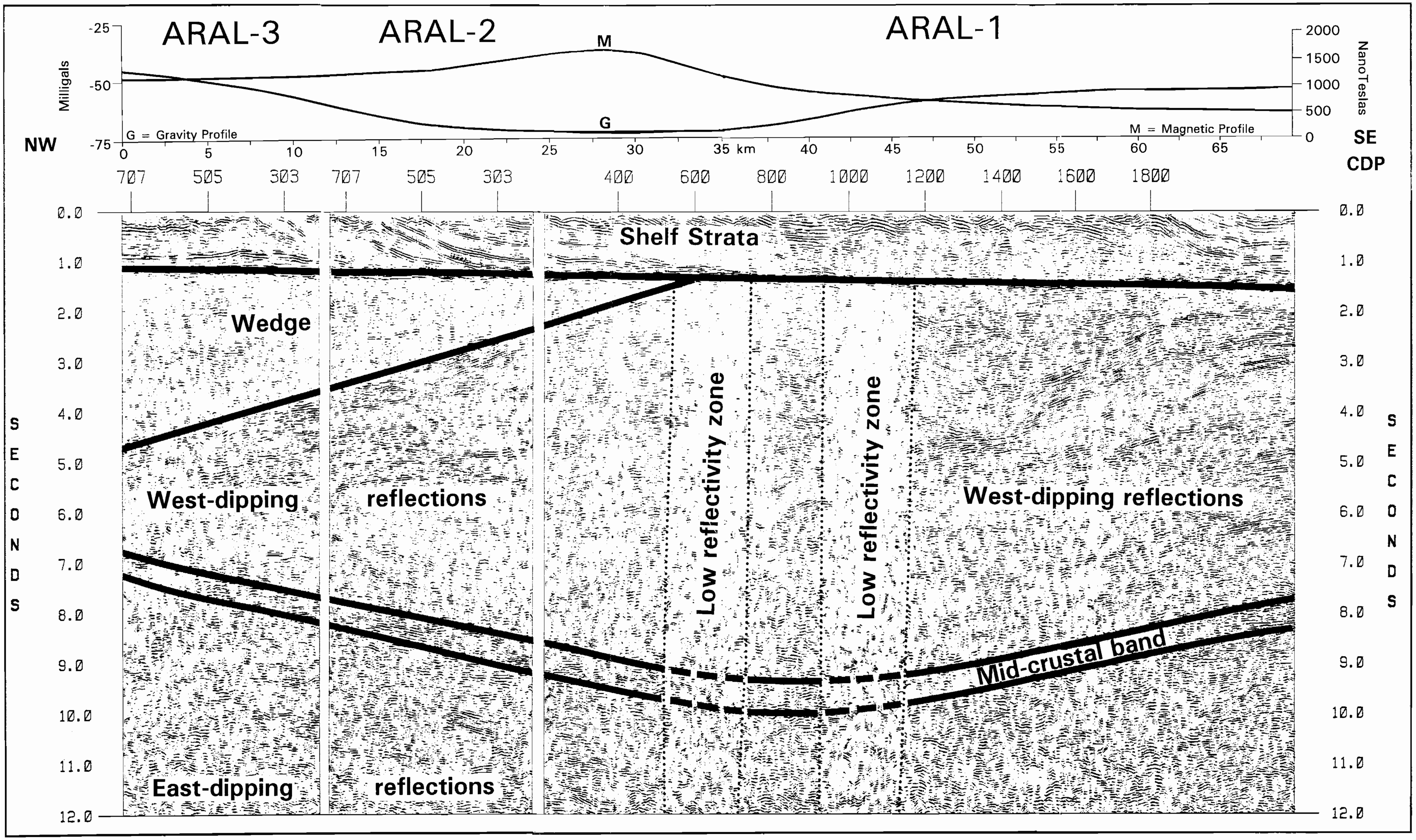

ARAL-3
ARAL-2
ARAL-1
M
G
G = Gravity Profile
M = Magnetic Profile
Milligals
NanoTeslas
NW
SE
CDP
km
Shelf Strata
Wedge
West-dipping
reflections
Low reflectivity zone
Low reflectivity zone
West-dipping reflections
Mid-crustal band
East-dipping
reflections
SECONDS
SECONDS

Figure 2.3a. ALD of the wedge (uninterpreted): The northwestern end of Profile ARAL images a distinct wedge-shaped body extending from the base of the shelf strata to 6 seconds. The individually processed lines ARAL-2 and ARAL-3 exhibit the low reflective region above the strongly reflective crust typified by west-dipping reflections. Subhorizontal reflections appear within the wedge and contrast with the west-dipping reflections southeast of the contact. Events along the contact are strongly reflective and suggest angular discordance. ALD = Automatic Line Drawing.

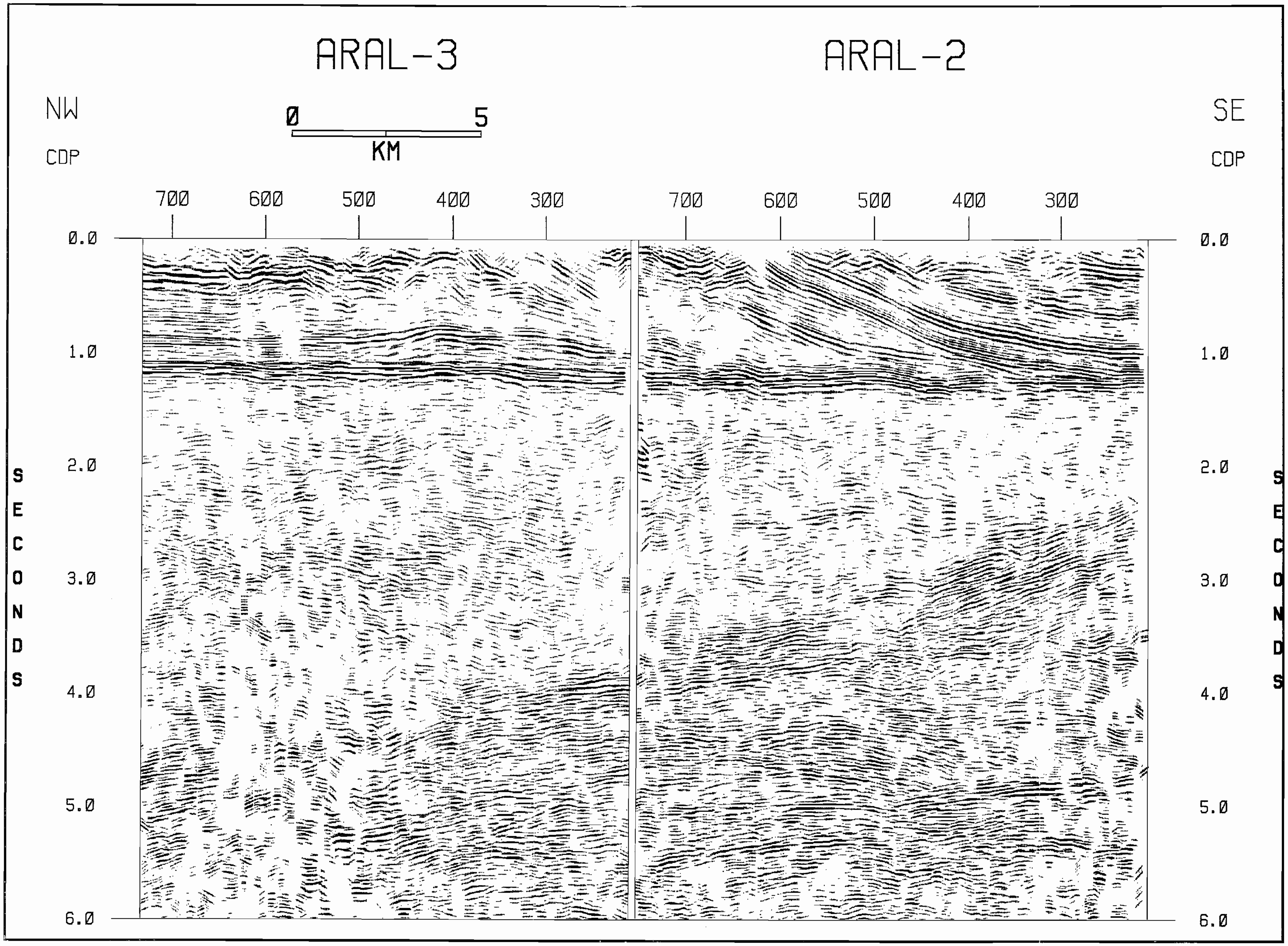

ARAL-3
ARAL-2
NW
SE
0 5
KM
CDP
CDP
700 600 500 400 300
700 600 500 400 300
0.0
1.0
2.0
3.0
4.0
5.0
6.0
SECONDS
SECONDS

Figure 2.3b. ALD of the wedge (interpreted): The wedge-shaped body stands out as a distinct crustal block in comparison with the adjacent crust and with the shallower shelf strata. Most of the reflections at the boundary between the wedge and the crust to the southeast do not parallel the contact, suggesting an angular relationship. Subhorizontal reflections within the wedge contrast in dip with those in the adjacent crust. The wedge shows no sign of thinning at the northwestern end of line ARAL-3, suggesting that it extends beneath the Cumberland Plateau. The Alleghanian structural front is visible at cmp 700, where folded rocks at the surface merge into flat-lying strata of the Cumberland Plateau. ALD = Automatic Line Drawing.

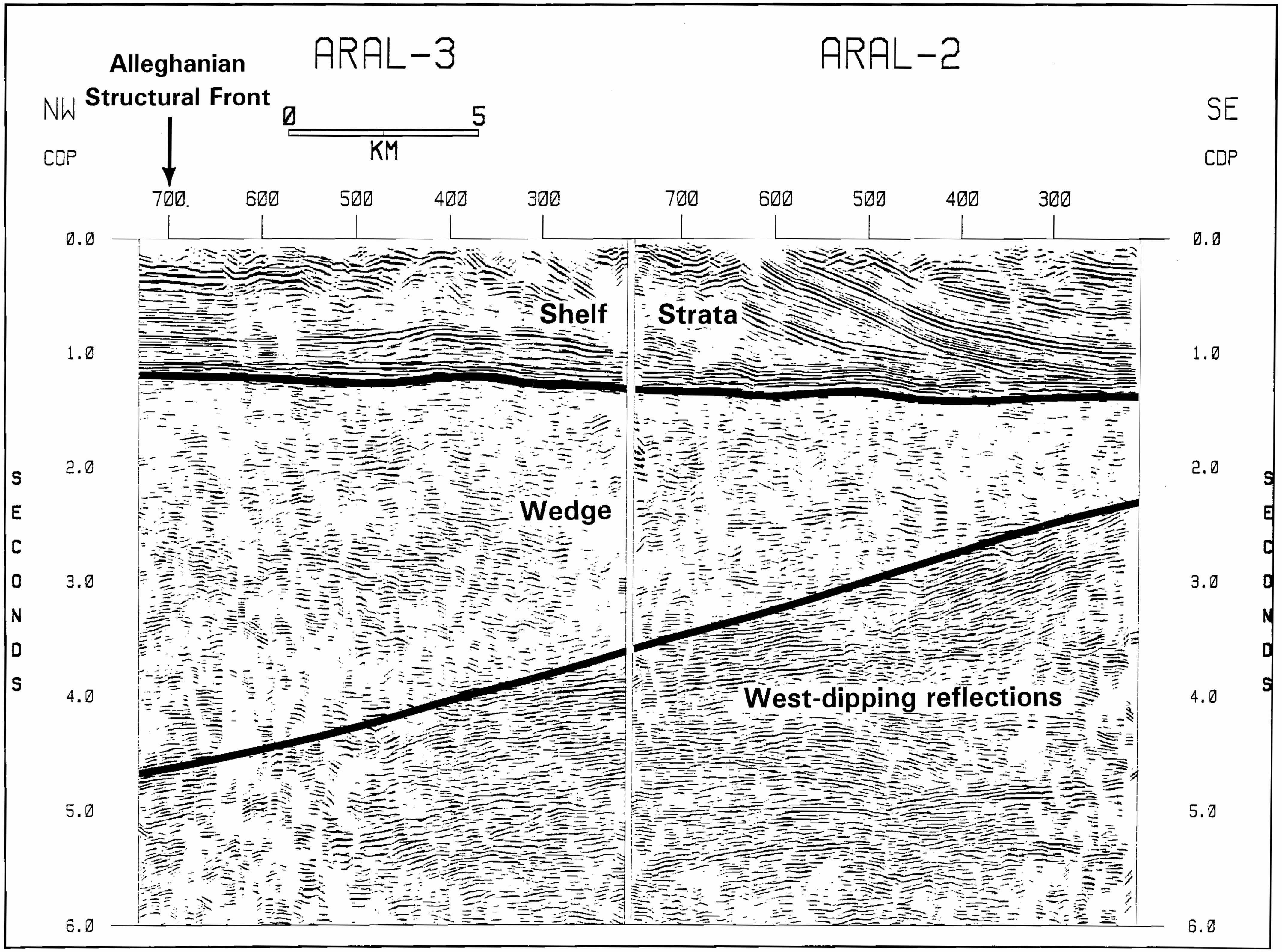

ARAL-3
ARAL-2
Alleghanian
Structural Front
NW
SE
0
5
KM
CDP
CDP
700
600
500
400
300
700
600
500
400
300
0.0
1.0
2.0
3.0
4.0
5.0
6.0
SECONDS
Shelf
Strata
Wedge
West-dipping reflections

Figure 2.4a. ALD of profile ARDU (uninterpreted): The short profile is composited from lines ARDU-1 and ARDU-2. Acquisition of these data with 20 second sweeps and 23 second record length permits extending the correlation of the data to 18 seconds. The low reflectivity region below 1.3 seconds might be related to the wedge imaged on profile ARAL, but the geometry of the body is indistinct. West dip is visible between cmps 150 and 400 from 2.5 to 3.8 seconds, but most of the reflections are subhorizontal. The bright band of reflections in the middle crust can be seen at 11 seconds. The Moho can be seen as a broad band of reflections between 14 and 15 seconds. The gravity (G) and magnetic (M) profiles above the reflection data are taken from the Bouguer anomaly map of Tennessee (Johnson and Stearns, 1967) and the Residual total intensity map of Tennessee (Johnson et al., 1979), respectively. The New York - Alabama Magnetic Lineament is formed by the steeply decreasing gradient near the northwest end of the profile. The position of the NYAML shown on the index map in Figure 1.1 is centered on the gradient. ALD = Automatic Line Drawing.

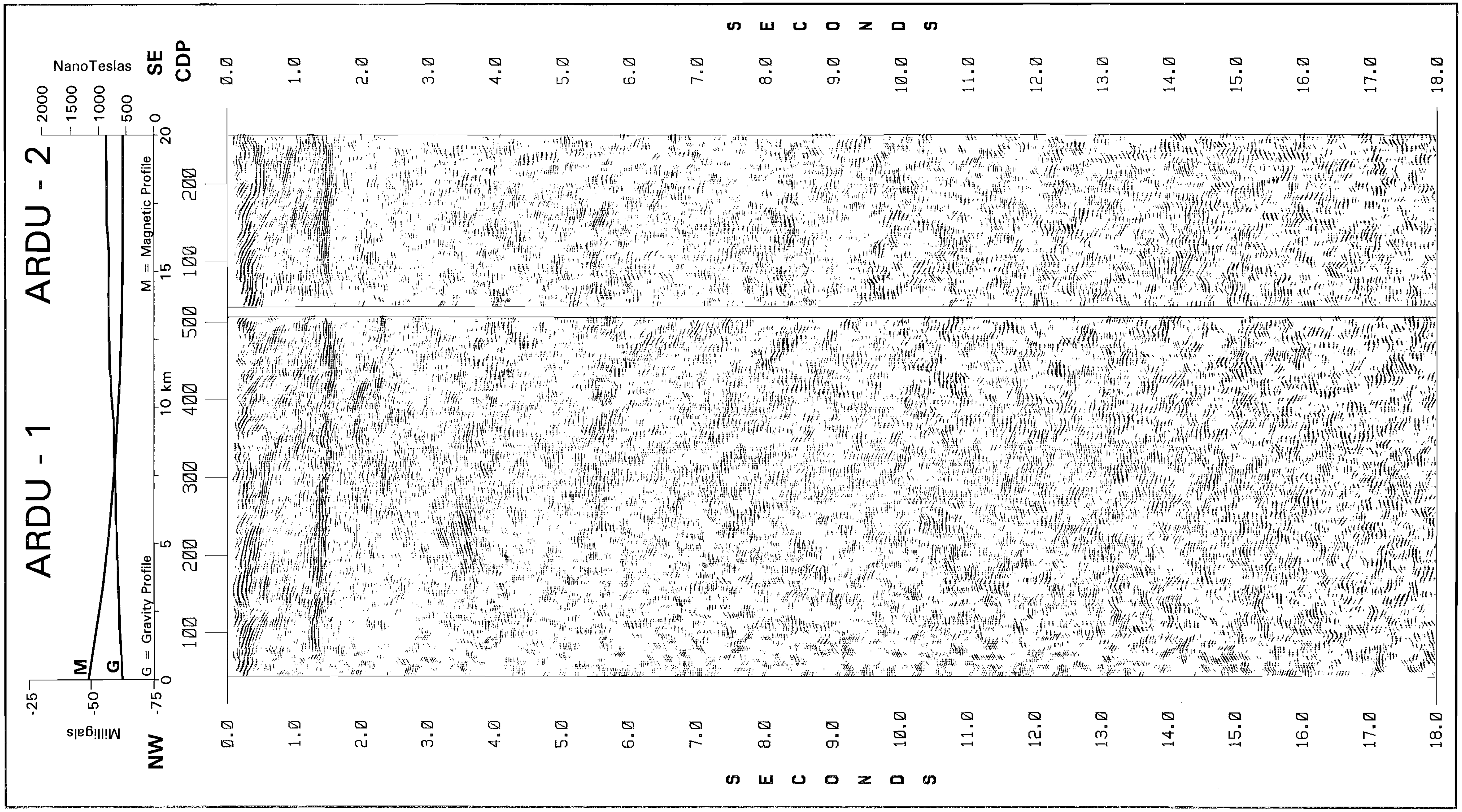
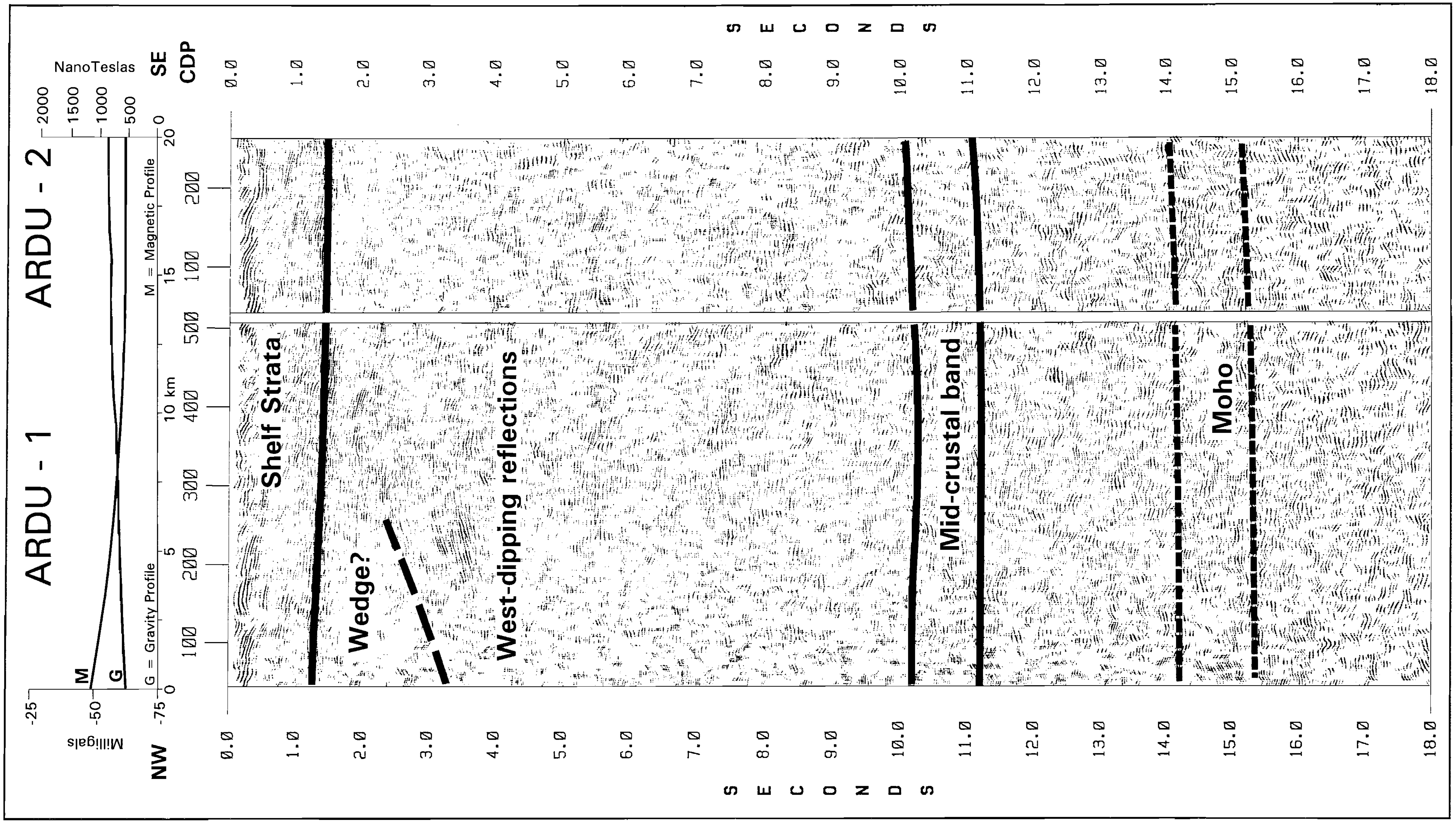

ARDU - 1
ARDU - 2
NanoTeslas
Milligals
SE
NW
CDP
SECONDS
M = Magnetic Profile
G = Gravity Profile
10 km

Figure 2.4b. ALD of profile ARDU (interpreted): The low reflectivity region below 1.3 seconds might be related to the wedge imaged on profile ARAL, but the geometry of the body is indistinct. West dip is visible between cmps 150 and 400 from 2.5 to 3.8 seconds, but most of the reflections are subhorizontal. The bright band of reflections in the middle crust can be seen at 11 seconds. The Moho can be seen as a broad band of reflections between 14 and 15 seconds. The gravity (G) and magnetic (M) profiles above the reflection data are taken from the Bouguer anomaly map of Tennessee (Johnson and Stearns, 1967) and the Residual total intensity map of Tennessee (Johnson et al., 1979), respectively. The New York - Alabama Magnetic Lineament is formed by the steeply decreasing gradient near the northwest end of the profile. The position of the NYAML shown on the index map in Figure 1.1 is centered on the gradient. ALD = Automatic Line Drawing.



ARDU - 1
ARDU - 2
NanoTeslas
Milligals
SE
NW
CDP
M = Magnetic Profile
G = Gravity Profile
10 km
SECONDS
Shelf Strata
Wedge?
West-dipping reflections
Mid-crustal band
Moho

Figure 2.5a. ALD of profile KR (uninterpreted): The single line profile reveals a distinct region of low reflectivity below 1.5 seconds that is interpreted as the northeastern extension of the wedge. Strongly reflective crust might be related to the region of west-dipping reflectivity on profile ARAL; however, most of the reflections within the more highly reflective region are subhorizontal. The middle crustal band of higher reflectivity is visible at 9 seconds. The gravity (G) and magnetic (M) profiles above the reflection data are taken from the Bouguer anomaly map of Tennessee (Johnson and Stearns, 1967) and the Residual total intensity map of Tennessee (Johnson et al., 1979), respectively. The New York - Alabama Magnetic Lineament is formed by the steeply decreasing gradient on the profile. The position of the NYAML shown on the index map in Figure 1.1 is centered on the gradient. ALD = Automatic Line Drawing.

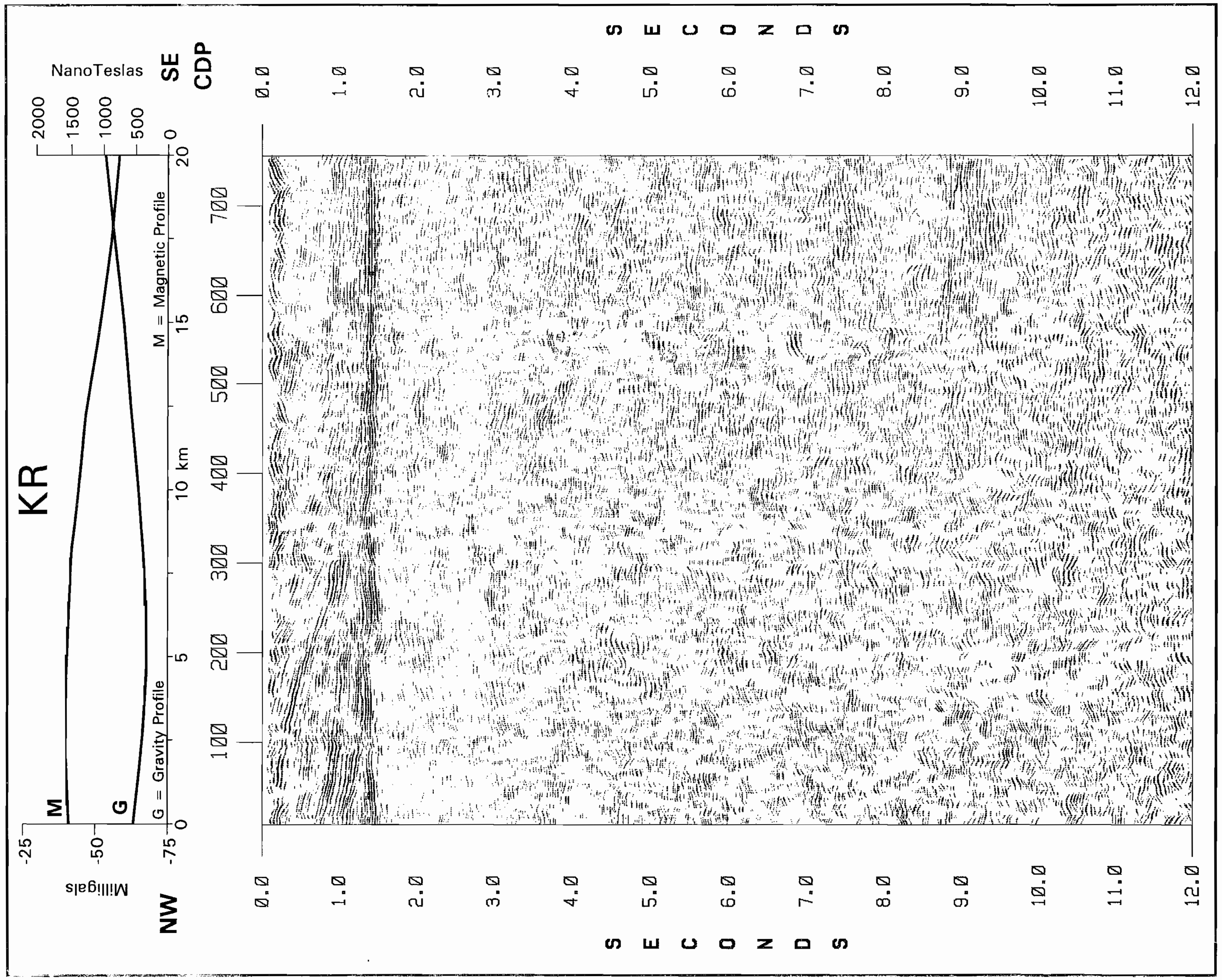

KR
NW
SE
CDP
NanoTeslas
Milligals
M = Magnetic Profile
G = Gravity Profile
km
SECONDS

Figure 2.5b. ALD of profile KR (interpreted): The distinct region of low reflectivity below 1.5 seconds can be interpreted as the northeastern extension of the wedge. Strongly reflective crust can be seen beneath the low reflective zone that might be related to the region of west-dipping reflectivity on profile ARAL; however, most of the reflections within the more highly reflective region are subhorizontal. The middle crustal band of higher reflectivity is visible at 9 seconds. The gravity (G) and magnetic (M) profiles above the reflection data are taken from the Bouguer anomaly map of Tennessee (Johnson and Stearns, 1967) and the Residual total intensity map of Tennessee (Johnson et al., 1979), respectively. The New York - Alabama Magnetic Lineament is formed by the steeply decreasing gradient on the profile. The position of the NYAML shown on the index map in Figure 1.1 is centered on the gradient. ALD = Automatic Line Drawing.

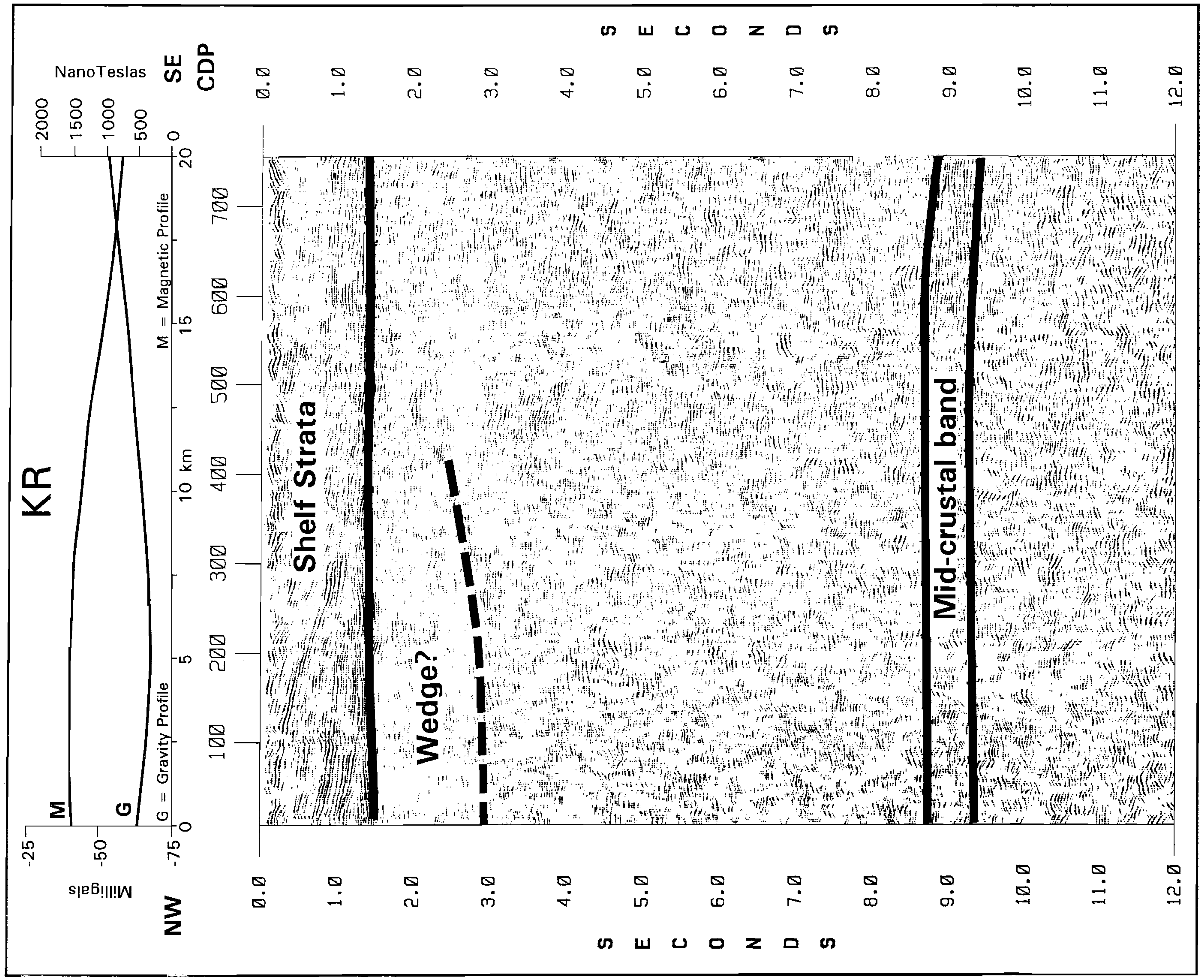

KR
NanoTeslas
Milligals
M = Magnetic Profile
G = Gravity Profile
km
SE
NW
CDP
Shelf Strata
Wedge?
Mid-crustal band
SECONDS

Figure 2.6a. ALD of profile FM (uninterpreted): The single line profile exhibits a region of lower reflectivity from 1.5 to 2.8 seconds on the northwestern half that could be evidence for the presence of the wedge. The reflections below and southeast of the region are subhorizontal but fairly bright and might correspond to the zone of west-dipping reflections on profile ARAL. The highly reflective band in the middle crust can be seen near 9 seconds dipping from the southeast to the northwest. Long sweep and record lengths permit the extension of correlation to 18 seconds. The Moho is imaged below 13 seconds on the southeastern end of the profile and dips about 8^{o} to 14 seconds on the northeastern end. The gravity (G) and magnetic (M) profiles above the reflection data are taken from the Bouguer anomaly map of Tennessee (Johnson and Stearns, 1967) and the Residual total intensity map of Tennessee (Johnson et al., 1979), respectively. The New York - Alabama Magnetic Lineament is formed by the steeply decreasing gradient across the profile. The position of the NYAML shown on the index map in Figure 1.1 is centered on the gradient. ALD = Automatic Line Drawing.

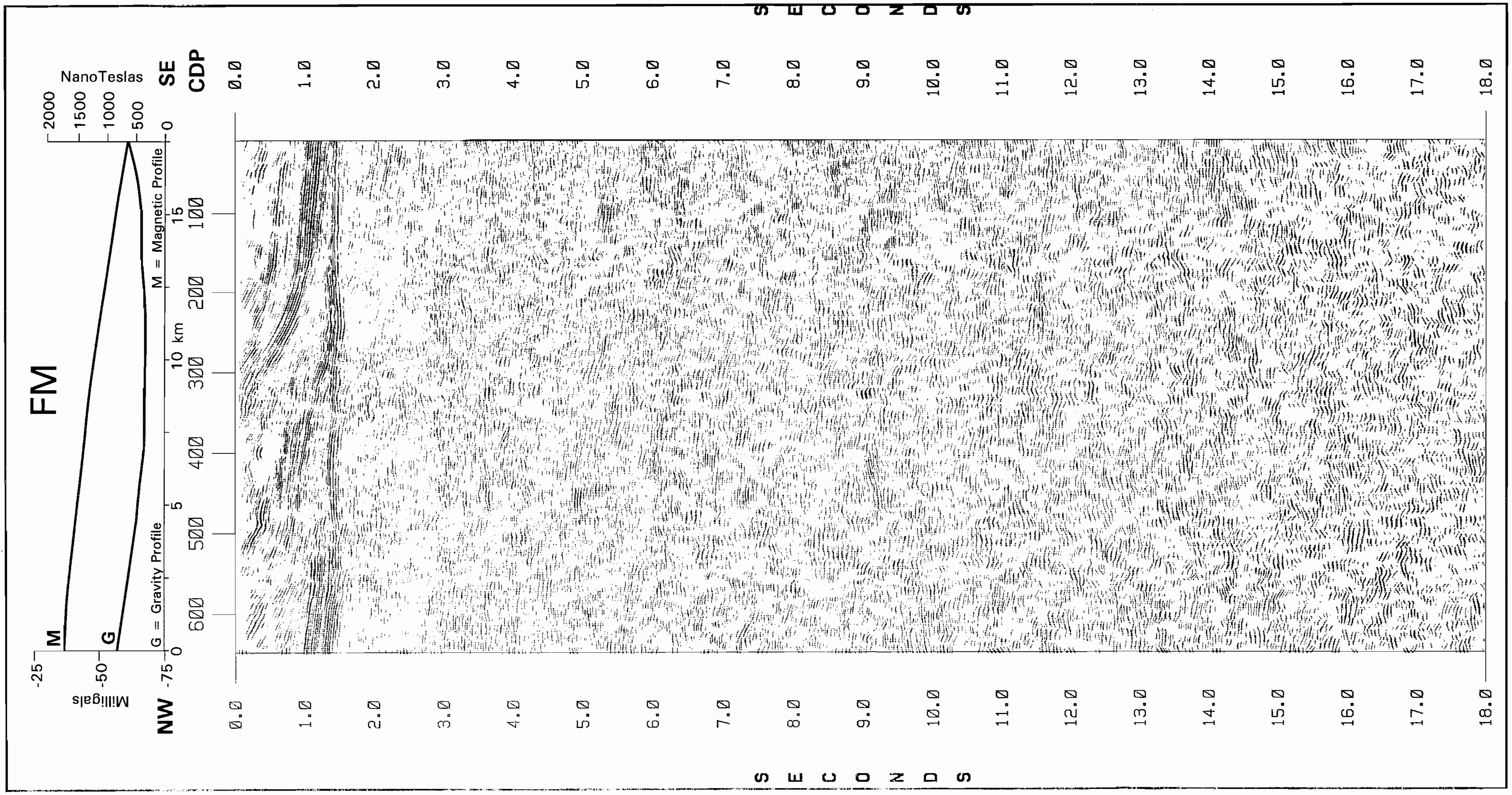

FM
NanoTeslas
2000
1500
1000
500
Milligals
-25
-50
-75
M
G
M = Magnetic Profile
G = Gravity Profile
0
5
10
15
km
SE
NW
CDP
100
200
300
400
500
600
SECONDS
0.0
1.0
2.0
3.0
4.0
5.0
6.0
7.0
8.0
9.0
10.0
11.0
12.0
13.0
14.0
15.0
16.0
17.0
18.0

Figure 2.6b. ALD of profile FM (interpreted): The region of lower reflectivity from 1.5 to 2.8 seconds on the northwestern half of the profile could be evidence for the presence of the wedge. The reflections below and southeast of the region are subhorizontal but fairly bright and might correspond to the zone of west-dipping reflections on profile ARAL. The highly reflective band in the middle crust can be seen near 9 seconds dipping from the southeast to the northwest. Long sweep and record lengths permit the extension of correlation to 18 seconds. The Moho is imaged below 13 seconds on the southeastern end of the profile and dips about 8^{o} to 14 seconds on the northeastern end. The gravity (G) and magnetic (M) profiles above the reflection data are taken from the Bouguer anomaly map of Tennessee (Johnson and Stearns, 1967) and the Residual total intensity map of Tennessee (Johnson et al., 1979), respectively. The New York - Alabama Magnetic Lineament is formed by the steeply decreasing gradient across the profile. The position of the NYAML shown on the index map in Figure 1.1 is centered on the gradient. ALD = Automatic Line Drawing.

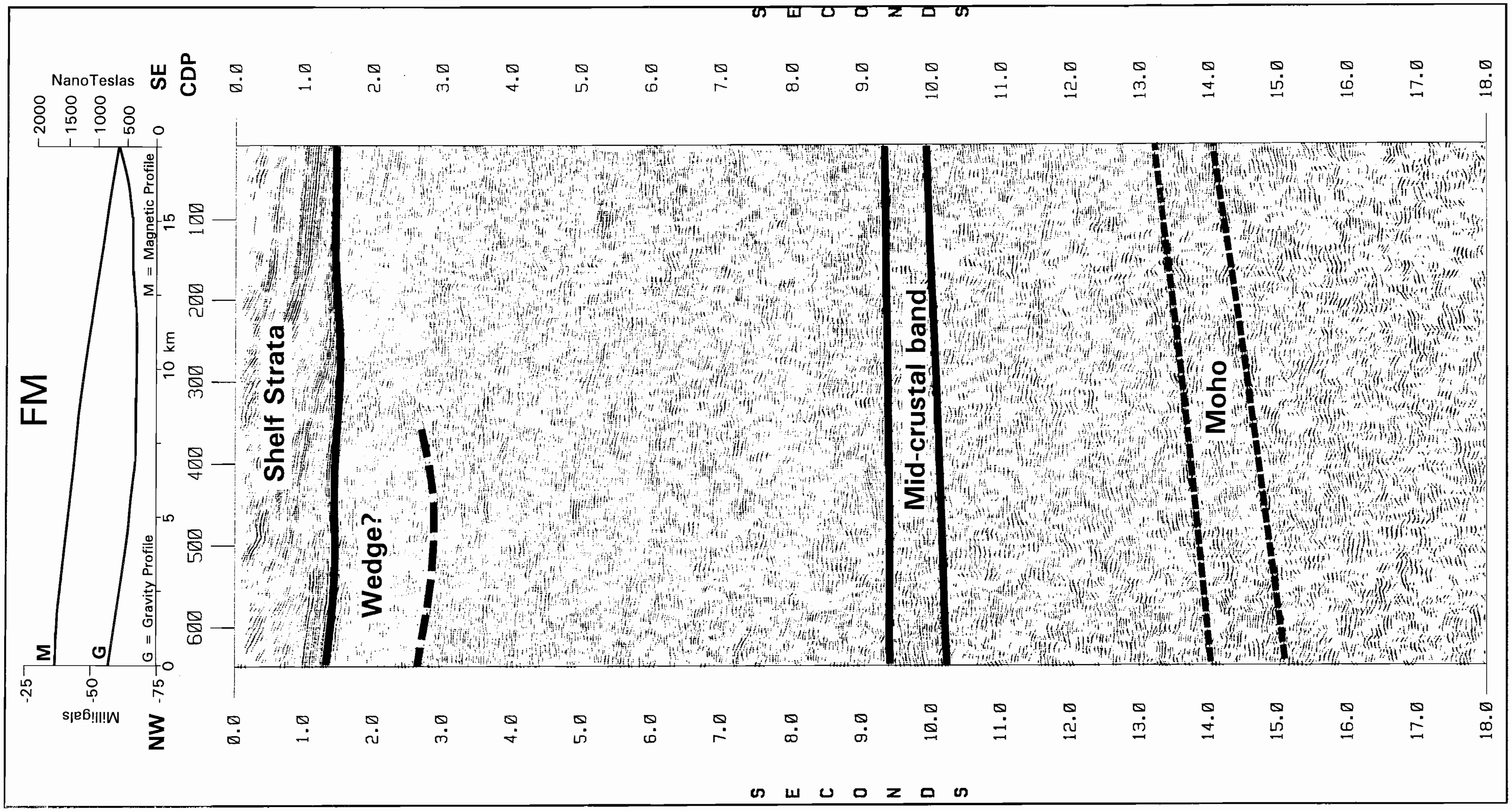

FM
NanoTeslas
2000
1500
1000
500
Milligals
-25
-50
-75
M
G
M = Magnetic Profile
G = Gravity Profile
0
5
10 km
15
NW
SE
CDP
100
200
300
400
500
600
Shelf Strata
Wedge?
Mid-crustal band
Moho
SECONDS
0.0
1.0
2.0
3.0
4.0
5.0
6.0
7.0
8.0
9.0
10.0
11.0
12.0
13.0
14.0
15.0
16.0
17.0
18.0

Figure 2.7a. ALD of profile LE-1 (uninterpreted): Overall loss of reflectivity on this profile results in few images of the deeper crust. Stronger reflectivity below 1.5 seconds suggests that the low reflective wedge does not extend to this line or lies deeper in the crust where it can not be distinguished. The loss of signal might indicate a lateral change in crustal composition or reflectivity. The gravity (G) and magnetic (M) profiles above the reflection data are taken from the Bouguer anomaly map of Tennessee (Johnson and Stearns, 1967) and the Residual total intensity map of Tennessee (Johnson et al., 1979), respectively. The New York - Alabama Magnetic Lineament is immediately northwest of the reflection line. The position of the NYAML shown on the index map in Figure 1.1 is centered on the steep gradient. ALD = Automatic Line Drawing.

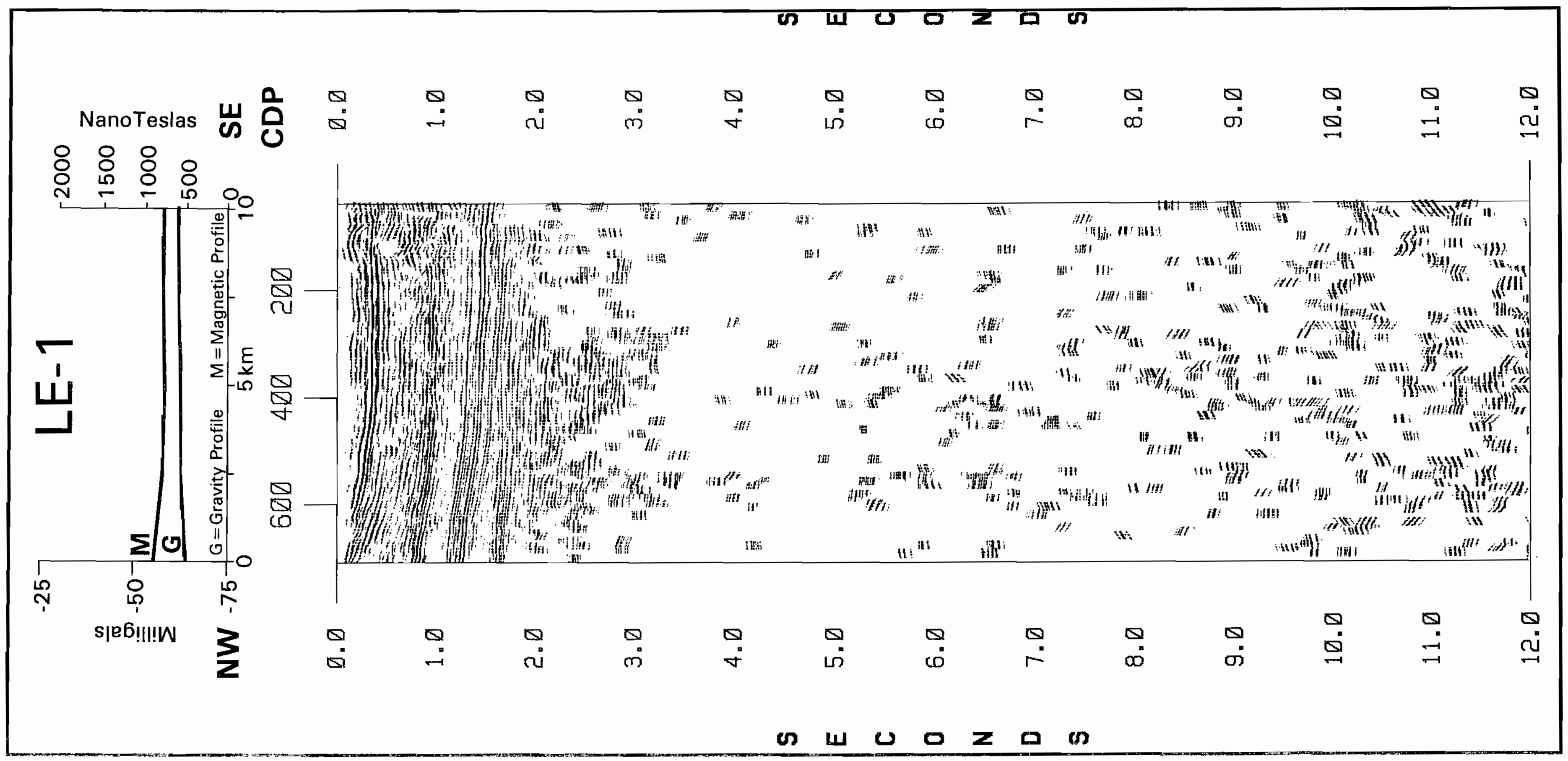
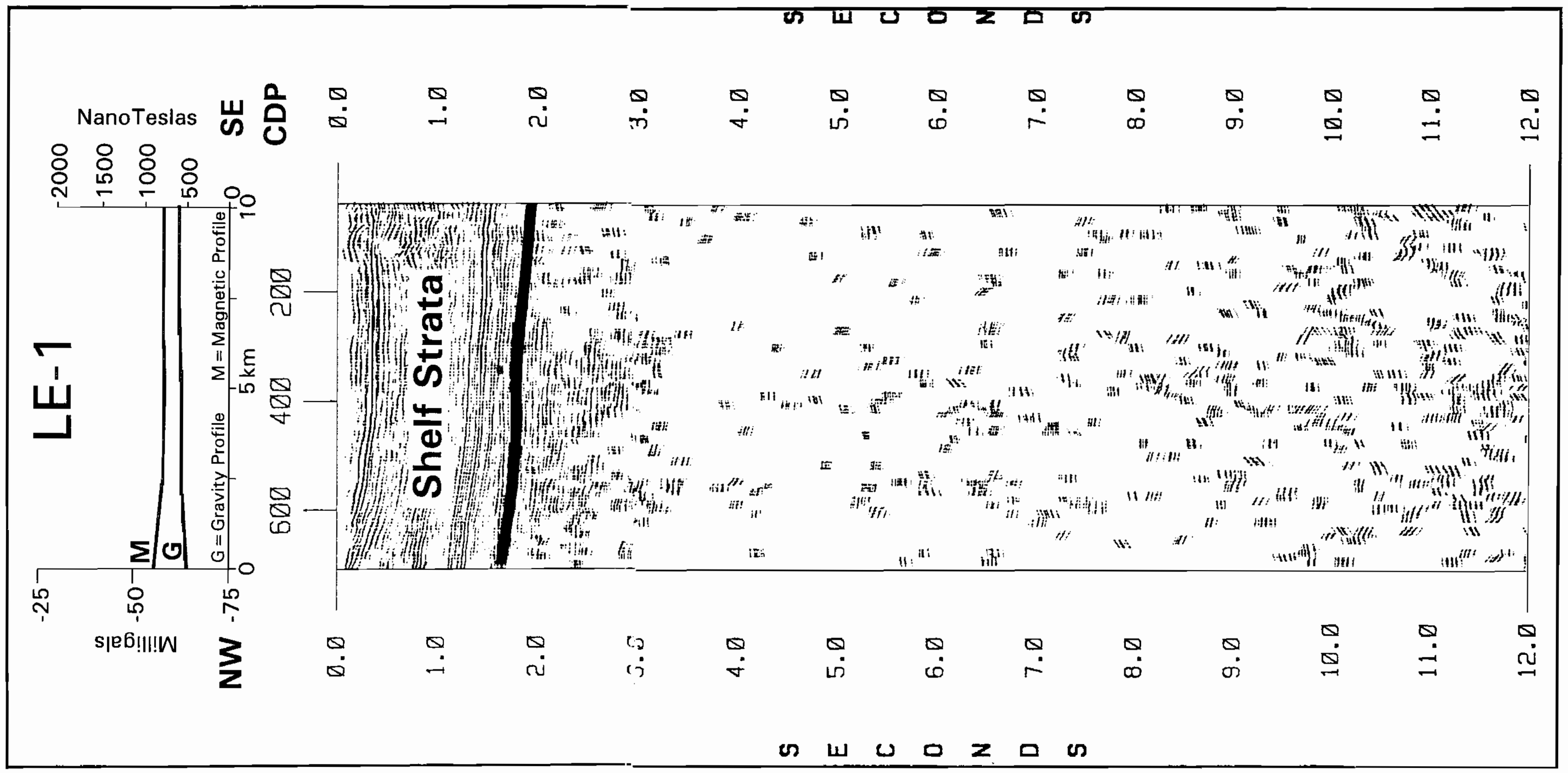

LE-1
NanoTeslas
Milligals
G=Gravity Profile
M=Magnetic Profile
5 km
NW
SE
CDP
SECONDS
SECONDS

Figure 2.7b. ALD of profile LE-1 (interpreted): Loss of reflectivity on the profile results in few images of the deeper crust. Stronger reflectivity below 1.5 seconds suggests that the low reflective wedge does not extend to this line or lies deeper in the crust where it can not be distinguished. The loss of signal might indicate a lateral change in crustal composition or reflectivity. The gravity (G) and magnetic (M) profiles above the reflection data are taken from the Bouguer anomaly map of Tennessee (Johnson and Stearns, 1967) and the Residual total intensity map of Tennessee (Johnson et al., 1979), respectively. The New York - Alabama Magnetic Lineament is immediately northwest of the reflection line. The position of the NYAML shown on the index map in Figure 1.1 is centered on the steep gradient. ALD = Automatic Line Drawing.



LE-1
NanoTeslas
2000
1500
1000
500
SE
CDP
M
G
G=Gravity Profile
M=Magnetic Profile
5 km
Milligals
-25
-50
-75
NW
Shelf Strata
SECONDS

Figure 2.8a. ALD of profile LE-2 (uninterpreted): Loss of reflectivity has resulted in poor returns from the deeper crust. The low reflectivity of this line might indicate a change in composition of the crust in the northeastern part of the study area. Alternatively, as the strike of profile LE-2 is essentially north-south, the lower reflectivity could indicate a change in strike of crustal structures to orientations less likely to be imaged on these data. The gravity (G) and magnetic (M) profiles above the reflection data are taken from the Bouguer anomaly map of Tennessee (Johnson and Stearns, 1967) and the Residual total intensity map of Tennessee (Johnson et al., 1979), respectively. The New York - Alabama Magnetic Lineament is northwest of the reflection line. The position of the NYAML shown on the index map in Figure 1.1 is centered on the steep gradient. ALD = Automatic Line Drawing.

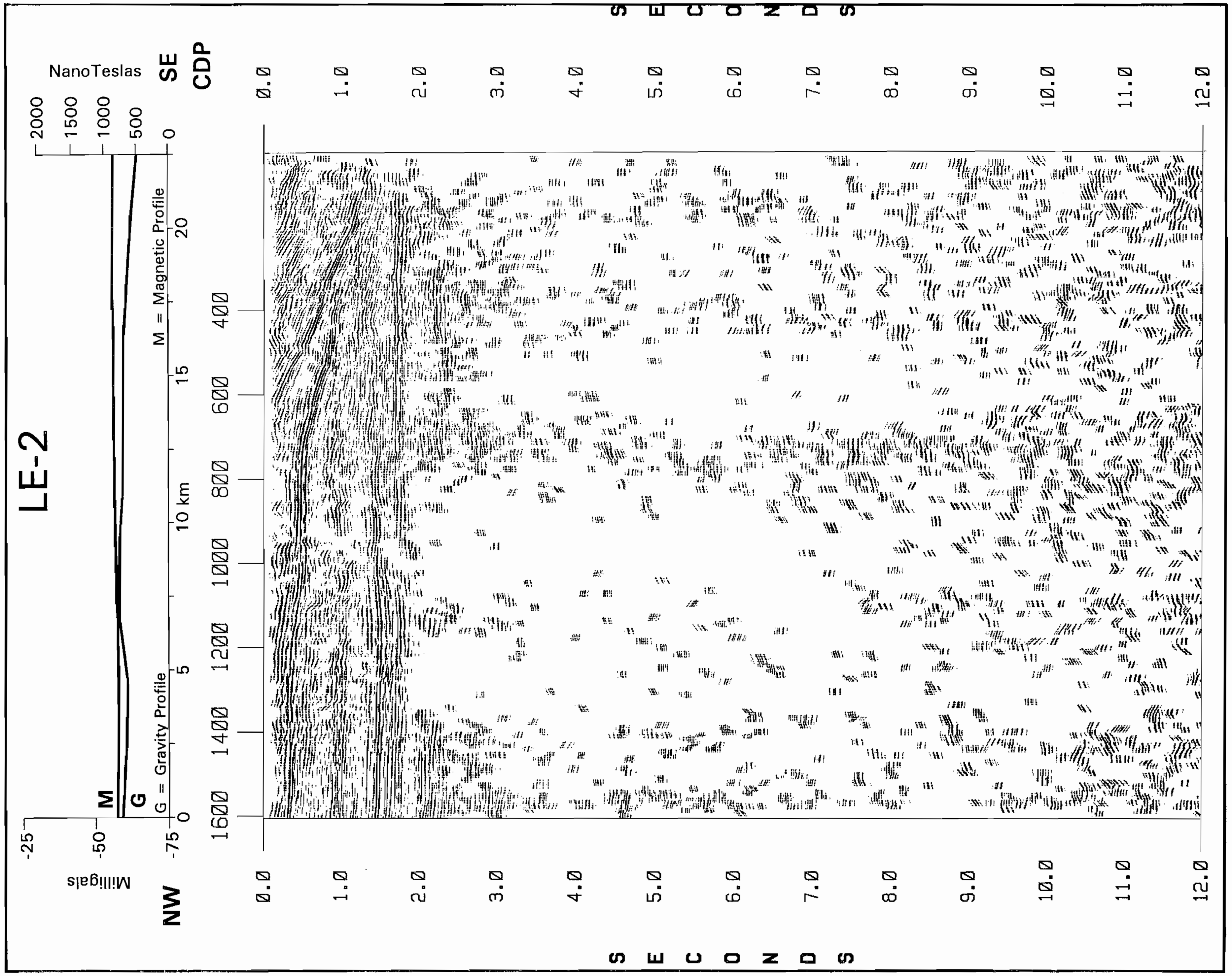
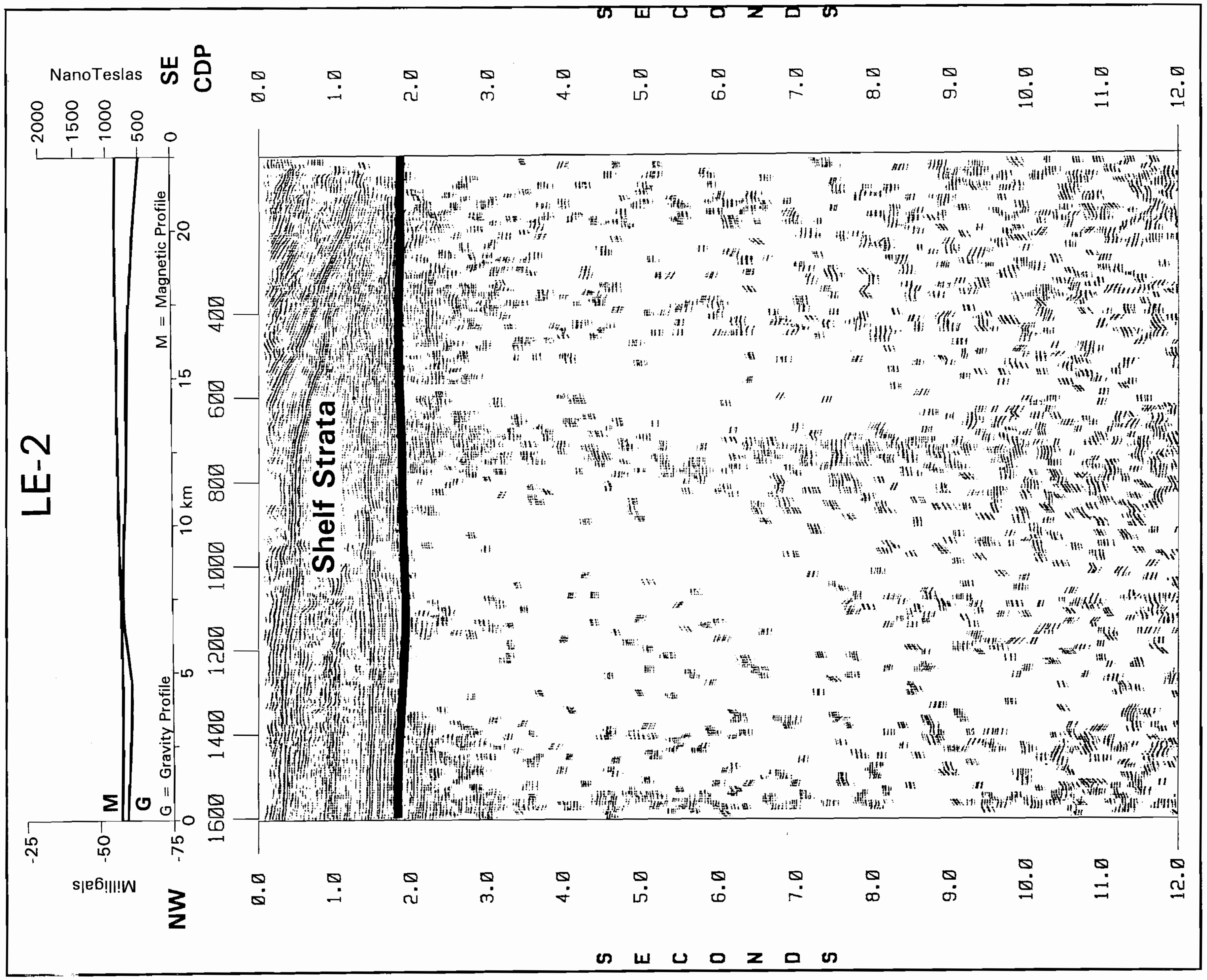

LE-2
NanoTeslas
Milligals
SE
NW
CDP
SECONDS
M = Magnetic Profile
G = Gravity Profile
10 km

Figure 2.8b. ALD of profile LE-2 (interpreted): The low reflectivity of this line might indicate a change in composition of the crust in the northeastern part of the study area. Alternatively, as the strike of profile LE-2 is essentially north-south, the lower reflectivity could indicate a change in strike of crustal structures to orientations less likely to be imaged on these data. The gravity (G) and magnetic (M) profiles above the reflection data are taken from the Bouguer anomaly map of Tennessee (Johnson and Stearns, 1967) and the Residual total intensity map of Tennessee (Johnson et al., 1979), respectively. The New York - Alabama Magnetic Lineament is northwest of the reflection line. The position of the NYAML shown on the index map in Figure 1.1 is centered on the steep gradient. ALD = Automatic Line Drawing.


LE-2
Shelf Strata
NanoTeslas
Milligals
NW
SE
CDP
M = Magnetic Profile
G = Gravity Profile
10 km
SECONDS

meters, an average crustal velocity of 6500 m/sec two-way travel-time was used for all depths. The commonly used p-wave velocity of 6000 m/sec is too low when compared to velocities determined by Prodehl et al. (1984) using the 1965 U.S.G.S. refraction data across the study area. Prodehl et al. suggested average upper crustal velocities of 6200 m/sec to 7-10 km depth, 6700 - 6800 m/sec between 17 and 34 km, and 7100 - 7400 m/sec from 40-47 km.

Description of imaged features

The subhorizontal reflection at the base of the shelf strata is visible at about 1.5 second (4.9 km) on all profiles. The reflection extends from beneath the subhorizontal rocks of the Cumberland Plateau on the northwest end of profile ARAL (Figure 2.2) across the profile beneath the dipping reflections from the Alleghanian thrust sheets. The Alleghanian structural front is imaged at CMP 700 on line ARAL-3. Rocks involved in Alleghanian deformation in the study area are Cambro-Ordovician in age and include primarily well-indurated carbonates and shales with occasional sandstone units.

The rocks of interest to this study are those of the crust and upper mantle beneath the shelf strata. These rocks comprise the crystalline basement between 4.9 to 39 km (1.5 to 12 seconds). The reflection profiles reveal that unexpected and areally extensive features are present.

The wedge

A pronounced wedge-shaped region of relatively low reflectivity is imaged on the western end of profile ARAL (Figures 2.2a and b, and Figure 2.3a and b). The wedge persists from the northwestern end of ARAL-3 to the southeastern end of ARAL-2, and might be present on the northwestern end of line ARAL-1. The wedge thickens to the northwest beneath the Cumberland Plateau where it reaches a thickness of over 3.4 seconds (11 km). The wedge thins to 1 second (3.25 km) at the southeastern end of ARAL-2 over a horizontal distance of 26 km. This geometry yields a dip of about 15^{o} for the contact between the wedge and the underlying basement. Visible within the wedge are subhorizontal reflections that exhibit increasing east dips as the reflections approach the contact between the wedge and the underlying crust. These reflections do not appear to extend across the contact, and their geometry suggests that they might roll into the contact.

The projection of the wedge to the southwest onto profile ARDU (Figure 2.4) is speculative. A high amplitude west-dipping package between cmps 150 and 300 at 3.2 to 3.8 seconds on profile ARDU bears a resemblance to high amplitude events along the contact between the wedge and the west-dipping reflections on profile ARAL near cmp 1550 at 3.0 to 4.0 seconds, and might form the basis for interpreting the wedge on profile ARDU. These reflections might be related to the contact between the wedge and the adjacent crust, although the distinctive geometry present on profile ARAL is not as well imaged.

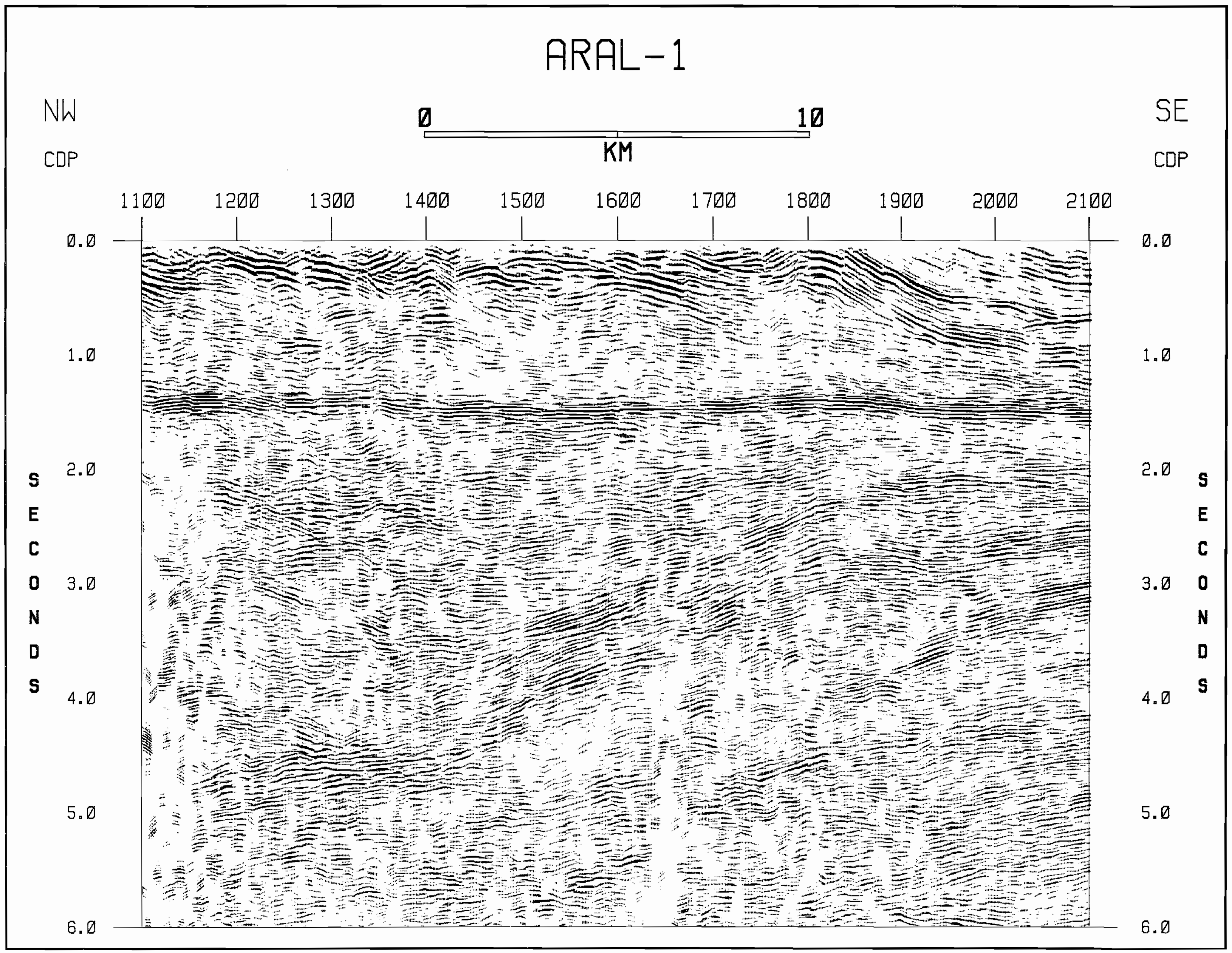
The presence of the wedge on profile KR (Figure 2.5a and b) can be inferred from the weaker reflectivity extending from the northwestern end of the line to about cmp 450. The region appears to taper from about 3 seconds (10 km) beneath cmp 7 to essentially zero at the center of the profile. Although west-dipping reflections appear within the less reflective region, those beneath the region consist of higher amplitudes. On profile FM (Figure 2.6a and b), fainter reflectivity is visible extending across the profile from about 1.5 seconds (4.9 km) to 2.7 seconds (9 km). The region is underlain by subhorizontal reflections of higher amplitude than that of the subhorizontal reflections within the lower reflective area. The lower reflectivity between the detachment surface and the more highly reflective part of the crust on profiles LE-1 and LE-2 (Figures 2.7a, b; and 2.8a, b; respectively) could be indicative of the presence of the wedge; however, no clear indication of the internal reflection character or the contact between the wedge and surrounding basement is present on these profiles.

West-dipping reflections

Pronounced west-dipping reflections dominate the southeastern half of profile ARAL (Figure 2.2a and b, and Figure 2.9a and b) and are visible on profile ARDU. Profiles KR, FM, LE1, and LE2 exhibit mainly subhorizontal reflections. The west-dipping reflections on profile ARAL are evident between 1.5 (4.9 km) and 9.0 seconds (29 km) where they dip approximately 30^o to the northwest from beneath the detachment surface of the Alleghanian allochthon to middle crustal depths. The reflections appear to dip less steeply near 9 seconds and merge into a band of higher reflectivity in the middle crust. West-dipping events are less pervasive on profiles KR and FM (Figures 2.5 and 2.6, respectively) to the northeast of profile ARAL (Figure 2.2). Profiles KR and FM are dominated by subhorizontal reflection packets with occasional west- and east-dipping events from approximately 2.5 seconds to around 9 seconds. The strong reflection strength within the crust beneath the detachment surface down to middle crustal depth forms the basis for extrapolating the crustal block typified by west-dipping reflections to profiles KR and FM. The change in orientation from west dip to subhorizontal could be related to a change in strike of the features responsible for the west dip or represent post-layering rotation.

Profiles LE-1 and LE-2 (Figures 2.7 and 2.8, respectively) contain suggestions of west dip within the more highly reflective regions of the crust, although most of the reflection segments appear subhorizontal. Reflection continuity is not as pervasive as the continuity on profile ARAL. West dips in the upper to middle crust are more obvious on LE-2 than on profiles LE-1or FM. The north-south orientation of LE-2 might be more favorable for better imaging of the west-dipping reflections than that of the northwest-southeast trending profiles LE-1 and FM, and might indicate that the strike of the layering responsible for the reflections has changed from the vicinity of profiles ARAL and KR to this location.

Figure 2.9a. ALD of west-dipping reflections (uninterpreted): Bright continuous reflections with marked west dip pervade the upper and middle crust southeast of and beneath the wedge. Reflections range in dip from 20^o to 30^o on these unmigrated data, suggesting that the layer boundaries responsible for the reflections range in dip from 21^o to 35^o (assuming that the seismic profile trends perpendicular to structural strike of the reflective interfaces). ALD = Automatic Line Drawing.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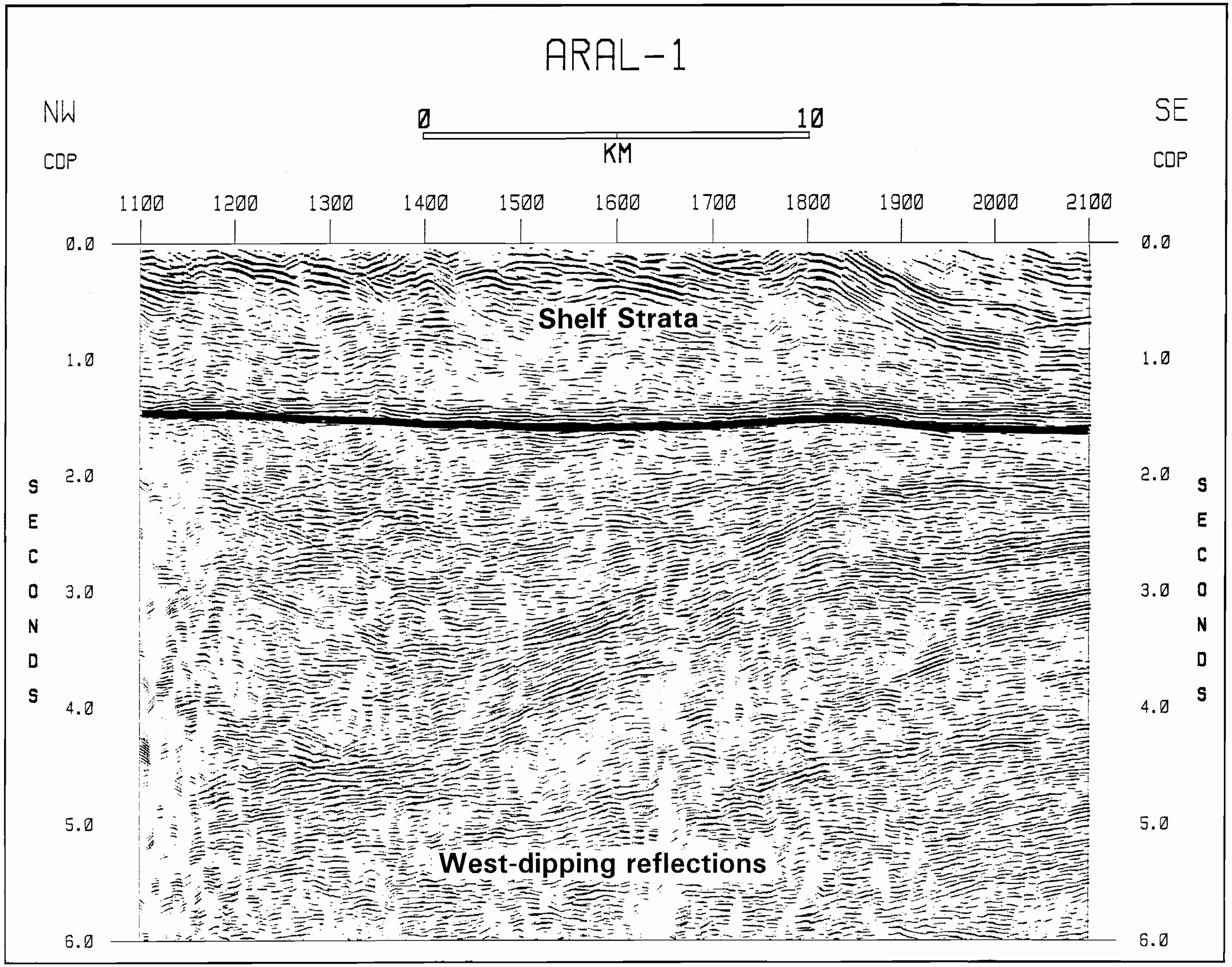
SECONDS

Figure 2.9b. ALD of west-dipping reflections (interpreted): Bright continuous reflections with marked west dip pervade the upper and middle crust southeast of and beneath the wedge. Reflections range in dip from 20^{o} to 30^{o} on these unmigrated data, suggesting that the layer boundaries responsible for the reflections range in dip from 21^{o} to 35^{o} (assuming that the seismic profile trends perpendicular to structural strike of the reflective interfaces). The dark line follows the contact between the shelf strata and the older crystalline basement. ALD = Automatic Line Drawing.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SECONDS
Shelf Strata
West-dipping reflections

Mid-crustal reflections

A band of highly reflective events is visible within the middle crust on profiles ARAL, ARDU, KR, and FM (Figures 2.2, 2.4, 2.5, and 2.6, respectively). On profile KR where they are best imaged, the events extend from about 8.7 to 9.5 seconds and are composed of stacked, subhorizontal, relatively long reflection segments. On profile ARAL, the band exhibits a broad, synformal geometry in which it dips from 7 seconds (23 km) on the northwestern end of the profile to 9 seconds (29 km) near the center of the profile at cmp 1000. From there, the band continues to the southeast as a fainter, west-dipping package to the end of the profile ARAL, where it can be seen at 7.8 seconds (25 km). On profile ARDU, a reflective band can be distinguished between 10 and 11 seconds, but it is less consistent than the band on profiles KR and ARAL. Profile FM exhibits a discontinuous band best identified on the northwestern end of the profile between 9.5 and 10.3 seconds. The band might continue across the profile more or less subhorizontally or be interpreted as dipping to the southeast where it reaches the end of the profile at approximately 11.5 seconds (37 km). Profiles LE-1 and LE-2 (Figures 2.7 and 2.8, respectively) do not exhibit a prominent middle crustal band.

East-dipping reflections

East-dipping reflections can be seen below the wedge mingled with short subhorizontal and more discontinuous west-dipping events on the northwestern end of profile ARAL (Figure 2.2). The most reflective of these events are imaged between 10.0 and 11.0 seconds (33 and 36 km) at cmp 350 on line ARAL-2 near the center of the profile. East-dipping reflections persist to 12 seconds across the profile to approximately cmp 1500. Southeast of cmp 1500, most of the deep crustal reflections are west-dipping. East-dipping events are less continuous on the shorter profiles. Profile ARDU contains east dipping reflections from 11.5 seconds (37 km) at cmp 100, line ARDU-1, to nearly 14 seconds (45.5 km) at cmp 500.

The Moho

On the basis of previous work (e.g, Prodehl et al., 1984; James et al., 1968) the Moho in this region is expected to be present at a depth of 45 km. Profiles ARDU and FM (Figures 2.4 and 2.6, respectively) were originally acquired with sufficient record lengths to permit extended correlation to 18 seconds while retaining a full octave (14 to 28 hertz) of frequency content. The Moho appears on these data as a diffuse band of higher reflectivity about 1 second thick.

On profile FM (Figure 2.6), the Moho appears as a brighter band of reflections dipping to the northwest in the vicinity of 14 seconds (45.5 km). The band of reflections is approximately 1 second wide. The top of the band can be seen at 13 seconds on the southeastern end of the profile and dips to 14 seconds on the northwestern end. The Moho plunges more than 3 km over a distance of 20 km for a dip of about 8^{o} on these unmigrated data, while the middle crustal band of reflectivity at about 9 seconds remains essentially horizontal.

The dip on the Moho apparent on profile FM is not observed on profile ARDU (Figure 2.4). The top of the band of Moho reflections can be seen extending across the data at about 14 seconds (45.5 km). The reflections form a subhorizontal band about 1 second thick, with the base of the reflection package at or below 15 seconds (49 km).

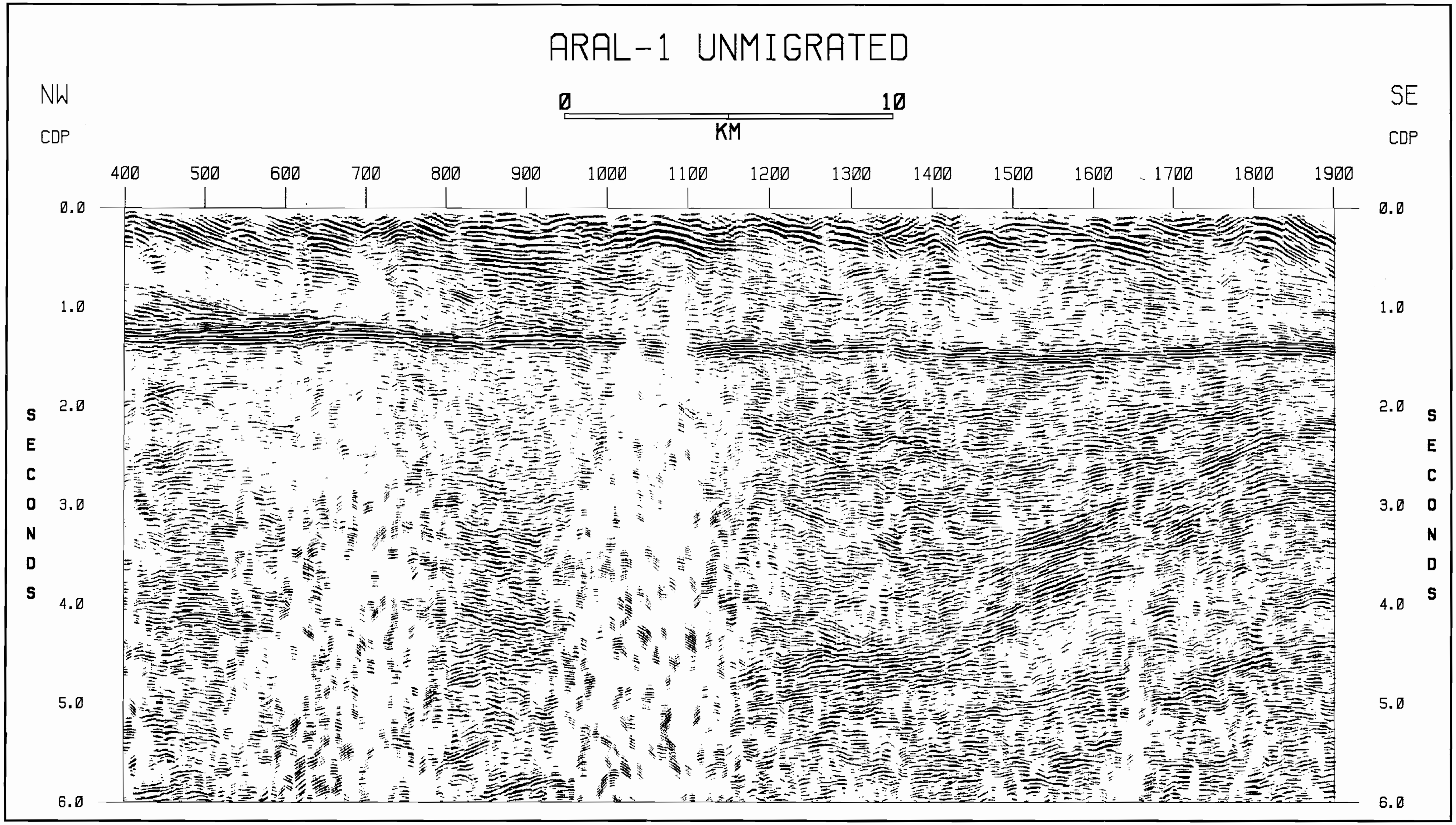
For the purposes of this study, the Moho will be interpreted to lie at the top of the bands of reflectivity. Reflections beneath the reflection band are slightly brighter and lower in frequency content, but do not differ significantly from those within the band attributed to the Moho. The transitional Moho model proposed by Prodehl et al. (1984) might be substantiated by these reflection seismic data. Furthermore, the data support crustal models of a thick crust beneath eastern Tennessee interpreted by previous authors (e.g. Prodehl et al., 1984; Long and Liou, 1986; James et al., 1968).

Low-reflectivity zones

Subvertical zones of low reflectivity are clearly visible on Profile ARAL (Figure 2.2). These features are not associated with known acquisition or processing problems and are considered to contain information about the crust. The relatively continuous west-dipping reflections do not maintain their continuity across the zones on the unmigrated profiles; however, the highly reflective base of the Alleghanian allochthon continues without interruption above some of the regions of low reflectivity. The low reflectivity zones extend to depths shallow enough to be investigated on migrated reflection data.

Although migration is not appropriate for most of the seismic lines in this study (see explanation, Appendix 2), line ARAL-1 is long enough to be migrated to 6 seconds near the center of the line. The uninterpreted and interpreted unmigrated automatic line drawings of line ARAL-1 are shown with the uninterpreted and interpreted migrated displays in Figures 2.10 and 2.11, respectively. Much of the energy concentrated in the zones on the migrated displays is from migration artifacts enhanced by the automatic line drawing processing; however, improvement in the shallower section is evident. West-dipping events northwest of the zones have migrated updip and to the southeast into the zones, while those to the southeast have migrated away. The zones have been narrowed by the migration of events to more accurate positions, and the interpreted position of the zone near cmp 1100 has been shifted to the southeast. The subhorizontal reflection from the base of the shelf strata has improved in continuity above the low reflective zone between cmps 1020 and 1100, indicating that the low reflective zone does not extend above that boundary. The data are interpreted to indicate the presence of discontinuous layers at the edges of subvertical dike swarms. The continuity of the layer at the base of the shelf strata suggests that the swarms were emplaced prior to erosion of the basement to that level.

Figure 2.10a. ALD of unmigrated line ARAL-1 (uninterpreted): Subvertical zones of low reflectivity like those seen at cmps 700 and 1050 are present throughout the data set. Line ARAL-1 of Profile ARAL can be migrated to investigate the preservation of the zones following the movement of dipping events adjacent to them. ALD = Automatic Line Drawing.



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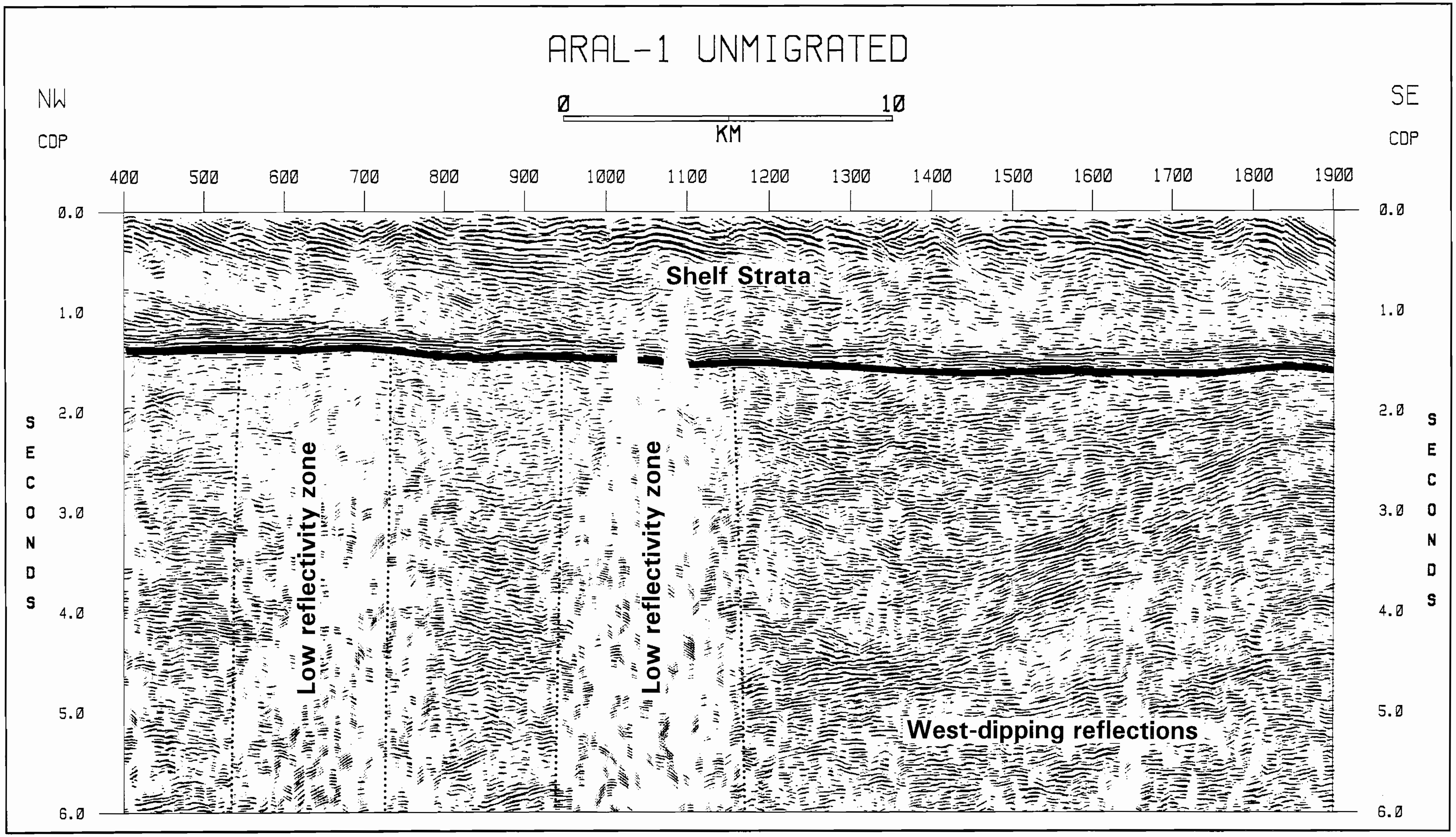
SECONDS

Figure 2.10b. ALD of unmigrated line ARAL-1 (interpreted): Subvertical zones of low reflectivity like those seen at cmps 700 and 1050 are present throughout the data set. Line ARAL-1 of Profile ARAL can be migrated to investigate the preservation of the zones following the movement of dipping events adjacent to them. ALD = Automatic Line Drawing.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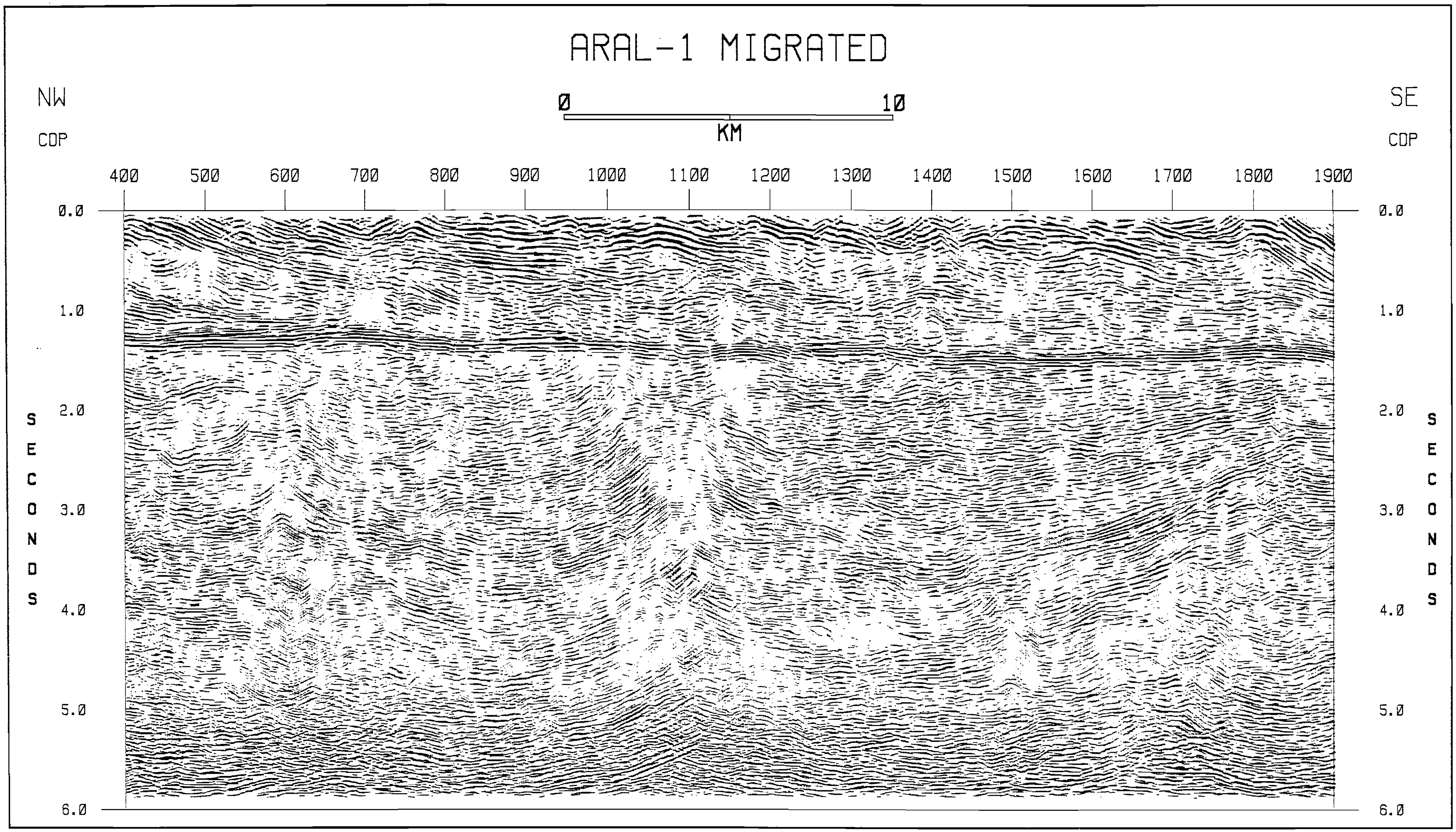
SECONDS
Shelf Strata
Low reflectivity zone
Low reflectivity zone
West-dipping reflections

Figure 2.11a. ALD of migrated line ARAL-1 (uninterpreted): Line ARAL-1 is long enough to be migrated to a depth of 6 seconds at the center of the line (see Appendix 2). The subvertical zones of low reflectivity at cmps 700 and 1050 are dominated by migration artifacts that have been enhanced by the line drawing processing. Some west-dipping reflections have migrated updip to into the zones while others have migrated away from the zones, resulting in a shift of the position of the low reflective zone at cmp 1100 to the southeast. The incomplete collapse of diffraction energy at the low reflective zone boundaries is interpreted as evidence for terminations of reflective surfaces responsible for the west-dipping reflections against subvertical dike swarms. The strong reflection at the base of the shelf strata at 1.4 seconds exhibits good amplitude recovery across the low reflective zone at cmp 1050 (compare to Figure 2.10). The improvement permits the interpretation that the subvertical bodies are older than the overlying shelf rocks. ALD = Automatic Line Drawing.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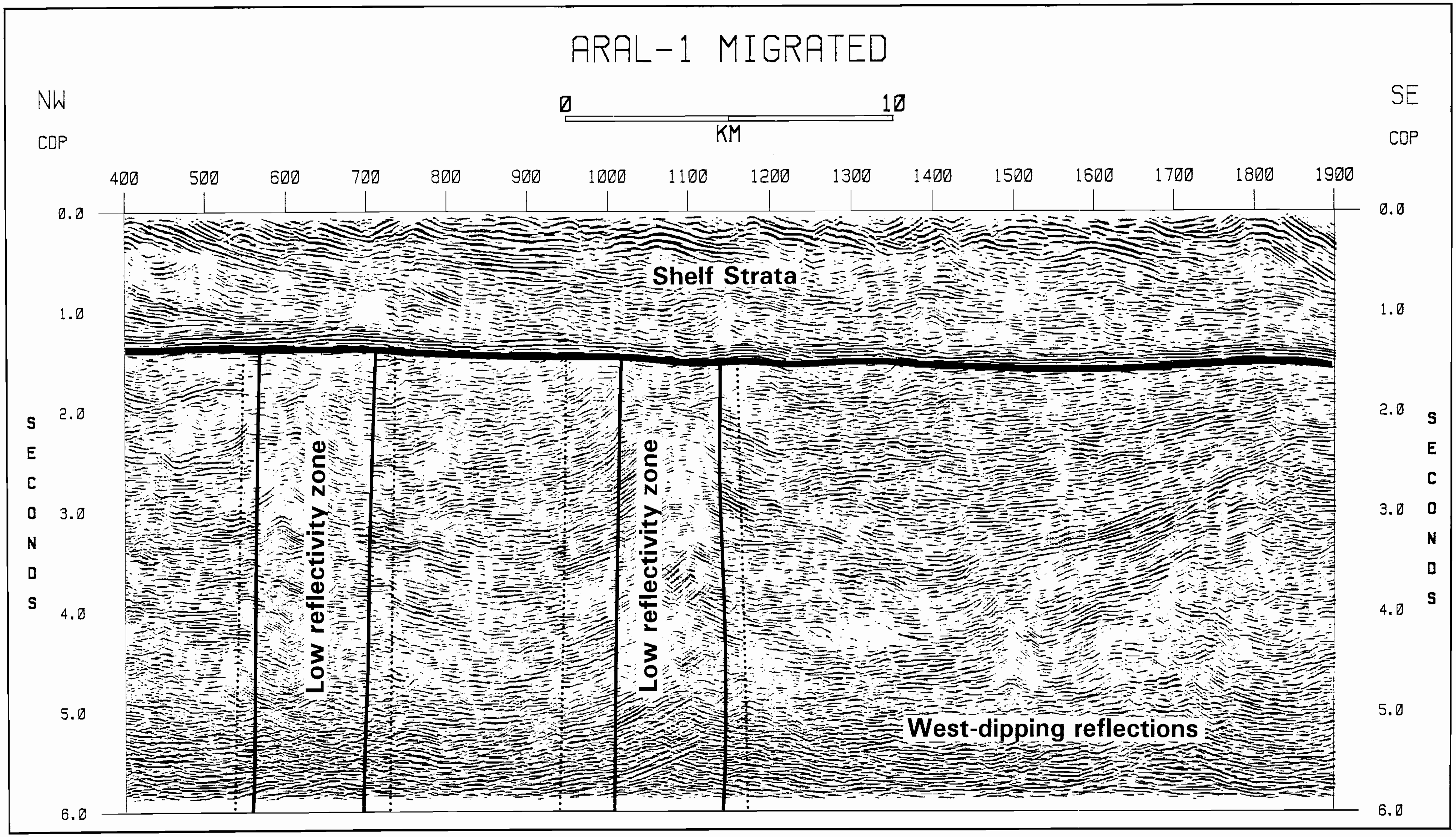
SECONDS

Figure 2.11b. ALD of migrated line ARAL-1 (interpreted): Line ARAL-1 is long enough to be migrated to a depth of 6 seconds at the center of the line (see Appendix 2). The subvertical zones of low reflectivity at cmps 700 and 1050 are dominated by migration artifacts that have been enhanced by the line drawing processing. Some west-dipping reflections have migrated updip to into the zones while others have migrated away from the zones, resulting in a shift of the position of the low reflective zone at cmp 1100 to the southeast. The unmigrated positions of the zones are shown as dotted lines and the migrated positions are shown as solid lines. The incomplete collapse of diffraction energy at the low reflective zone boundaries is interpreted as evidence for terminations of reflective surfaces responsible for the west-dipping reflections against subvertical dike swarms. The strong reflection at the base of the shelf strata at 1.4 seconds exhibits good amplitude recovery across the low reflective zone at cmp 1050 (compare to Figure 2.10). The improvement permits the interpretation that the subvertical bodies are older than the overlying shelf rocks. ALD = Automatic Line Drawing.



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SECONDS
SECONDS
Shelf Strata
Low reflectivity zone
Low reflectivity zone
West-dipping reflections

Summary

Reflection seismic profiles across the New York - Alabama Magnetic Lineament reveal the presence of discrete bodies and discontinuities within the crust beneath the late Precambrian and Paleozoic shelf strata. Features include a wedge-shaped block on profile ARAL typified by a region of low reflectivity of considerable areal extent. Pervasive west-dipping reflections appear beneath and to the southeast of the wedge on profile ARAL. Most of the reflection profiles image a subhorizontal band of higher reflectivity at middle crustal depths. Below the band, east-dipping reflections are imaged on the northwestern end of profile ARAL. The Moho can be seen on profiles ARDU and FM as a 1 second thick band of high reflectivity. The band exhibits dip of about 8^o on profile FM, suggesting that structure exists on the Moho in the northeastern part of the study area. The depth to the top of the Moho is 44 to 47 km using an average crustal velocity of 6500 m/s two-way travel-time justified by previous work (e.g. Prodehl et al., 1984) and lithologies present in the study area. Subvertical bands of low reflectivity visible on most of the profiles are probably real and can be interpreted to indicate the presence of dike swarms.

Chapter 3: Pseudomagnetic Field Investigations

Overview

The size and linear extent of the New York - Alabama Magnetic Lineament has provoked considerable speculation about the cause of one of North America's most pronounced geophysical phenomena. The Bouguer gravity data across the New York - Alabama Magnetic Lineament do not exhibit as distinctive a signature as the magnetic data, suggesting a lack of mafic material in conjunction with the strong magnetic anomaly. The relationship between the two potential fields might be more subtle than, for example, that accompanying the Mid-continent Rift System, where the strong magnetic signature is matched by a strong positive gravity signature.

The differences between the magnetic and the gravity fields in eastern Tennessee might reflect the presence of different sources within the crust or combinations of the same sources in which the relative contributions to the fields are different. The pseudomagnetic field technique provides a method for the extraction of the magnetic field that could be expected from the gravity data. In areas that have a pronounced magnetic signature and a nondescript gravity signature such as eastern Tennessee, the approach permits a comparison to be made to the more pronounced observed field. An advantage of using the pseudomagnetic field for comparison to the observed magnetic field over using the gravity field directly is that the pseudomagnetic field can be upward continued to the elevation at which the comparable aeromagnetic data were acquired. In addition, the magnetic anomaly shapes are usually asymmetric over sources due to the dipole effect. The shapes of pseudomagnetic anomalies are more directly comparable to the observed total intensity field. These advantages encourage the use of the pseudomagnetic technique in regions with a pronounced magnetic field.

The experimental approach used to determine the relationship of the gravity field to the magnetic field is outlined in Figure 3.1. The comparison of pseudomagnetic results with observed magnetic anomalies provides a means of explaining the differences between the potential fields and facilitates the development of better interpretations of the subsurface geology.

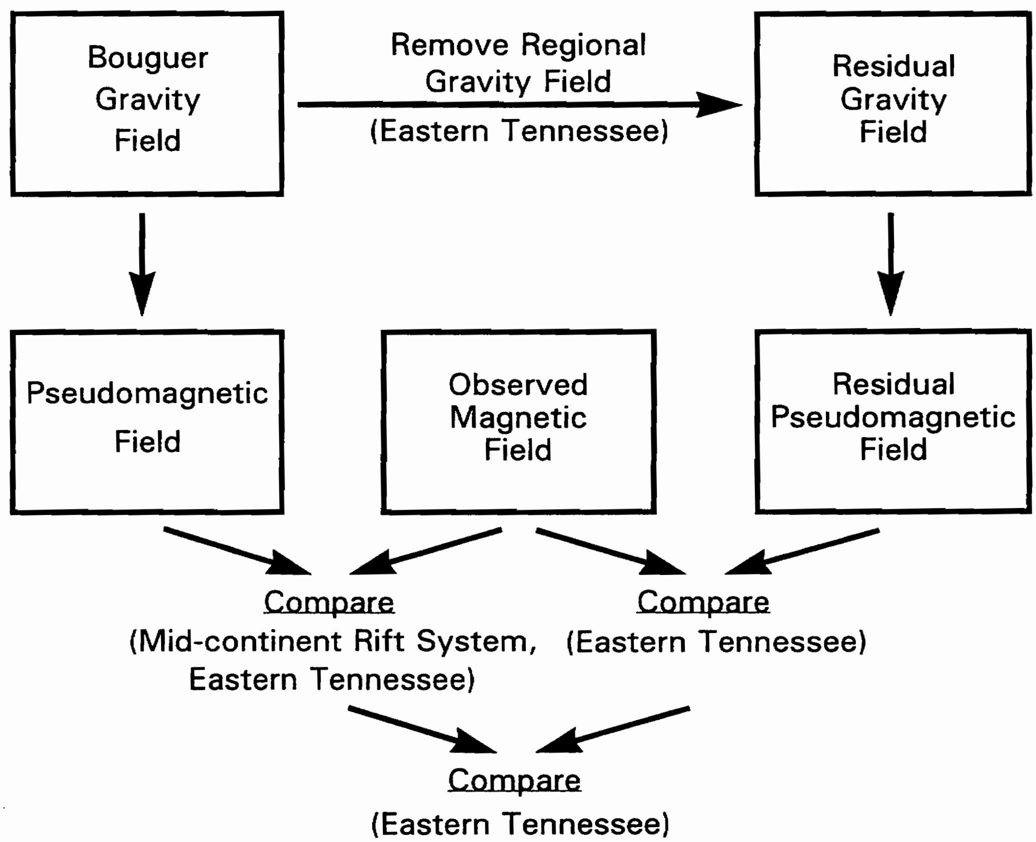


Figure 3.1. Experimental approach: The pseudomagnetic field is extracted from the gravity data and compared to the observed magnetic field. If appropriate, a residual gravity field is obtained by subtracting a regional field and the pseudomagnetic field obtained from the residual gravity field is compared to the observed magnetic field. If the residual gravity field yields a better fit to the observed magnetic field than the total gravity field, an interpretation of the sources of the magnetic anomalies can be better constrained.

Pseudomagnetic Fields

Theory

Robinson (1971) described a method based on Poisson's relation for extracting the pseudomagnetic field from gravity data. The extraction of the pseudomagnetic field from the gravity data addresses the implicit assumption common to many potential field investigations that dense rocks commonly have high magnetic susceptibilities; however, rocks can exhibit induced or remanent magnetism in the absence of density contrasts with adjacent units (e.g. Robinson et al., 1985).

Poisson's relation provides an explicit relationship between the gravity field and the magnetic field by relating density to susceptibility. Poisson's relation is given by:

$$A(x, y, z) = \frac{I}{G\rho}\frac{\partial}{\partial\alpha}U(x, y, z)$$

where A(x,y,z) is the magnetic field potential and U(x,y,z) is the gravity field potential at a point due to a source of uniform density, ρ, and uniform magnetism, I, in the direction α.

The results from simple modeling experiments performed by Robinson (1971) on synthetic data depict the accuracy of the pseudomagnetic technique in replicating the observed magnetic fields for a point source and for a semi-infinite slab. His results indicate that the pseudomagnetic field method should work well for situations where the sources of magnetic anomalies are expected to be related to sub-vertical dikes or similar isolated features. Robinson cautions that the pseudomagnetic field approach works best when the gravity anomaly approaches zero in an area smaller than the region covered by the two-dimensional operator.

Pseudomagnetic fields can be used to discriminate between localized sources and sources extending over large lateral distances. To test the applicability of this method, I used the Mid-continent Rift System (MCRS) as a possible group of point sources and the New York - Alabama Magnetic Lineament (NYAML) as a possible combination of point sources and subsurface bodies of large areal extent. If the Lineament and surrounding region is a combination of sources, the pseudomagnetic field should delineate the smaller bodies and less reliably reproduce the signatures of the sources of regional extent.

Sources of data used for pseudomagnetic field modeling

The data used as input to modeling were taken from the Geophysics of North America CD and include the Society of Exploration Geophysics (SEG) Bouguer gravity data base and the Decade of North American Geology (DNAG) aeromagnetic data base. The original station locations for the SEG gravity before gridding can be found on the Gravity Anomaly Map of the Eastern United States (Society of Exploration Geophysicists, 1982), and more complete discussions of gravity data acquisition can be found in Godson and Scheibe (1982) and Godson (1986). The primary source of data for the S.E.G. gravity map was the Defense Mapping Agency. The digital data were gridded to 4 X 4 km spacing, and can be expected to image bodies with diameters > about 10 km depending on the original station spacing, stated to be at least 5 minutes. The search radius in regions of sparse data was 40 km. Bouguer gravity reductions were performed using a density of 2.67 gm/cc. Terrane corrections in areas of substantial relief were performed by the U.S.G.S.

The DNAG digital magnetic data base has a grid interval of 2 km. The data were intended for use at a scale of 1:5,000,000 and are not reliable for the investigation of smaller anomalies. Sources of the magnetic data used in the compilation can be found on the data distribution index map with the Magnetic Anomaly Map of North America publication. The data were prepared using the Definitive Geomagnetic Reference Field (DGRF) for the data used in the compilation magnetic data. Other references include Godson

(1986). Grauch (1993) warned of problems associated with the DNAG digital magnetic data base. She discussed the presence of datum shifts among the surveys used to compile the data base of interest here. Grauch commented on the lack of resolution resulting from the wide spacing between flight lines relative to the flight elevations and warned against using the data set for the evaluation of small anomalies. Because the anomalies of interest to this study are large, the DNAG data base is expected to be adequate.

Reduction of the gravity and magnetic data for pseudomagnetic field evaluation consisted of converting the data to a common map projection, regridding the data to a common station spacing, extraction of the pseudomagnetic field from the gravity field, and comparison of the resultant pseudomagnetic map to the observed magnetic map. The map projections were converted from Albers equal area (gravity data) and from S.T.M. (magnetic data) to U.T.M. coordinates using the U.S.G.S. program UTILTRAN. Gravity and magnetic data sets were regridded to 1 km spacing to force the data points in each set to fall at the same spatial location. Spatial aliasing was avoided by using a large operator and by investigating only those anomalies large enough to be properly sampled by the 4 X 4 km grid.

Pseudomagnetic fields were calculated for an elevation of 1 km above the surface. For this study, a two dimensional operator of 41 points was determined to accurately represent the anomalies of interest to this study. Amplitudes of the pseudomagnetic field maps are not scaled to match those on the observed magnetic data, so product maps (e.g. Figure 3.5) generated to facilitate interpretation (Peavy et al., in prep). The product maps reflect a procedure described in Davis (1973) in which the mean is subtracted from each data point and the result divided by the standard deviation of the set. The approach provides a means of normalizing the calculated and observed values and facilitates the comparison of the fields.

Pseudomagnetic field of the Mid-continent Rift System

The Mid-continent Rift System provided an excellent target for testing the pseudomagnetic approach on a reasonably well understood geologic and geophysical feature. The rift system has been interpreted by most authors as an assemblage of mafic flows emplaced during incipient rifting (e.g. King and Zietz, 1971; Serpa et al., 1984; Behrendt et al., 1990). Behrendt et al. (1990) found evidence for crustal thicknesses up to as much as 58 km beneath the Rift System based on the GLIMPCE (Great Lakes International Multidisciplinary Program on Crustal Evolution) reflection seismic experiment. They indicated that these data corresponded well with previous refraction data from Halls (1982) and Luetgert and Meyer (1982), and assigned the thickness of the crust to underplating that effectively converted an Archean upper "seismic mantle" into Proterozoic lower "seismic crust".The prominent gravity and magnetic anomalies suggest a good correlation between the sources for the potential fields, and the interpreted limited areal extent of the sources is expected to generate reasonable pseudomagnetic fields.

The gravity field shown in Figure 3.2 is composed of a strong, essentially N-S trending high flanked by more isolated gravity lows. Values range from 30 mgal to -110 mgal over a 96 by 96 km grid. The strong low gravity values along the flanks of the high have been related to sedimentary deposits and might be related to the sedimentary basins reported by Chandler et al., 1982. Their model represents a good fit between the observed gravity field and the calculated field. They attributed the strong positive signature to a dense body that penetrates the crust from the near surface to Moho depths and the low flanks adjacent to the high as responding to shallow, low density bodies adjacent to the high density material. These data are consistent with interpretations by Behrendt et al. (1988).

The total intensity magnetic field is shown in Figure 3.3. The magnetic map, with a sample interval of 2 km, shows more detail than the more

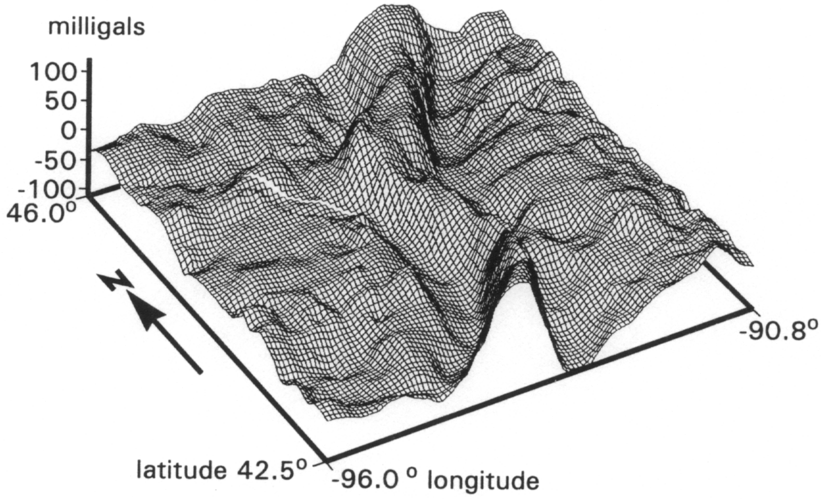


Figure 3.2. Mid-continent Rift System gravity field: The central ridge of high gravity values has been interpreted as sourced by mafic intrusions emplaced along the rift axis during Precambrian time. The gravity data were sampled at an interval of 4 X 4 km in the Geophysics of North America database.

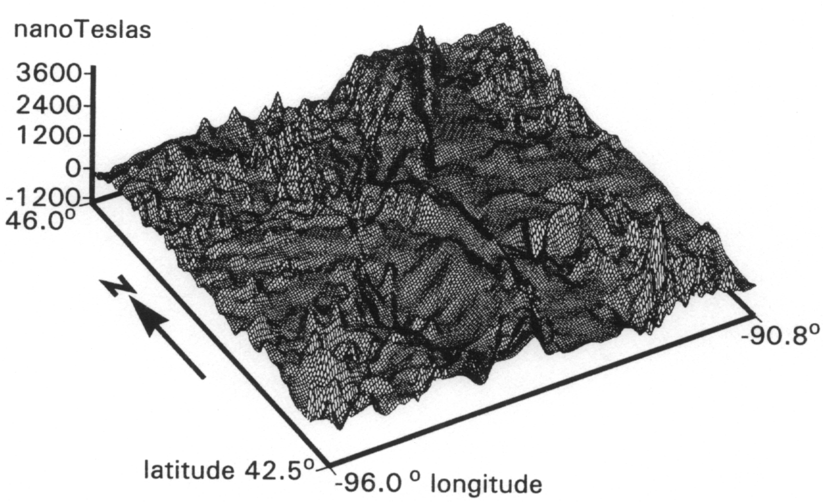


Figure 3.3. Mid-continent Rift System magnetic field: The central ridge of high magnetic values has been interpreted as being sourced by mafic intrusions emplaced along the rift axis during Precambrian time. An excellent correspondence between the gravity and magnetic fields suggests that the pseudomagnetic field will be comparable to the observed magnetic field. The magnetic data were sampled at an interval of 2 X 2 km in the Geophysics of North America database.

coarsely sampled 4 X 4 km gravity map, and indicates the essentially N-S trend of highly susceptible rocks along the axis of the MCRS. Values range from 2600 nT to -1200 nT over the grid. Several of the localized susceptible features exhibit the low-high pair expected for total intensity data at the magnetic inclination of 72^{o} for this latitude. These pairs are probably related to local subvertical intrusions, the amalgamation of which produces the strong regional signature of the MCRS. Anomalies outside the vicinity of the System are probably localized dikes.

A constraint on the source of the magnetic field is the magnetic "basement", which is defined to be the Curie point for the magnetically susceptible minerals responsible for the field. For magnetite, the Curie point is 575^{o} C, while the Curie point for ilmenite is variable depending on composition but much lower. Thus the contributors to the gravity field include deeper sources than those responsible for the magnetic field. In the MCRS, the consistency of the gravity and magnetic fields with the lithologies and structure of the rift feature reduces the chance that the two potential fields have different sources. Furthermore, the low heat flow value of less than 62 mW / m^2 (Morgan and Gosnold, 1989) deepens the level of the Curie point in the crust. Thus, the pseudomagnetic field extracted from the Bouguer gravity field is expected to mimic the observed magnetic field well.

Figure 3.4 is a display of the pseudomagnetic field over the MCRS. The regional presence of the high values along the axis of the rift is clear. Because the values are not scaled to match the gravity or magnetic field values, a normalized comparison of the pseudomagnetic field to the observed fields is required. The product map in Figure 3.5 provides a comparison, where positive correlations between the observed magnetic map and the pseudomagnetic map are indicated by high product values. Although remanent magnetism, found to be a strong contributor to the signature of the MCRS by King and Zietz (1971), was not considered in this study, the concentration of

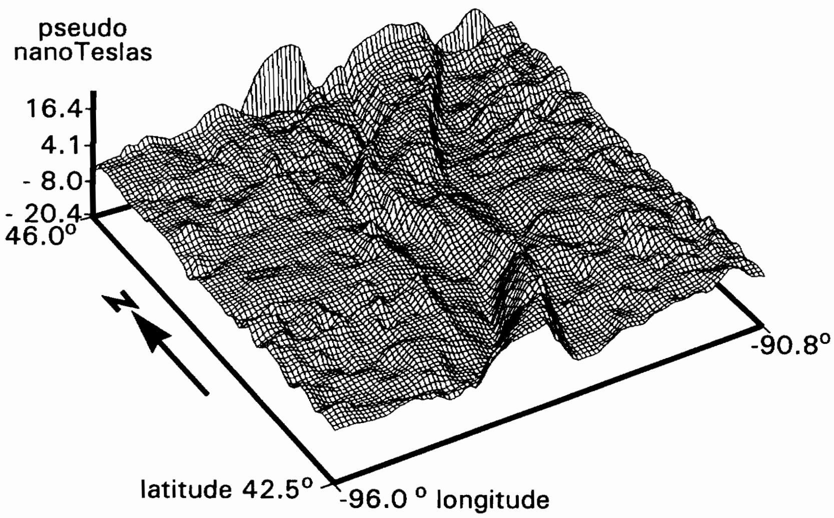


Figure 3.4. Mid-continent Rift System pseudomagnetic field: The gravity field was interpolated to a 1 X 1 km grid before the pseudomagnetic field was produced. The field was calculated using a 41 X 41 point operator following tests to ensure that the anomalies were properly sampled. The strong positive signature along the axis of the rift supports the hypothesis that the pseudomagnetic field can be used to approximate the magnetic field when sources satisfy Poisson's relation. The interpretation that the observed fields result from the presence of mafic intrusions is supported.

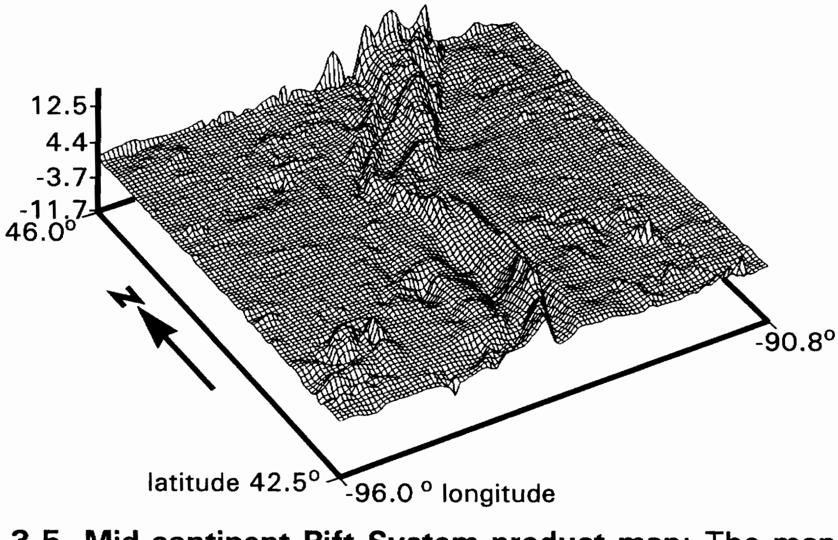


Figure 3.5. Mid-continent Rift System product map: The map is a normalized comparison of the pseudomagnetic map and the observed magnetic map. The high values along the axis of the rift indicate that the pseudomagnetic field has reproduced the magnetic map well, as expected for the strong correspondence between the signatures in the gravity and magnetic data.

high values within the MCRS indicates a good correspondence between the fields. Thus, the pseudomagnetic field approach to extracting the would-be magnetic field from the gravity field appears to have utility in the evaluation of the correspondence between gravity and magnetic field sources.

Pseudomagnetic fields in eastern Tennessee

Pseudomagnetic fields were calculated for the Bouguer gravity map and for a residual gravity map. The comparison of both pseudomagnetic maps with the magnetic map permitted the elimination of dense, susceptible bodies as sources for the New York - Alabama Magnetic Lineament (NYAML). The residual map provided additional assurance that the gravity anomalies within the East-continent Gravity High (ECGH) west of the NYAML were accompanied by high amplitude, short wavelength magnetic anomalies.

The gravity map (Figure 3.6) reveals the ECGH, the gravity low along the Tennessee - North Carolina border, and a faint string of low values along the NYAML, shown by arrows. The interpretation that the ECGH reflects the presence of dense bodies in the subsurface is corroborated by the surface exposure of a peridotite dike in the area. The high magnetic values in the north-central part of the map correspond in position to the ECGH. The anomalies are typified by short wavelengths suggesting localized sources like those expected for small, subvertical mafic dikes.

The NYAML is too broad and continuous to be sourced by small, near surface anomalies (King and Zietz, 1978), and forms a steep gradient between high magnetic values on the northwest and low values to the southeast (Figure 3.7). The gradient separates magnetic susceptibility values as high as 2600 nT from values less than zero over a distance of 25 km. The magnetic gradient corresponds to a string of gravity lows to -70 mgal from a regional value of about -55 mgal. The low gravity values in combination with the strong magnetic signature suggests the juxtaposition of felsic, magnetically susceptible rocks against felsic, less susceptible rocks.

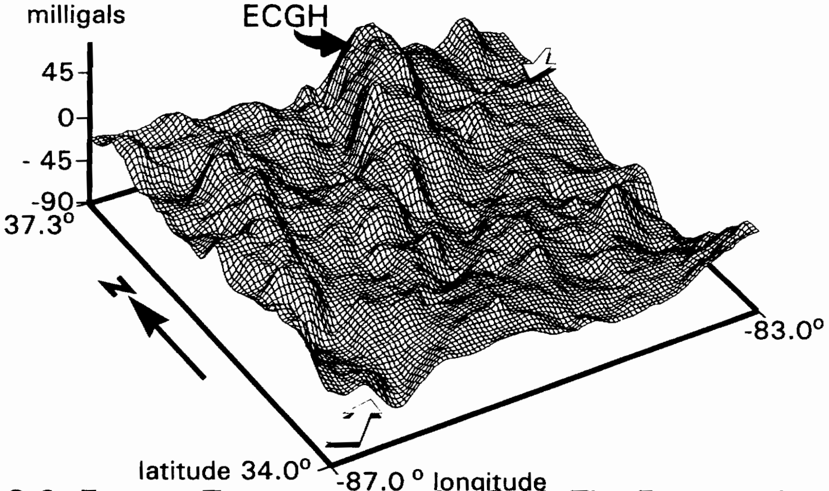


Figure 3.6. Eastern Tennessee gravity field: The East-continent Gravity High (ECGH) is the prominent ridge near the upper center of the map. The string of gravity lows can be seen in line with the New York - Alabama Magnetic Lineament indicated by the arrows. The gravity data were sampled at an interval of 4 X 4 km in the Geophysics of North America database.

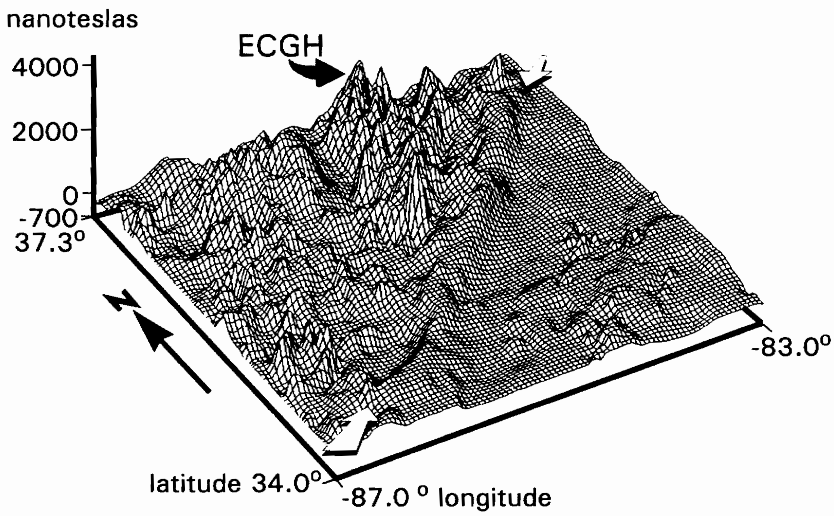


Figure 3.7. Eastern Tennessee magnetic field: The field exhibits short wavelength, high amplitude anomalies in the north-central part of the map that correspond in location to the East-continent Gravity High (ECGH). The New York - Alabama Magnetic Lineament, shown by the arrows, is characterized by a steep gradient trending N45°E and has the long wavelength suggestive of a deep source. The presence of a low gravity signature along the NYAML suggests that the source does not satisfy Poisson's relation. The magnetic data were sampled at an interval of 2 X 2 km in the Geophysics of North America database.

The pseudomagnetic field in Figure 3.8 does not show a strong correlation with the observed magnetic map except in the vicinity of the East-continent gravity high. The product map in Figure 3.9 supports these observations, where the high values are located almost exclusively in the northwestern portion of the map. A faint negative correlation can be seen extending along the position of the New York - Alabama Magnetic Lineament and can be attributed to the presence of the string of faint gravity lows. A localized negative correlation value can be seen in the upper right corner of the map area. This value correlates with a strong magnetic signature in eastern Kentucky that does not have a corresponding gravity signature. The small magnetic anomaly might indicate the presence of a body that is too small to be resolved by the more coarsely sampled gravity data set.

Removal of a regional gravity signature from the Bouguer gravity map permits the extraction of a pseudomagnetic field from the residual gravity field. The removal eliminates contributions to the gravity field from deep crustal sources. The presence of surface exposures of peridotite dikes and kimberlites suggests that those sources might be better represented on a residualized map than on a map of the Bouguer gravity field. Therefore, the pseudomagnetic field map generated from a residual gravity map might be better for interpreting the smaller sources.

The crust in eastern Tennessee has been determined by numerous studies to be from 42 to 50 km thick (e.g. Prodehl et al., 1984; Long and Liou, 1986; James et al., 1968; Tatel et al, 1953). The topography of the Moho can be approximated from these studies and from reflection seismic data reprocessed as part of this research. Figure 3.10a illustrates the combined results from refraction profiles, earthquake arrival times (P_n), crustal modeling, and reflection profiles and suggests a general trend for the Moho that rises from more than 50 km near the southern boundary of Tennessee to less than 45 km toward the north.

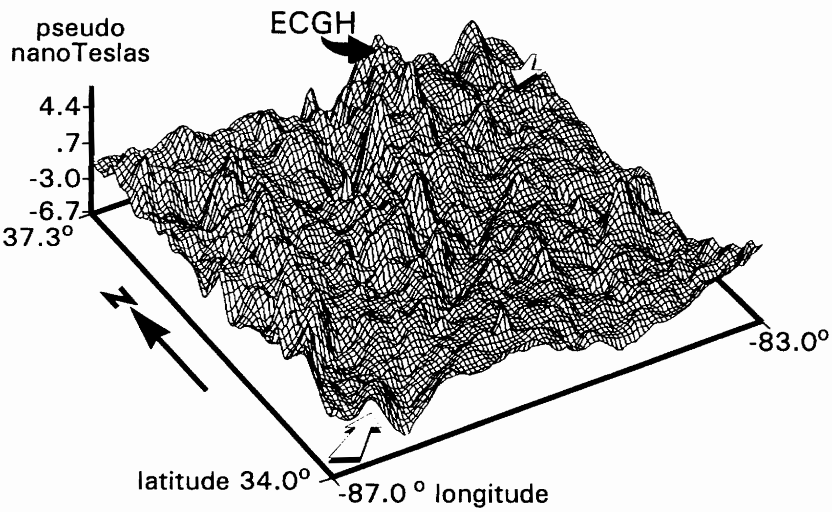


Figure 3.8. Eastern Tennessee pseudomagnetic field: The pseudomagnetic field reproduces the short wavelength anomalies in the north-central part of the map, but does not reproduce the signature of the New York - Alabama Magnetic Lineament (arrows). The field was calculated using a 41 X 41 point operator.

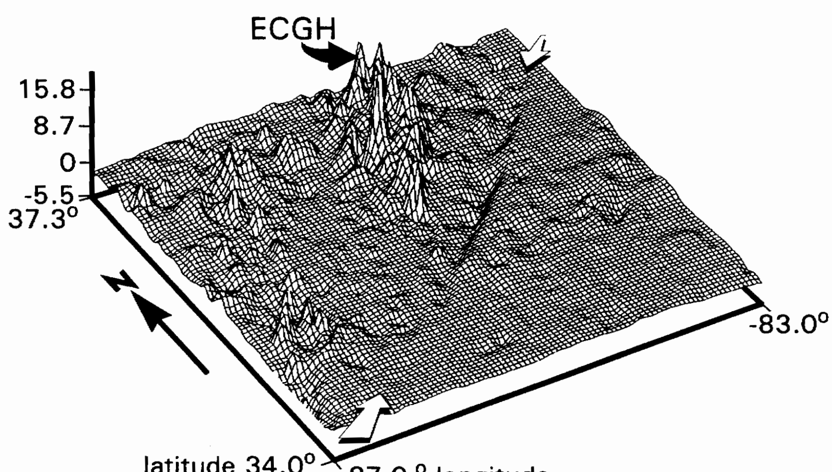


Figure 3.9. Eastern Tennessee product map: The map is a normalized comparison of the pseudomagnetic map and the observed magnetic map. The high values in the north-central part of the display indicate that the pseudomagnetic field has reproduced the magnetic map well in that area. The absence of positive signatures along the New York - Alabama Magnetic Lineament (arrows) indicates low correspondence between the maps. Low values suggest that the gravity field has an inverse Poisson's relationship to the magnetic field.

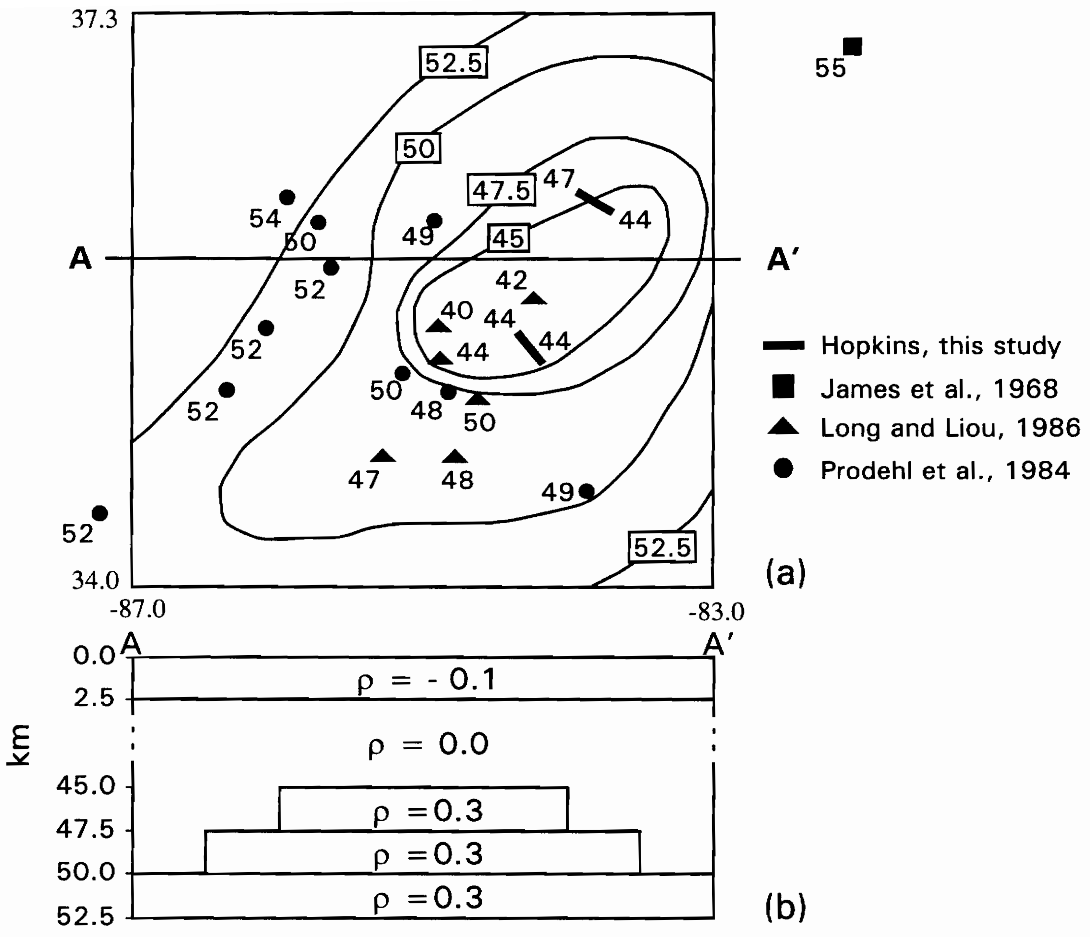


Figure 3.10. Regional gravity model: The model is developed from independent data sources including refraction and reflection seismic investigations. The model represents the presence of a thinner region of crust within the study area in comparison with the overall 50 km thick crust. The contour map in (a) illustrates the ridge of thinner crust trending approximately N50^{o}E. In (b), the layers along the profile A - A' are indicated in which the densities shwon are differences from the standard crustal density of 2.67 gm/cc. The shallow layer is used as a scale factor.

Values obtained from the refraction and reflection data were hand-contoured to develop the model shown in Figure 3.10a. The contour map indicates the presence of two thinner regions of the crust connected to form a ridge trending N50°E, approximately the same azimuth as the crustal thickness maps of Taylor (1989) and James et al. (1968). Five layers define the model in Figure 3.10b. All layers are 2.5 km thick, with the tops of the shallowest layers at depths of 45 km, the tops of layers 2 and 4 at 47.5 km, and the top of layer 5, the solid layer extending across the study area, at 50 km. The base of the model was set at 52.5 km. The layers were assigned densities of 0.3 gm/cc above 2.67 gm/cc to represent the shallower depth of the upper mantle. The 20 km thick upper layer is used as a scale factor and is assigned a density contrast of -0.1 gm/cc.

The model formed the input to a three dimensional gravity modeling program written by Edwin S. Robinson at Virginia Tech using the equations of Plouff (1976). The regional gravity field obtained from 3DGRAV is shown in Figure 3.11. The values obtained for the regional field range from -41.05 mgal along the crest of the gravity ridge to -59.95 mgal along the flanks.

The regional gravity field was subtracted point by point from the observed gravity field to produce the residual field. The residual gravity field (Figure 3.12) indicates the presence of high gravity values up to 60 mgal in the vicinity of the ECGH and a broad regional low to the southeast. The high values suggest the presence of a ridge of dense material with a markedly different trend from the deep ridge modeled for the base of the crust. The East-continent Gravity High trends approximately N25°E in contrast with the ridge on the Moho which trends N50°E.

The pseudomagnetic field calculated for the residual gravity field is shown in Figure 3.13. High magnetic susceptibility values in the vicinity of the East-continent Gravity High are visible. A comparison of the residual pseudomagnetic field map with the pseudomagnetic field map derived from

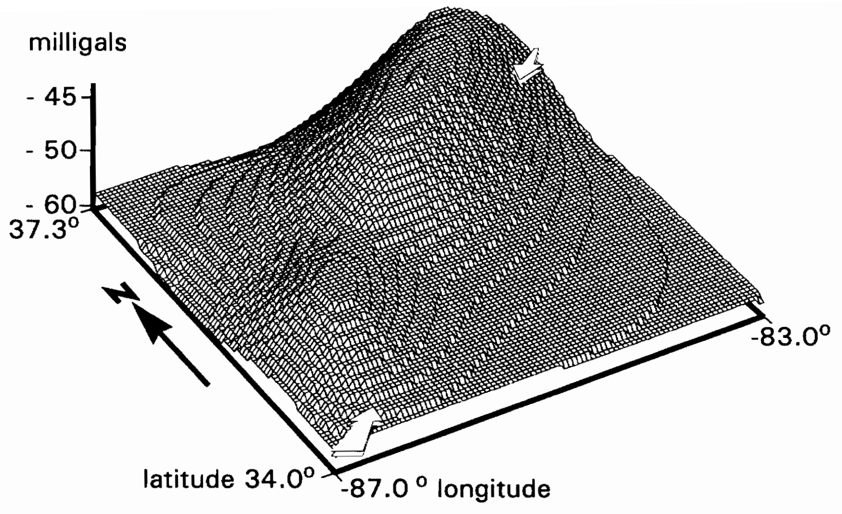


Figure 3.11. Regional gravity field: The regional field is based on crustal thickness values shown in Figure 3.10. Two regions of thinner crust can be inferred from the limited data and form a ridge trending N50°E, subparallel to the trend of the New York - Alabama Magnetic Lineament shown by the arrows.

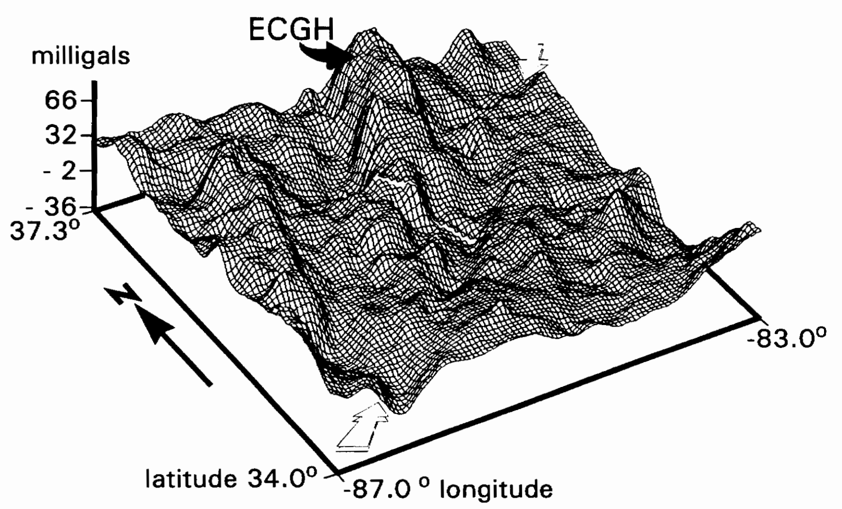


Figure 3.12. Residual gravity field: The field results from the removal of the regional field in Figure 3.11 from the Bouguer gravity map shown in Figure 3.6. The map differs from the Bouguer map mainly as a bulk shift to higher gravity values. The East-continent Gravity High (ECGH) is visible in the north-central part of the map. The position of the New York - Alabama Magnetic Lineament is shown by the arrows.

the total Bouguer gravity map (refer to Figure 3.6) reveals that the two maps are similar. The removal of the regional field from the total Bouguer gravity map has not strongly affected the appearance of the shallower sources of gravity anomalies

The correspondence of the ridge of high values is clearly imaged on the product map in Figure 3.14. A comparison of product maps for the residual field and for the total field suggests that no better correspondence can be found between the residualized pseudomagnetic field and the observed magnetic map than appears between the pseudomagnetic field map derived from the total Bouguer gravity map. The absence of improved correspondence between the residual and Bouguer gravity pseudomagnetic fields indicates that a deeper crustal contribution to the Bouguer gravity field does not mask a shallower gravity signature that would yield a pseudomagnetic field similar to the magnetic field of the NYAML. Hence, the sources of the small anomalies in the vicinity of the ECGH are interpreted to be relatively shallow localized bodies of strong density and susceptibility contrast with the surrounding crust, while the New - York Alabama Magnetic Lineament appears to be sourced by a regionally extensive body of relatively low density compared to the crust to the southeast of the steep magnetic gradient.

Summary

The correspondence between the pseudomagnetic field and the observed magnetic field in eastern Tennessee suggests that the sources of the East-continent Gravity High are shallow, localized, and satisfy Poisson's relation by being both dense and magnetically susceptible. The presence of the Norris Lake peridotite in eastern Tennessee, characterized by a localized gravity and magnetic high, supports the interpretation that the gravity and magnetic signatures over the ECGH are produced by buried mafic intrusions.

The residual pseudomagnetic field does not bear a stronger correspondence to the New York - Alabama Magnetic Lineament than the pseudomag-

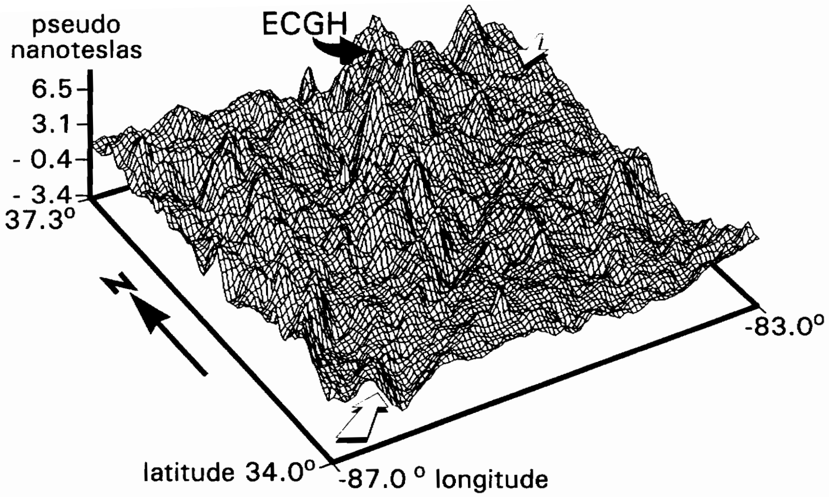


Figure 3.13. Eastern Tennessee residual pseudomagnetic field: The pseudomagnetic field does not differ significantly from the field calculated using the Bouguer gravity field (compare to Figure 3.8). The similarity indicates that the thickness variation in the crust did not affect the pseudomagnetic field calculation. The New York - Alabama Magnetic Lineament is indicated by the arrows.

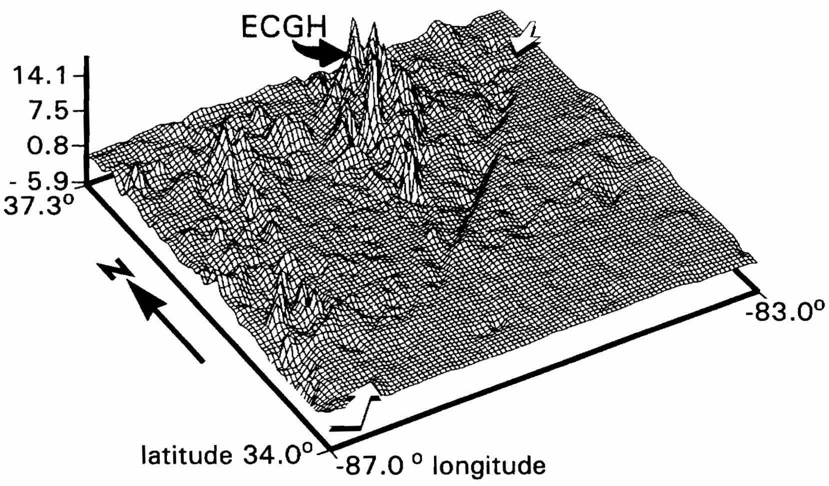


Figure 3.14. Eastern Tennessee residual product map: The map is a normalized comparison of the residual pseudomagnetic map and the observed magnetic map. The high values in the vicinity of the East-continent Gravity High remain a positive correlation following the removal of the crustal thickness variation. The similarity of the residual product map with the product map calculated for the Bouguer gravity pseudomagnetic field indicates that the effect of thin-

netic field determined from the Bouguer gravity. The lack of improvement following the removal of a regional gravity field is interpreted to mean that the low gravity values along the Lineament are indicative of the relative density of the magnetically susceptible rocks responsible for the NYAML. This correspondence suggests that the magnetically susceptible bodies in that region are laterally extensive and do not satisfy Poisson's relation. Hence the source of the New York - Alabama Magnetic Lineament is appropriately interpreted as evidence of a contact between crustal blocks of granitic or granodioritic compositions with different magnetic susceptibilities.

Chapter 4: Gravity and Magnetic 2-D Models

While three dimensional modeling is necessary to characterize essentially non-linear gravity and magnetic signatures, the New York - Alabama Magnetic Lineament is exceptional in its length and continuity and lends itself to two dimensional modeling. Reflection seismic data across the lineament provide an opportunity to constrain the geometries of bodies beneath the Alleghanian allochthon and Cumberland Plateau in eastern Tennessee and facilitate the selection of the most appropriate models among several that satisfy the potential field data. Reflection profile ARAL (Figure 2.2) is the most illuminating and reveals the presence of distinctive features within the crystalline basement. Two dimensional modeling coupled with the reflection data suggests compositional variations among the features.

Method

Modeling was carried out at Virginia Tech using the commercially available computer modeling software GM-SYS, version 1.89b, marketed by Northwest Geophysical Associates, Corvallis, Oregon. Observed magnetic values were obtained from the Residual Total Intensity Aeromagnetic Map of Tennessee by Johnson et al. (1979) (Figure 4.1). The data used for magnetic modeling along profile ARAL were flown by GeoMetrics, Inc. in 1978, using a proton-precession magnetometer in a stinger configuration. Flight lines were spaced at 3.3 km, flown at a constant barometric altitude of 1 km above sea level, and readings were taken at intervals of approximately 100 m. The flight path was recovered using either a recording video camera or Doppler radar. Diurnal corrections were applied and the 1975 International Geomagnetic Reference Field (IGRF) was removed.

Gravity data were obtained from the Bouguer Gravity Anomaly Map of Tennessee by Johnson and Stearns (1967) (Figure 4.2). Stations, shown as dots on the map, were spaced at nominally 10 km in the vicinity of profile

Figure 4.1. Residual Total Intensity Aeromagnetic Map of eastern Tennessee: The map area shows the reflection profile ARAL composed of lines ARAL-1, ARAL-2, and ARAL-3. The magnetic profile was take from values along the line A-A'. The steep gradient of the New York - Alabama Magnetic Lineament is visible trending N45^{o}E. Contour interval 20 nt. Map from Johnson et al., 1979.

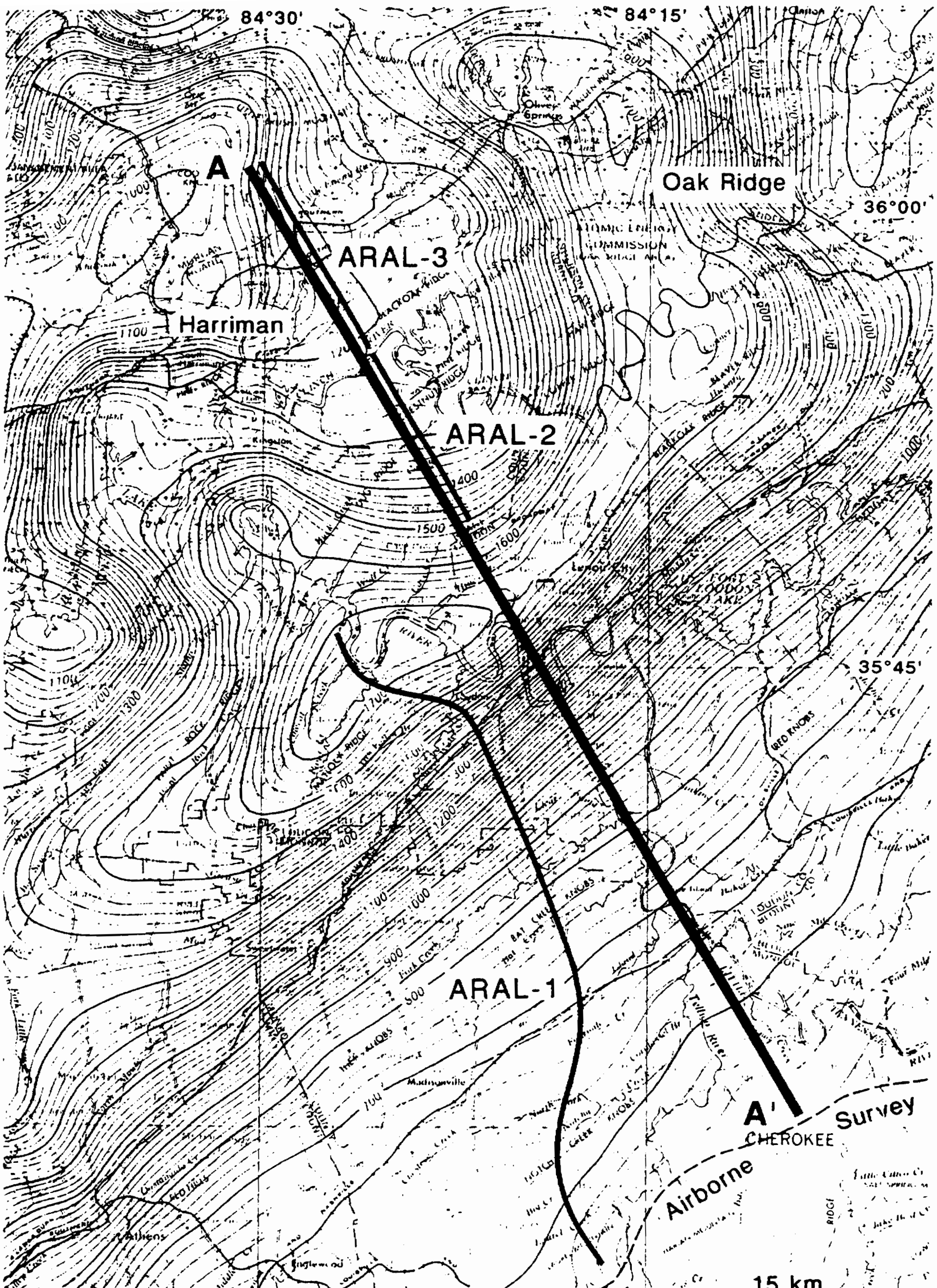
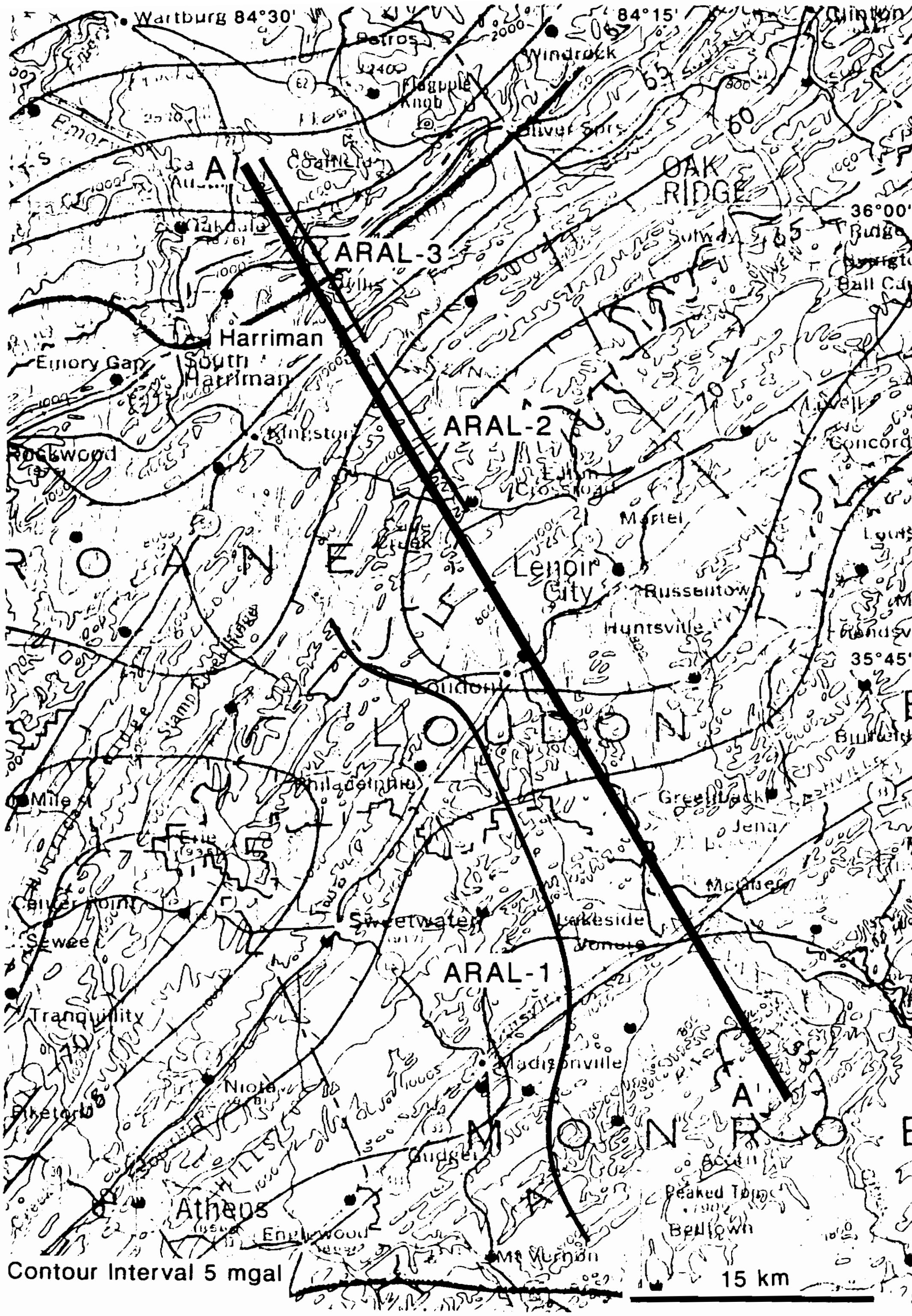

84°30'
84°15'
36°00'
35°45'
A
A'
ARAL-3
ARAL-2
ARAL-1
Oak Ridge
Harriman
CHEROKEE
Limit of Airborne Survey
15 km
Contour Interval 20 nT

Figure 4.2. Bouguer Gravity Anomaly Map of eastern Tennessee: The map area shows the seismic reflection profile ARAL, composed of lines ARAL-1, ARAL-2, and ARAL-3. The gravity profile was take from values along the line A-A'. The position of the New York - Alabama Magnetic Lineament is shown on the map trending N45^{o}E along a group of gravity lows (hachured contours). Gravity stations are shown as dots. Contour interval = 5 mgal. Map from Johnson and Stearns, 1967.


Wartburg 84°30'
84°15'
Clinton
Petros
Windrock
Oliver Spgs.
OAK RIDGE
36°00'
A
ARAL-3
Harriman
Emory Gap
South Harriman
Kingston
ARAL-2
Rockwood
R O A N E
Lenoir City
Martel
Russelltown
Huntsville
Concord
35°45'
Loudon
L O U D O N
Philadelphia
Greenback
Jena
Sweetwater
Lakeside
Vonore
ARAL-1
Sewee
Tranquillity
Niota
Madisonville
A'
M O N R O E
Athens
Englewood
Mt Vernon
Peaked Top
Belltown
Contour Interval 5 mgal
15 km

ARAL. No terrain corrections were applied to the data, and the standard reduction density of 2.67 gm/cc was used in producing the gravity map.

Gravity and magnetic values were taken from the maps at 2.5 km intervals along a straightened approximation to the trend of profile ARAL, shown as A - A′ on Figures 4.1 and 4.2. The reflection seismic line ARAL is shown as a crooked line following the roads along which it was acquired. The trend of the New York - Alabama Magnetic Lineament in eastern Tennessee is N45^{o}E and forms the basis for assigning the trend of the two dimensional profile to an azimuth of 315^{o} (perpendicular to the trend of the lineament). The Earth's field was set at 53500 nT, the inclination used was 65^{o}, and the declination was - 4^{o}.

Forward modeling proceeded from simple to more complex subsurface geometries. The geometries interpreted on profile ARAL and used to constrain the models are shown in Figure 4.3. An effort was made to adhere to the geometries visible on the reflection seismic data with the recognition that the seismic impedance contrasts definitive of the subsurface bodies might not be indicative of susceptibility contrasts. In the development of the gravity models, the values input to the program represent density differences from 2.67 gm/cc, the standard Bouguer gravity correction value. Remanence was not considered for these models, but could improve the results for future investigations. The following models represent density and susceptibility values and geometries capable of reproducing the observed gravity and magnetic fields.

Description of models and profiles

The simple model exhibited in Figure 4.4 is composed of two bodies. The Alleghanian allochthon is shown at the top thickening from 3 km on the northwest end of the profile to 8 km on the southeast end above the older crustal rocks. The depth to the Moho in this and subsequent models is 50 km, consistent with refraction results (e.g. Prodehl et al., 1984). Susceptibilities range from 0.0600 for the allochthon to 0.1130 for the crust below. The fit

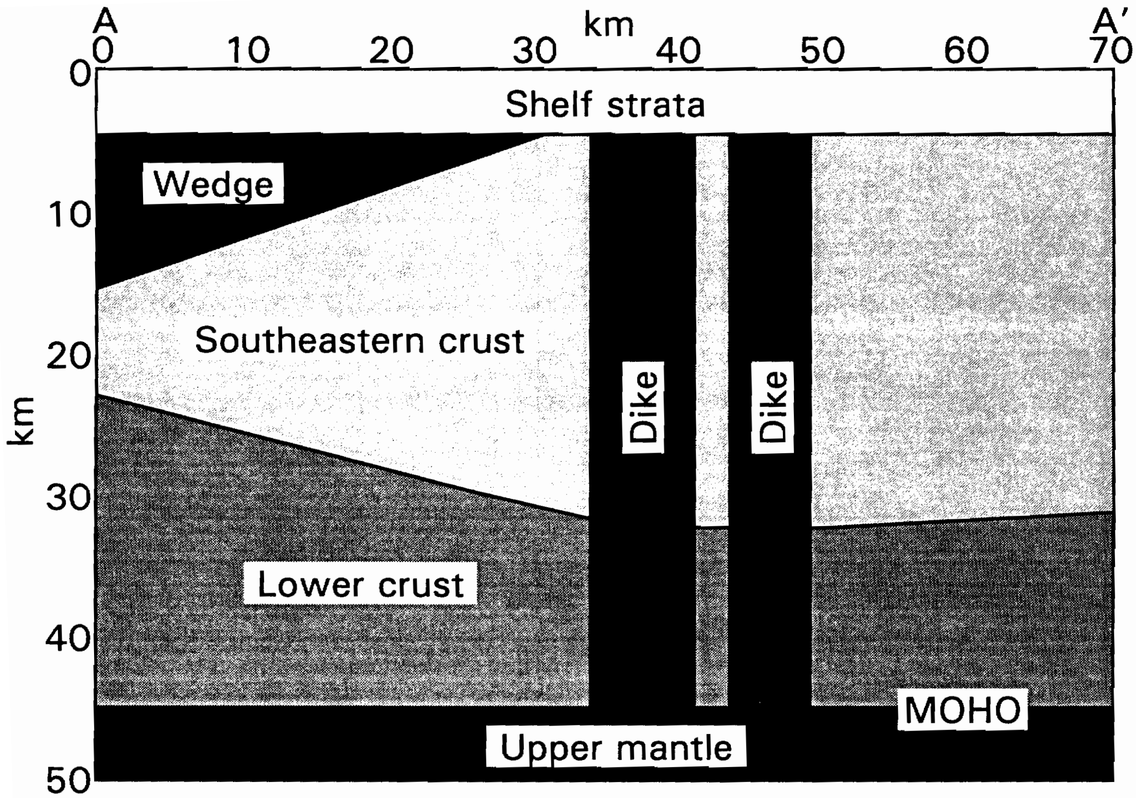


Figure 4.3. 2-D model geometries: The bodies shown above are visible on reflection profile ARAL (Figure 2.2) and have been projected onto profile A - A', Figures 4.1 and 4.2. The crustal blocks differ acoustically from each other and might have compositional differences sufficient to produce the gravity and magnetic signatures observed along the New York - Alabama Magnetic Lineament. Because the reflection profile ARAL is crooked, the straightened projection A - A' is shorter.

between the observed magnetic values, shown as dots, and the calculated values, shown as a solid line is good. However, the calculated gravity profile does not reflect the gravity low in the observed data near the center of the profile. In addition, thickening of the shelf strata at the top of the model to 8 km on the southeast end of the profile is not supported by the seismic reflection data, where the maximum thickness of the upper layer is about 4.5 km assuming a two-way travel time of 5500 m/s for the shelf strata (1.6 seconds).

The model in Figure 4.5 consists of a constant thickness allochthon of 4 km (about 1.4 seconds) and two dikes within the deeper crust. The dikes are positioned to align with the subvertical low reflective zones on the reflection profile near the cmps 700 and 1100, where the low reflectivity might result from the presence of two dike swarms. In addition to the susceptibility contrast between the dikes and the surrounding crust, the model exhibits two crustal magnetic susceptibilities separated by the left edge of the left dike at about 37.5 km. The northwestern half of the crust is modeled as highly susceptible at 0.0240 while the southeastern half has a susceptibility of 0.0190. The crust between the dikes is of the latter type. The dikes are shown as less susceptible than either crustal type with susceptibilities of 0.0178.

In the gravity model, the crustal rocks are assumed to have the same density with significantly lower densities assigned to the dikes. The magnetic profile reveals that the peak in the observed values lies 5 km to the northwest of the peak in the calculated values, but has approximately the same amplitude and the same gradient. The gravity profile indicates a similar offset. The low value on the observed profile lies 20 km northeast of the calculated low. While a change in the position of the dikes can align the peaks, such a modification is not supported by the seismic data. Migrated profile ARAL (Figure 2.11) shows that the position of the southeastern dike has moved slightly to the southeast, exacerbating the poor fit to the gravity data. Shifting the dikes

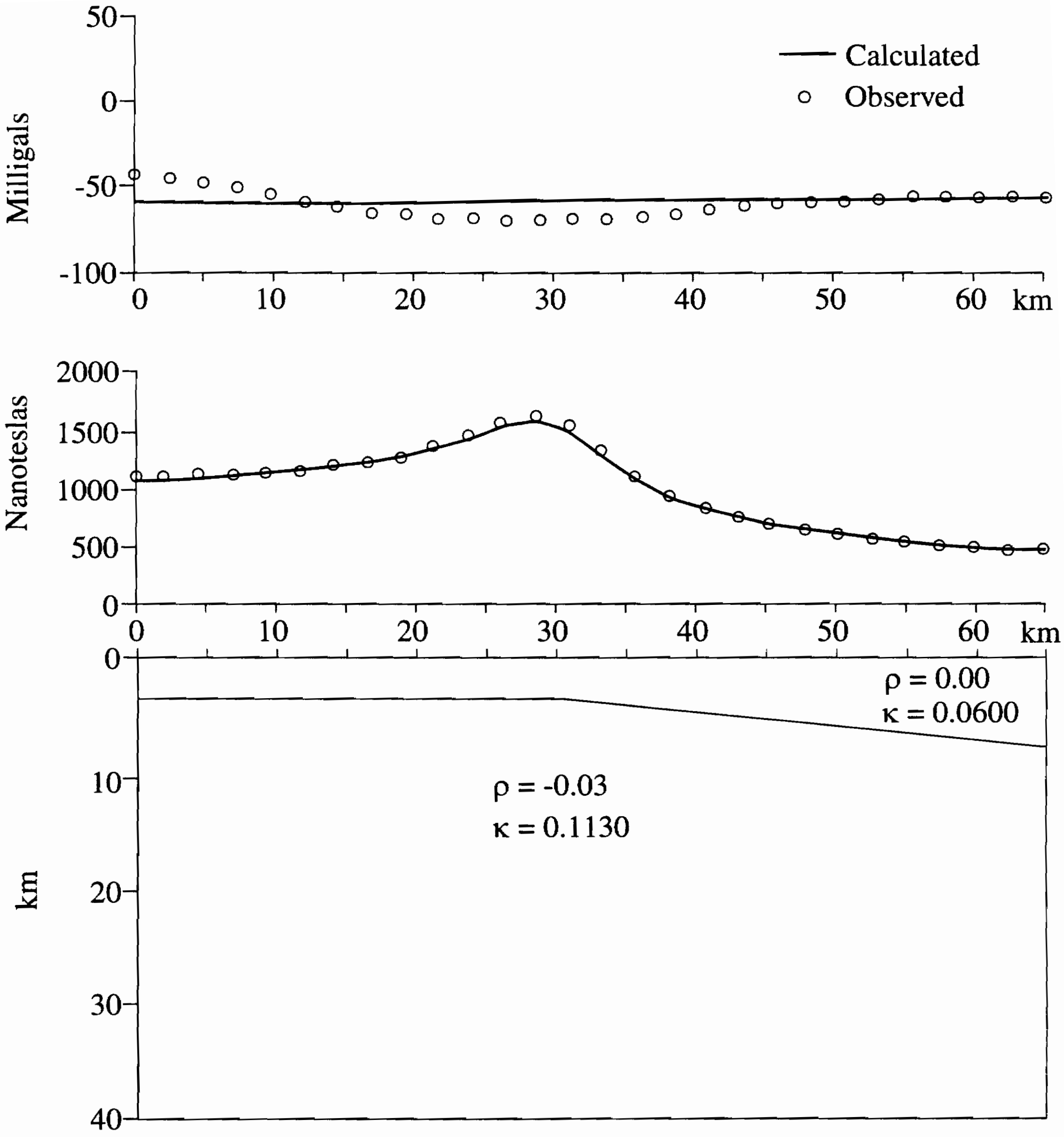


Figure 4.4. Two layer model: The simple model includes only density and susceptibility contrasts between the crust and the shelf strata above. The calculated magnetic values are reasonably accurate, but the gravity model fails to account for the gravity low observed in the data. In addition, the block containing the shelf strata has been thickened to the southeast beyond the limit allowed by the reflection data in order to produce the magnetic fit.

ρ = density difference from 2.67 gm/cc, κ = susceptibility

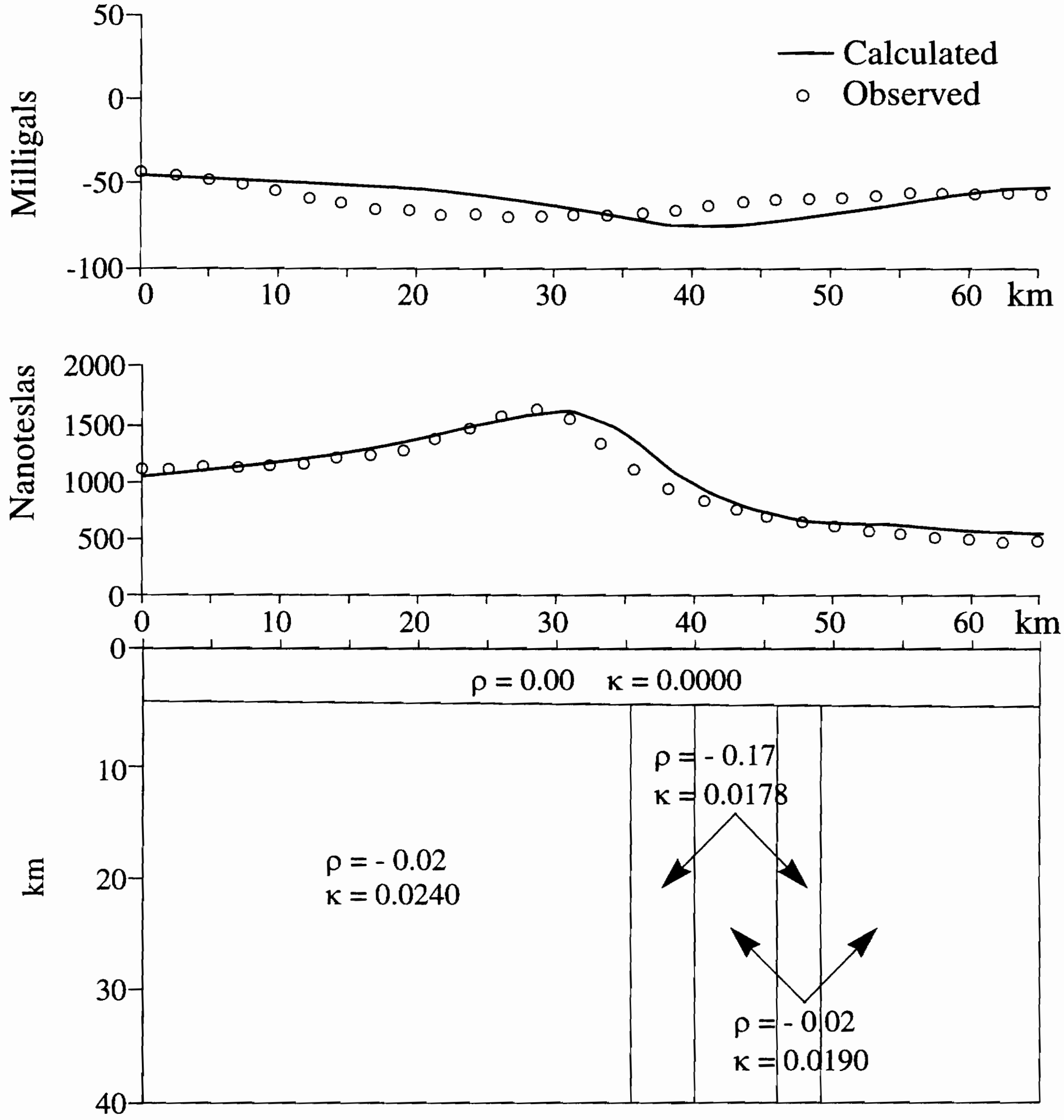


Figure 4.5. Two dike model: The subvertical zones of low reflectivity at cmps 700 and 1050 on profile ARAL are modeled as dikes. The best fit to the observed gravity and magnetic data has been obtained by assigning the dikes low densities and susceptibilities relative to the adjacent crust. Furthermore, the crust northwest of the dikes has a different susceptibility than that of the crust between the dikes and to the southeast. The calculated gravity and magnetic values do not accurately reproduce the observed data, but the curve shapes can be seen shifted to the southeast.

ρ = density difference from 2.67 gm/cc, κ = susceptibility to the northeast to produce the alignment would result in placement of the dikes in areas where reflection continuity is excellent, thereby ignoring a valuable constraint on the potential field models.

An alternative hypothesis suggested by the preceding model is that of two crustal rock types in vertical contact without the dikes (Figure 4.6). The model yields a magnetic shape that is broader than the peak of the observed profile, the magnitude of the calculated magnetic anomaly is larger, and the values do not tend be asymptotic with the northwestern and southeastern ends of the observed data, resulting in significant error. Furthermore, the model fails to reproduce the low values near the center of the observed gravity profile.

The following models revolve around the presence of the wedge observed on the reflection seismic data over the northwestern half of the profile. The wedge is a clearly distinct body on the seismic data and provides a good candidate for susceptibility and density contrasts within the crust.

The model displayed in Figure 4.7 is comprised of the wedge accompanied by dikes. The susceptibility values are the same as those for the dikes and crust in Figure 4.5 with the addition of a wedge with susceptibility 0.0245. Density contrasts are as strong between the wedge and the northwestern crust as they are between the northwestern crust and the southeastern crust containing the dikes.

The reproduction of the observed magnetic values by the calculated values is imperfect. The calculated peak values, while of essentially the same amplitude, are shifted a few km to the southeast and the calculated gradient on the northwest is not as steep as the observed gradient on that side. The calculated gradient on the southeast is somewhat steeper than that produced by the observed data. The calculated gravity profile is a good match for the observed gravity data. The calculated values effectively reproduce the shape of the gravity profile and position the gravity low at the correct location. This

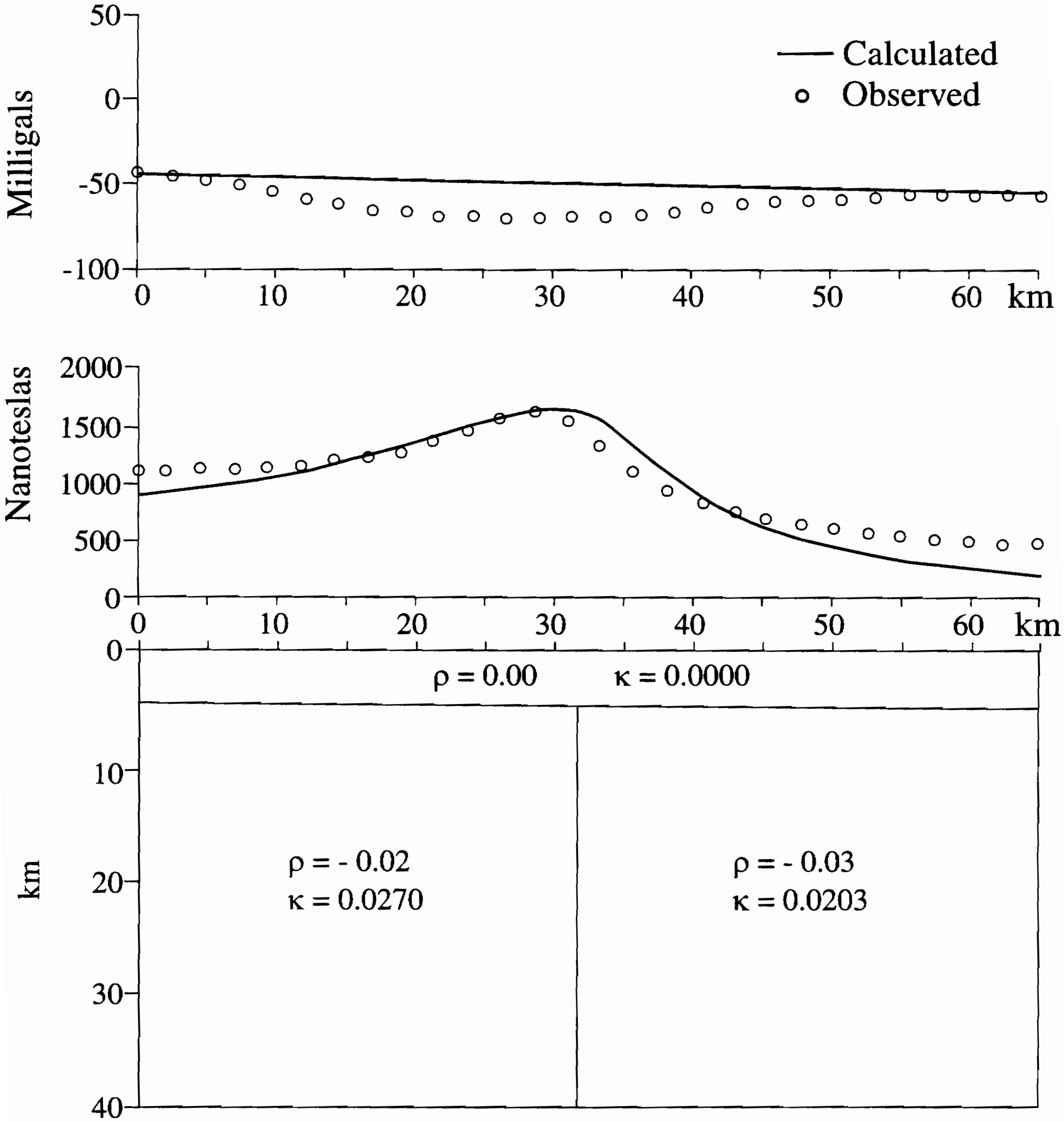


Figure 4.6. Two crustal blocks: The model illustrates a crustal suture along the flank of the subvertical zone of low reflectivity at cmp 700 on profile ARAL. The crust to the northwest is slightly denser and more susceptible than the crust to the southeast. The gravity data is not well reproduced by the model, and the peak in the magnetic data is shifted to the right in the calculated profile.

ρ = density difference from 2.67 gm/cc, κ = susceptibility

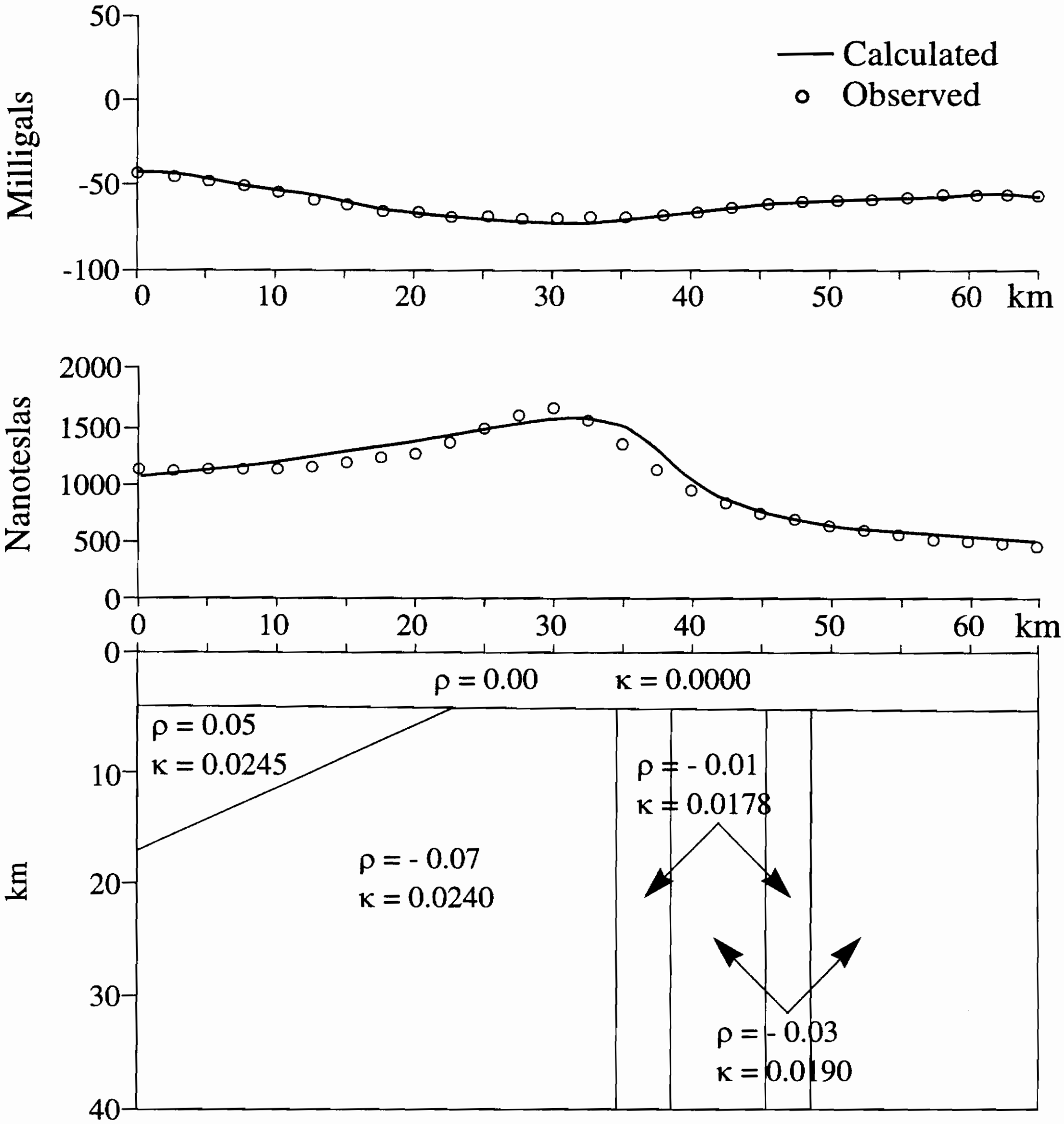


Figure 4.7. Wedge and dikes: The wedge appears on the northwestern end of the line beneath 4.5 km and is clearly image by reflection profile ARAL. The model illustrates density and susceptibility values for the wedge, the crust adjacent to the wedge, the dikes, and the crust between and to the southeast of the dikes. The calculated gravity low and magnetic peak are shifted to the right from the observed profiles.

ρ = density difference from 2.67 gm/cc, κ = susceptibility

model is a significant improvement over the previous dike model in imaging the gravity structure of the crust, but both dike models fall short in providing an accurate representation of the observed magnetic profile.

The model displayed in Figure 4.8 shows the wedge beneath the allochthon on the northwestern end of the line with a susceptibility of 0.0520. The susceptibility of the lower crust beneath the middle crustal band is 0.0440. The middle crust has a susceptibility of 0.0450. The division between the lower and upper crust is based on the presence on the seismic data of a middle crustal band of higher reflectivity ranging in depth from 7 to 9 seconds (23 to 29 km) on profile ARAL. The model includes the layer of shelf strata, which is shown with a density contrast and a susceptibility of zero. This portrayal of the crust fits the observed magnetic data well. Calculated peak values fall atop the observed peak values and the steep gradient and flanking values are well represented. The gravity data are as well reproduced as are the magnetic data, but the calculated gravity low is slightly offset to the northwest from the observed gravity low. All of the boundaries in the model match boundaries observed on the seismic reflection profile ARAL.

The model displayed in Figure 4.9 is that of a broad, dipping band of high susceptibility (0.0600) between two regions of lower susceptibility (0.0500). The shelf strata are represented with zero susceptibility and density contrast. The left boundary of the higher susceptibility region corresponds to the position of the wedge boundary with the remainder of the crust. The right boundary of the region is speculative. Density contrasts shown on the model for the gravity profile are different for each body: the wedge is denser than 2.67 gm/cc while the dipping layer and the remainder of the crust is less dense than 2.67 gm/cc. Both the magnetic and gravity observations are well imaged by the calculated results from this model, although the presence of the southeastern boundary, hence the thickness, of the layer is unsubstantiated by the reflection data.

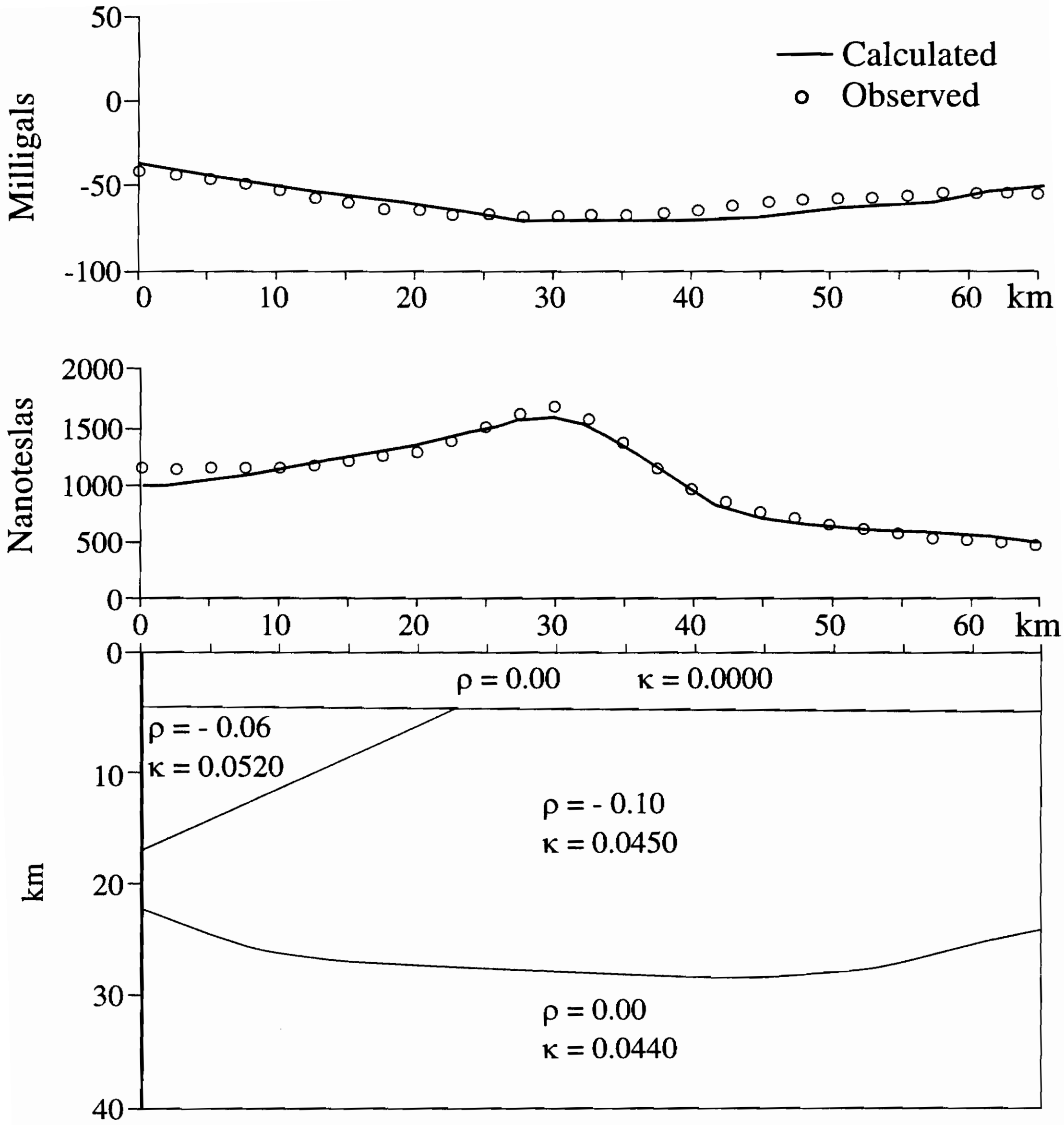


Figure 4.8. Wedge and layered crust: The densities assigned to the wedge and lower crust are higher than those assigned to the upper crust, while the susceptibility of the wedge is higher than that of the other bodies. Gravity and magnetic profiles calculated from this model produce an excellent fit to the observed data. This model is considered to be the best two-dimensional model to apply to the region in eastern Tennessee because all of the bodies shown are clearly delineated on the reflection profile ARAL (Figure 2.2).

ρ = density difference from 2.67 gm/cc, κ = susceptibility

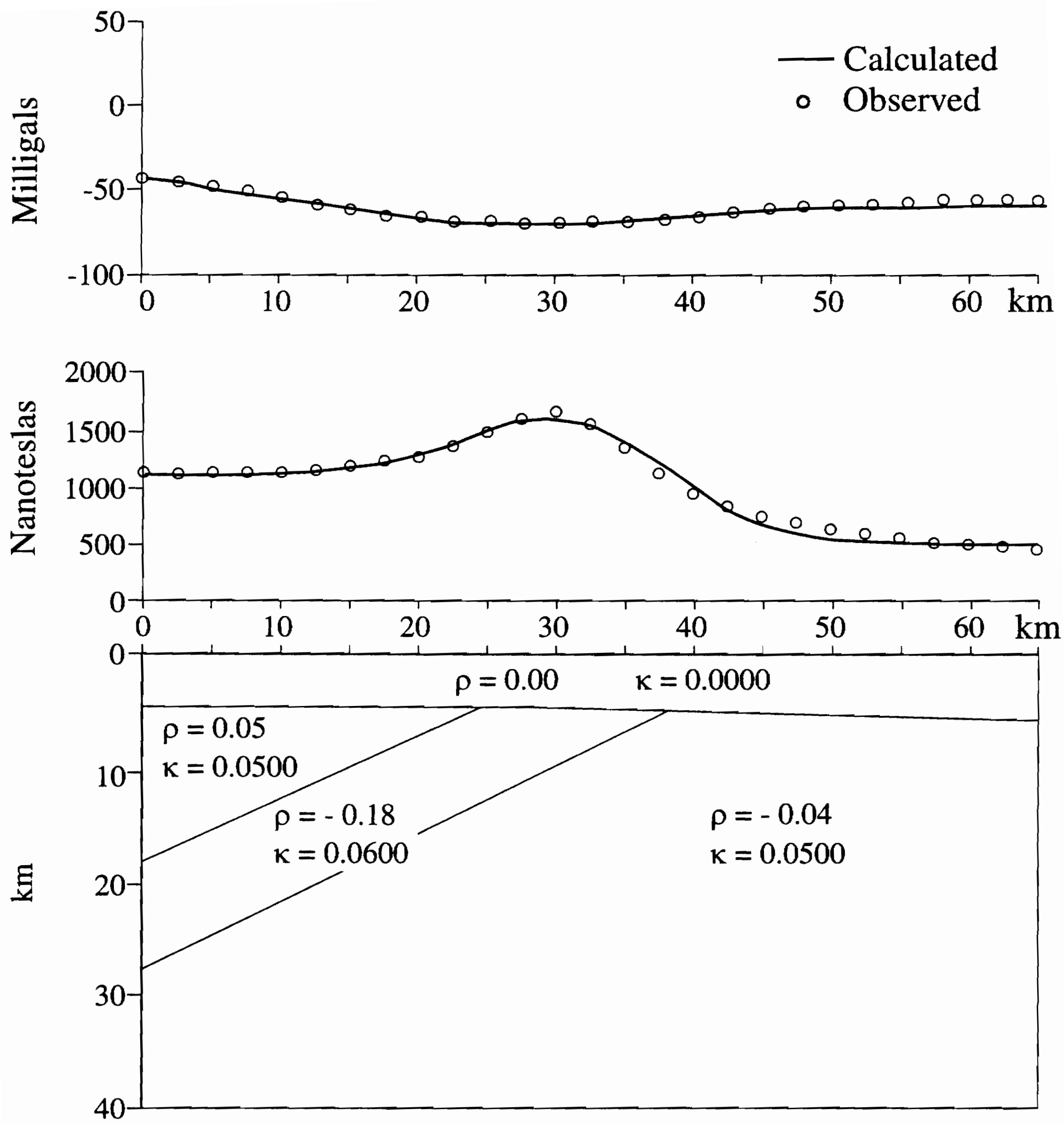


Figure 4.9. Tabular dipping layer: The calculated values from this model produce an excellent fit to the observed profiles; however, the wide dipping layer beneath the wedge is speculative. The top of the layer can be constrained by the base of the wedge on reflection profile ARAL, but its lower contact with the crust can not be seen. The wedge is shown as a high density block, while the layer is shown as a the most susceptible body.

ρ = density difference from 2.67 gm/cc, κ = susceptibility

The model in Figure 4.10 is similar to the previous dipping layer model except that the layer thins with depth toward the northwestern end of the line. The susceptibilities of each body are different as are the density contrasts. This model results in calculated values that match the observation closer than the preceding values, particularly along the southeastern flank of the steep gradient. Here again, the presence of the dipping layer and the position of its southeastern boundary, while honoring the dip of west-dipping reflections on the seismic data, are speculative.

Interpretation

The models in Figures 4.4 and 4.6 are included to illustrate the fits produced by these simple geometries to the potential field data while including only bodies visible on the reflection data. The dike models in Figures 4.5 and 4.7 do not accurately reproduce the magnetic profiles, but the model in Figure 4.7 results in a good fit to the gravity data. These models addresses the possibility of wide spread subvertical felsic dike swarms throughout the eastern United States that could account for the New York - Alabama Magnetic Lineament. Although the alignment of enough dike swarms to produce the NYAML might be questioned, these possible sources can not be dismissed.

The last three candidates for the source of the New York - Alabama Magnetic Lineament are the models shown in Figures 4.8, 4.9, and 4.10. The model in Figure 4.8, composed of a wedge and two crustal blocks, is the best candidate for satisfying both potential field observations and the reflection profile ARAL (Figure 2.2). The model includes no boundaries that are not explicitly imaged as impedance contrasts on profile ARAL. The potential field values in the wedge are higher than those of the crust to the southeast, and are interpreted as evidence for the presence of a distinct lithology. The wedge might be composed of diorites or their metamorphic equivalent, and are situated adjacent to more granitic crustal rocks to the southeast. As this model satisfies the gravity, magnetic, and reflection data well, it is considered to

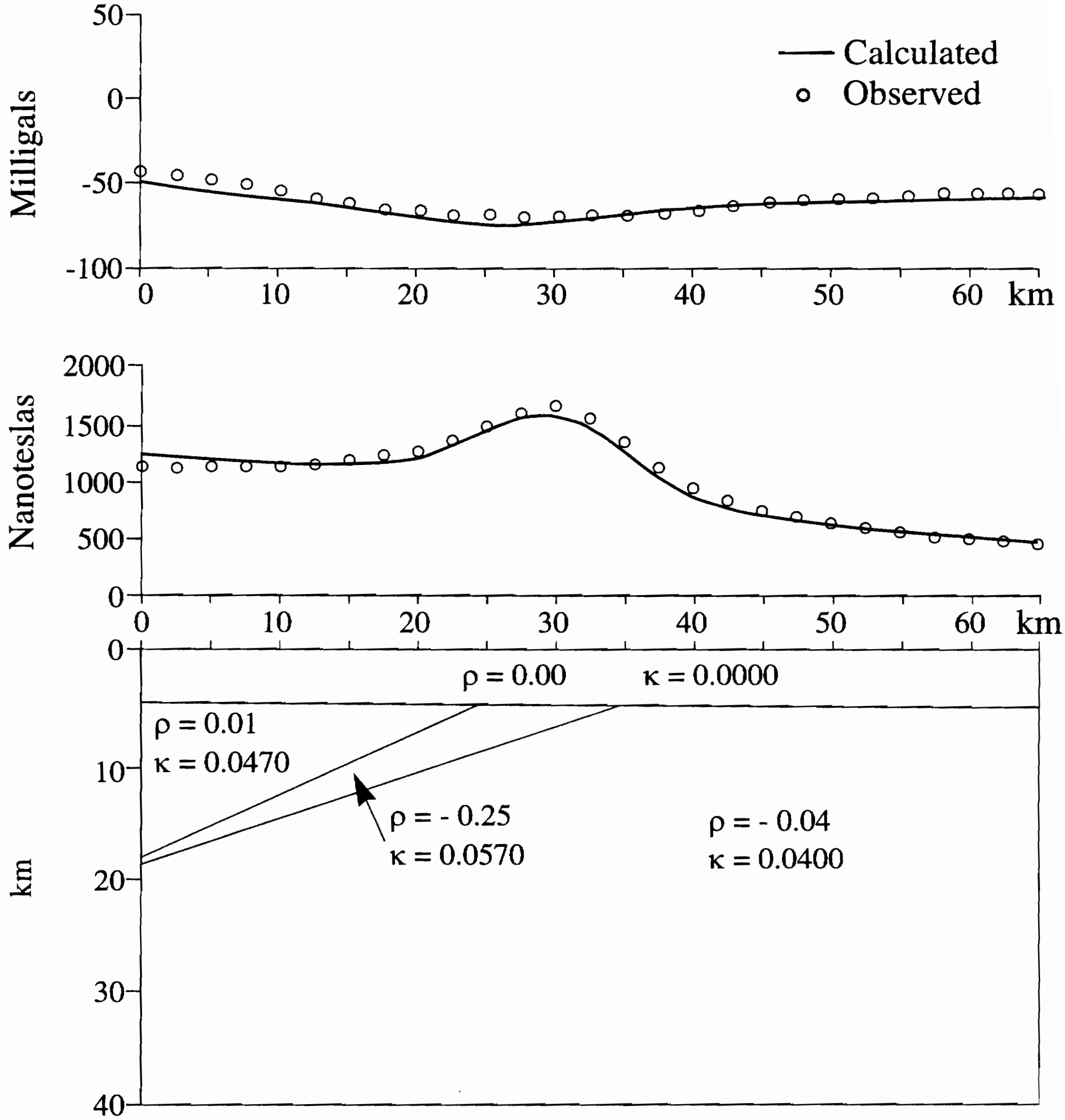


Figure 4.10. Tapered dipping layer: The low density, high susceptibility layer provides a good fit to the observed gravity and magnetic data; however, the lower contact of the layer is not imaged on the reflection data. The wedge is higher in density and susceptibility than the crust to the southeast, but lower in these properties than the dipping layer.

ρ = density difference from 2.67 gm/cc, κ = susceptibility

illustrate the source of the New York - Alabama Magnetic Lineament and the character of the adjacent crust.

The calculated values for the models displayed in Figures 4.9 and 4.10, composed of a wedge with a dipping layer, yield profiles that closely represent the observed total intensity and Bouguer gravity profiles. The placement of the lower contact was chosen on the basis of producing the best fit to the potential field data. The absence of an acoustic contrast on the reflection data in the presence of a density contrast across the lower contact between the layer and the adjacent crust is considered to seriously compromise the validity of these models.

In models 4.9 and 4.10, the density of the wedge is high enough to be interpreted as sourced by rocks of at least intermediate composition. Lithologies such as diorite could account for the higher density calculated for the wedge in the dipping layer models. The dipping layer in both models exhibits high susceptibility and low density in comparison with the wedge or adjacent crust. The layers might be interpreted as altered zones, possibly associated with intrusion or crustal scale fracturing leading to deposition of magnetically susceptible minerals while maintaining an essentially felsic composition. Alternatively, the dipping layer could be interpreted as an intrusion.

Summary

The strength of these gravity and magnetic models lies in their correlation with excellent seismic reflection data. While a wide range of geometries yield good approximations to the potential field data, only those geometries that can be interpreted from the seismic data are deemed to represent the most likely sources for the New York - Alabama Magnetic Lineament. The model pictured in Figure 4.8 is the preferred candidate because it satisfies the potential field data and is consistent with impedance contrasts on profile ARAL. In addition, the geometry and size of the wedge provides for a range of interpretations that could account for the length and linearity of the NYAML.

Chapter 5: The Eastern Tennessee Seismic Zone

The eastern Tennessee seismic zone (ETSZ) has received attention recently as representing a possible site for the occurrence of a damaging earthquake (Powell, et al., 1994; Nishenko and Bollinger, 1990). The south-eastern United States seismic network (SEUSSN) data base indicates that for an 11 year period from July 1977 through June 1988 eastern Tennessee was the most active seismic region within the reporting area (Figure 5.1). The only locus of more activity in the eastern United States is the New Madrid seismic zone along the boundary between Tennessee and Missouri. Powell et al. (1994) reported that the seismic moment release per unit crustal volume over that last 10 years indicates that the Eastern Tennessee Seismic Zone has the second highest strain release energy in the eastern U.S. They indicated that, although second to the New Madrid seismic zone, the region affected in eastern Tennessee is smaller, suggesting a more concentrated zone of activity. Bollinger et all. (1991) reported that nearly half of the felt earthquakes occurred within 25 km of Knoxville, Tennessee. Nishenko and Bollinger (1990) warned that the probability of a damaging earthquake in the eastern United States is at a moderate to high level, and that, given the lack of seismic wave attenuation for the well-indurated crystalline rock, the damage radius for a given magnitude event would be much larger than the radius expected in a region with higher attenuation such as California. They defined a damaging earthquake as one of $m_b \geq 6.0$. An event of such a magnitude in eastern Tennessee could pose serious consequences for the large population centers, hydroelectric, and nuclear power generation facilities there, motivating the desire to obtain a clear understanding of all physical relationships that might contribute to seismic hazard risk assessment.

Reflection seismic data shown in Figure 5.2 along the Eastern Tennessee Seismic Zone provide an opportunity to image structures and geometries

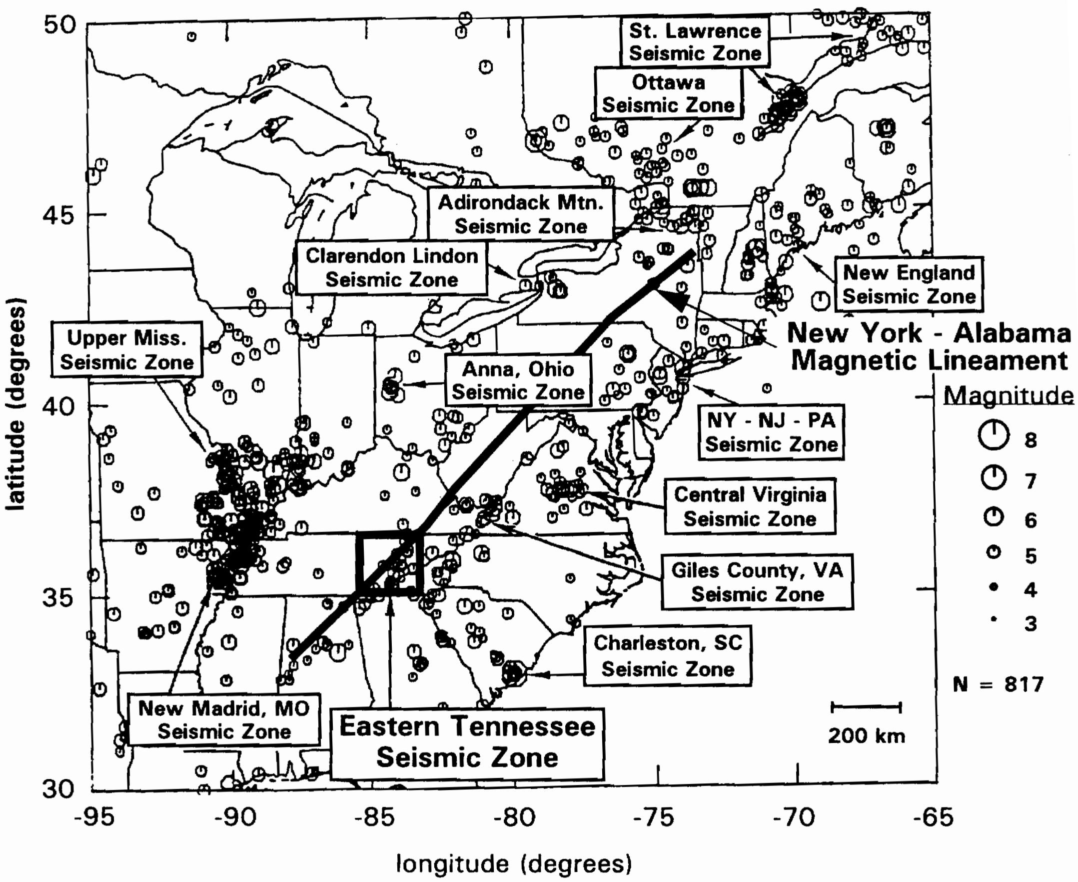


Figure 5.1. Central and eastern United States seismicity: Earthquakes of magnitudes over 3 are shown in numerous seismic zones throughout the region for 1568 to 1987. The eastern Tennessee seismic zone (ETSZ) forms a NE - SW trending band of events subparallel to the New York - Alabama Magnetic Lineament, shown as a heavy line. As the second most active seismic region in the east-central United States, the zone merits consideration as a potentially destructive locus of seismicity. The study area, shown as a dark square, includes part of the ETSZ. Seismicity map courtesy of G.A. Bollinger.

within the crystalline basement that might be important in developing a velocity model for the crust in eastern Tennessee. Geometries visible on profile ARAL (Figure 2.2) indicate that the use of a horizontally layered velocity model for regional earthquake location in the seismic zone might not be appropriate.

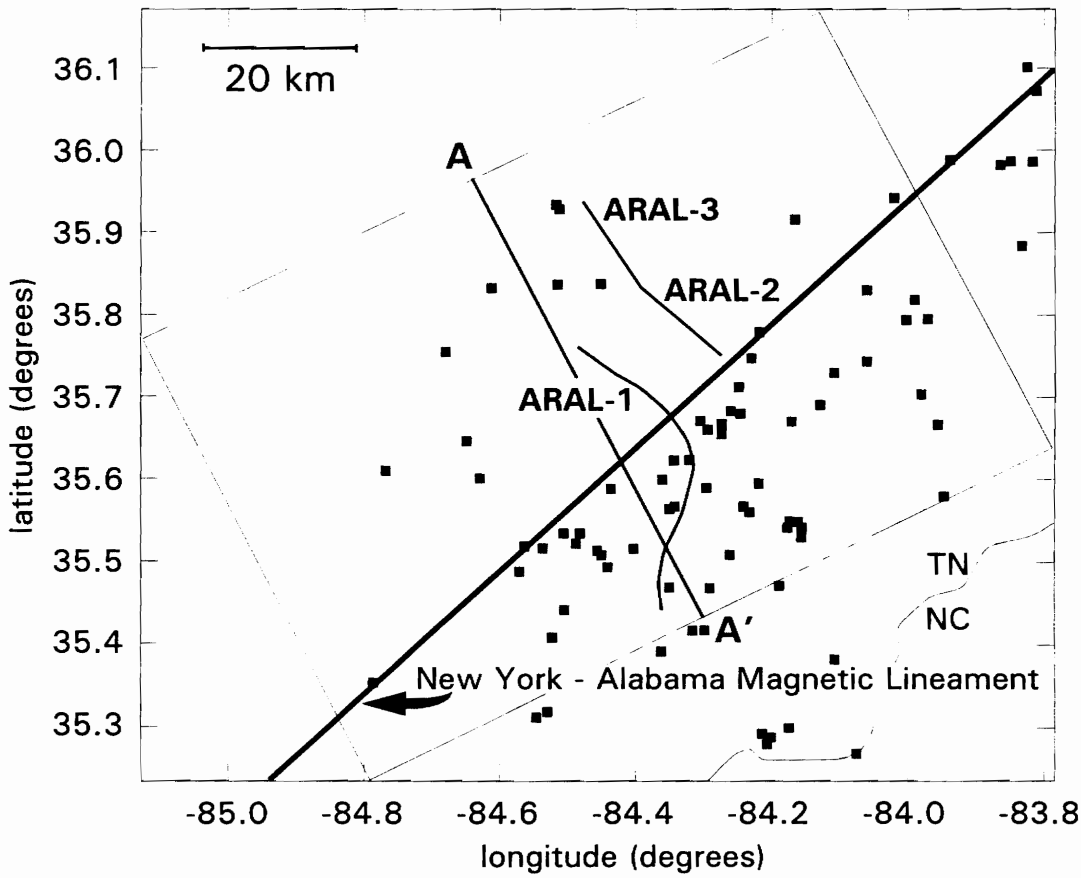


Figure 5.2. Eastern Tennessee earthquakes and reflection profile ARAL: Map view of the study area showing the locations of 62 earthquakes (squares) recorded from 1979 and 1991. The events chosen were of m_b > 2.0, had horizontal and vertical location errors of less than 5 km, and had been assigned no fixed depths. The New York - Alabama Magnetic Lineament is shown as a heavy line trending through the region at N45°E. The shaded area delineates the region from which earthquakes were projected onto line A-A′ representing the reflection profile ARAL. The earthquakes were projected perpendicular to the trend of A - A′.

The proximity and alignment of the New York-Alabama Magnetic Lineament with the eastern Tennessee seismic zone suggest that these two prominent geophysical phenomena are related. King and Zietz (1978) proposed that the source of the New York-Alabama Magnetic Lineament might be a mega-scale, strike-slip boundary within crystalline basement, and Johnston et al. (1985) postulated that the lineament represents the western boundary of a seismogenic crustal block.

Figure 5.3 shows the approximate positions of events projected onto profile ARAL (Figure 2.2) from 50 km northeast and southwest from within the shaded region in Figure 5.2. The events plotted are those recorded by the southeastern United States seismic network from 1979 to 1991 and consist of events with $m_b > 2.0$. All locations with vertical or horizontal errors over 5.0 km were eliminated from the catalog used here, as were all events to which fixed depths had been assigned. The foci of eastern Tennessee earthquakes range from 3 km to 29 km deep with most of the locations beneath the shelf strata within the crystalline basement (Bollinger et al., 1985). The majority of events plot within the region of the crust dominated by strong west-dipping reflections on profile ARAL. The earthquake foci are not shown on the reflection profile because the proximity of foci to reflections is coincidental. The errors associated with the locations of the earthquakes used in this analysis are too large to assign an event to a position on a reflection.

The available earthquake focal mechanism solutions for the Eastern Tennessee Seismic Zone (Teague et al., 1986) and regions to the northeast suggest that strike slip motion is the dominant strain release mechanism at the present time. The nodal planes determined by Teague et al. strike N-S (right lateral) and E-W (left lateral). Munsey and Bollinger (1985) found a similar sense of motion for the earthquakes in the Giles County seismic zone 400 km to the northeast in Virginia. The nodal planes in that study were oriented NNE and ESE. The azimuth of maximum horizontal compressional stress in

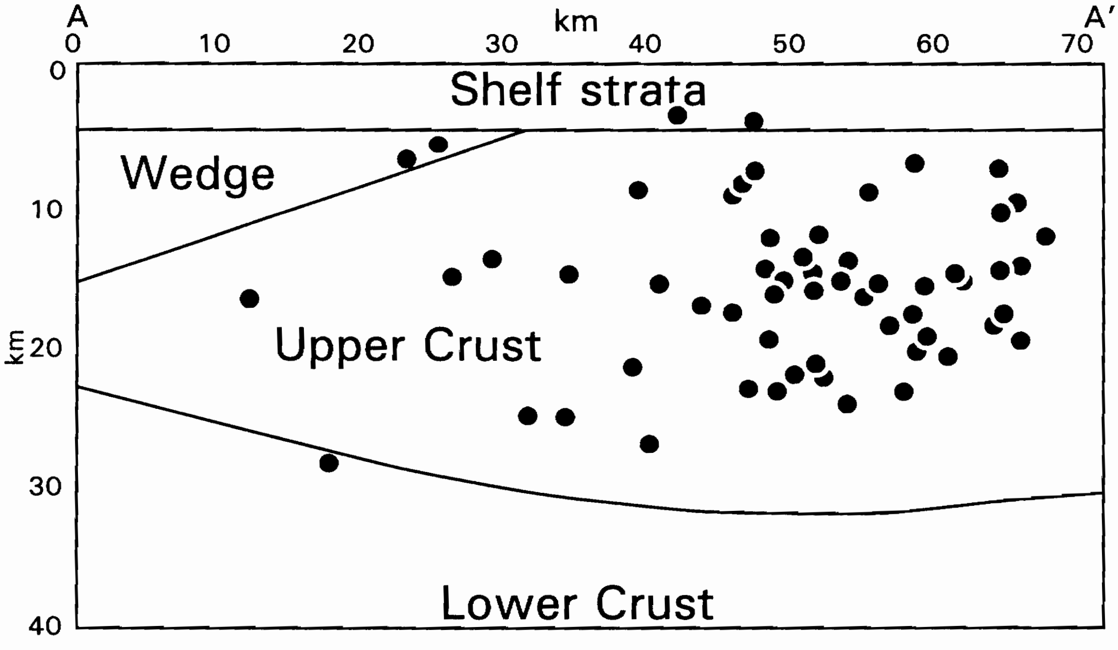


Figure 5.3. Earthquakes on profile ARAL: The relationship of foci to individual reflections can not be inferred as the projection distances are too great, but the reflection character of the seismogenic crust can be distinguished. Most of the events occurred below the shelf strata within the part of the crust characterized by bright west-dipping reflections.

the eastern United States has been interpreted to trend NE (Zoback and Zoback, 1989) and locally to trend N50°E. This orientation is nearly parallel to the trend of the New York-Alabama Magnetic Lineament.

Although stress-drop calculations are not available to evaluate whether the earthquakes are occurring along new fractures or reactivating older features, the possibility that they occur on older fractures is reasonable. The presence of subvertical fractures provides the necessary orientation for reactivation in a strike slip tectonic environment, and provides conduits for the introduction of pore fluids. Fluids and pore pressure transients along fractures have been interpreted to trigger earthquakes in regions loaded to a critical state (e.g. Costain et al., 1987a and 1987b). The uplift of the crust to present depths suggests vertical movement of at least 20 km and could account for the introduction of vertical fractures as confining pressures changed, although fractures oriented subvertically can not be imaged on these reflection data at the depths of the foci. The continuity of the west-dipping reflections does not argue for substantial displacement along subvertical failure planes, but localized movement is not precluded.

The concentration of seismogenic activity along the southeastern side of the New York-Alabama Magnetic Lineament merits discussion. Two-dimensional gravity and magnetic modeling (this paper) are interpreted as evidence that the NYAML and its associated gravity low result from a magnetic susceptibility and density contrast across a northeast-dipping boundary. The dip of the contact is approximately 30°. The surface locations of the earthquakes in the ETSZ fall to the southeast of the mapped position of the NYAML; however, the mapped position of the NYAML does not mark a vertical boundary between the contrasting crustal types. Hence, the alignment of the foci with the mapped position of the NYAML does not bear a relationship to the source of the NYAML unless the foci fall throughout one of the blocks responsible for the gradient. It is possible to suggest, based on the projection shown in Figure 5.3, that foci fall throughout the crust typified by west-dipping reflections on both sides of the NYAML. Until better hypocenter locations are determined for the earthquakes in eastern Tennessee, the relationship between the foci and the NYAML will be unclear.

Earthquake activity does not extend along the length of the NYAML, nor does it occur wherever west-dipping reflections have been reported in the crust (e.g. Pratt et al., 1989). Apparently, the seismicity in eastern Tennessee is due to a combination of factors, some of which are independent of the reflection character of the crust and of the NYAML.

The delineation of crustal blocks in which more seismic activity is focused might suggest diagnostic characteristics of the crust in eastern Tennessee that make the crust more seismogenic in the presence of triggering mechanisms. Such characterization can be done by using reflection data. The use of layer boundary information on reflection data will aid in the development of better velocity models for the region and provide for better earthquake locations. In turn, better locations will enable researchers to ascertain whether seismogenic crust can be characterized using reflection or potential field data.

Chapter 6: Discussion

The integration of reflection seismic data with potential field and earthquake data provides an opportunity to characterize the crust beneath the thin-skinned thrust sheets of the Valley and Ridge province in eastern Tennessee. Salient features observed in the data and revealed by modeling include:

- Moho reflections on profiles FM and on ARDU at 14 seconds suggest crustal thickness of about 45 km.
- East-dipping reflections on profile ARAL below 9 seconds.
- Band of sub-horizontal higher reflectivity in the middle crust on most reflection profiles.
- Wedge-shaped zone of distinct reflectivity on northwest half of reflection profile ARAL.
- Strong west-dipping reflections appear on the southeast half of seismic reflection profiles ARAL and ARDU to middle crustal depths.
- Magnetic signatures west of the New York - Alabama Magnetic Lineament (NYAML) are produced by numerous small bodies of limited areal extent.
- Length and linearity of the NYAML permits 2-D potential field modeling and provides out-of-the-plane reflection data control.
- The steep gradient of the NYAML can be modeled as sourced by the magnetic susceptibility contrast between the wedge-shaped body and the adjacent crust characterized by west-dipping reflections.
- Earthquake locations in the eastern Tennessee seismic zone trend parallel to and fall mainly to the southeast of the NYAML.
- Earthquakes do not extend along the entire length of the NYAML.
- Earthquake foci in the eastern Tennessee seismic zone plot mostly within the crystalline basement typified by west-dipping reflections.

Crustal thickness

The considerable thickness of the crust in eastern Tennessee can be interpreted from reversed refraction and reflection seismic data and from gravity modeling. The apparent preservation of thick crust through extensional and compressional tectonic events following the Grenville orogeny further suggests that the younger events did not significantly affect the basement in this region with the possible exception of the emplacement of dike swarms. The absence of faults over a few tens of meters of throw along the bright reflection at the base of the shelf strata (about 1.4 seconds, Figures 2.2 to 2.8) supports the hypothesis that large scale crystalline basement deformation had ceased prior to deposition of the late Precambrian shelf strata. Because the last orogeny to dominate eastern Tennessee culminated with the Grenville event, that orogeny or a previous event is probably responsible for the production of the thick crust.

The interpretation of the Moho at about 14 seconds on seismic reflection profiles ARDU and FM provides a basis for crustal thickness estimates. The depth estimate of 45 km is consistent with the work of others (e.g. Tatel et al., 1953; James et al., 1968; Prodehl et al., 1984) when higher crustal velocities are used. Applying a higher average velocity of 6.5 km/sec two-way travel-time to these data results in the location of the reflections attributed to the Moho at about 45 km.

Because the crystalline basement rocks had been eroded to sea level by the end of the Precambrian, the present crustal thickness represents a minimum. Removal of the imbricated Alleghanian thrust sheets from the total thickness estimate of 45 km results in a thickness of about 40 km for the part of the crust composed of crystalline rocks. Although the metamorphic grade of the basement in eastern Tennessee is speculative, the nearest exposures of Grenville basement are of upper amphibolite to granulite grade gneisses, and are probably good analogues for the grade in the study area.

The presence of high grade rocks in Blue Ridge thrust sheets to the southeast of the study area indicates that they were close enough to the surface to be involved in Alleghanian thrusting. Hatcher et al. (1989) suggested that the Blue Ridge master detachment formed along the brittle-ductile transition within Grenville basement. The Blue Ridge thrust sheet moved westward at that depth until it ramped upward, bringing the high grade metamorphic rocks to the surface east of the Tennessee border.

Green et al. (1988) suggested that the crustal thickness in the vicinity of the Grenville Front Tectonic Zone was originally 70 km on the basis of the present exposure of amphibolite - granulite grade metamorphic rocks and the ductile deformation seen in outcrop. The required burial depth for deformation and metamorphism was approximately 20 km. If the rocks in the study area have similar metamorphic grades, the surface of the basement beneath the shelf strata must have been buried to at least 20 km. Hence, the depth to the base of the crust in eastern Tennessee, presently at about 45 km, to have been at least 20 km deeper, requiring the total crustal thickness to have been at least 55 to 60 km. Such thicknesses are reported in collisional orogenic belts like the Himalayas.

Grenville Front and east-dipping reflections

The east-dipping reflections described on the northwestern end of Profile ARAL below 7 seconds resemble the reflections imaged by Milkereit et al. (1992) on industry data and by Green et al. (1988) on GLIMPCE data (Figure 1.7). Green et al. related the bright, east-dipping Grenville front reflections to ductile faults formed in high pressure and high temperature conditions during collision. Kinematic indicators in the mylonites record northwest-directed emplacement of the rocks along the Zone. Green et al. noted that the zone of deformation along the GFTZ increases in width from 32 km at the surface to about 50 km at 9 seconds due to the fanning of reflections with depth. Reflection dips range from 35^{o} near the near the surface to 25^{o} at depth below the eastern margin of the Zone.

COCORP seismic data in Ohio (Figure 1.10) reveal a zone of east-dipping reflections that extends from about 3 sec (9 km) to the bottom of Line OH-1 at 5 sec (15 km), which Pratt et al. (1989) attributed to the presence of the Grenville Front. On the COCORP data the reflections dip 25^o to 30^o (Culotta et al., 1990) and correlate with a high-frequency magnetic pattern. Culotta et al. used the magnetic pattern to extend the Grenville front into Tennessee where it falls 70 km west of the western end of Profile ARAL. This location is consistent with the extrapolation of the position marked in Kentucky by a cooperative study among the Kentucky, Ohio, and Indiana Geological Surveys (Drahovzal, et al., 1992), which positions the front west of the East-continent Gravity High on the Kentucky-Tennessee border.

Figure 6.1 indicates the relationship of the east-dipping reflections to the Grenville Front and the Grenville Front Tectonic Zone. If the east-dipping reflections on profile ARAL record the position of the eastern margin of the Grenville Front Tectonic Zone, the location and dip of the reflections on profile ARAL require that either (1) the margin falls at the contact with the shelf strata about 60 km northwest of the end of profile ARAL and dips about 20^o east dip to appear below 7 sec on the profile, or that (2) the margin is located approximately 40 km northwest of the profile and maintains a dip between 25^o and 30^o. Assuming the Grenville Front Tectonic Zone maintains the 32 km width that is in evidence at the surface in Canada, then the position of the Grenville Front lies between 60 and 90 km from the northeastern end of profile ARAL. As the published location of the Front falls at 80 km from the northeastern end of the line, the east-dipping reflections can be interpreted reliably as evidence of the Grenville Front Tectonic Zone beneath eastern Tennessee.

Mid-crustal band of high reflectivity

Nelson et al. (1992) described a gently dipping zone of reflectivity at

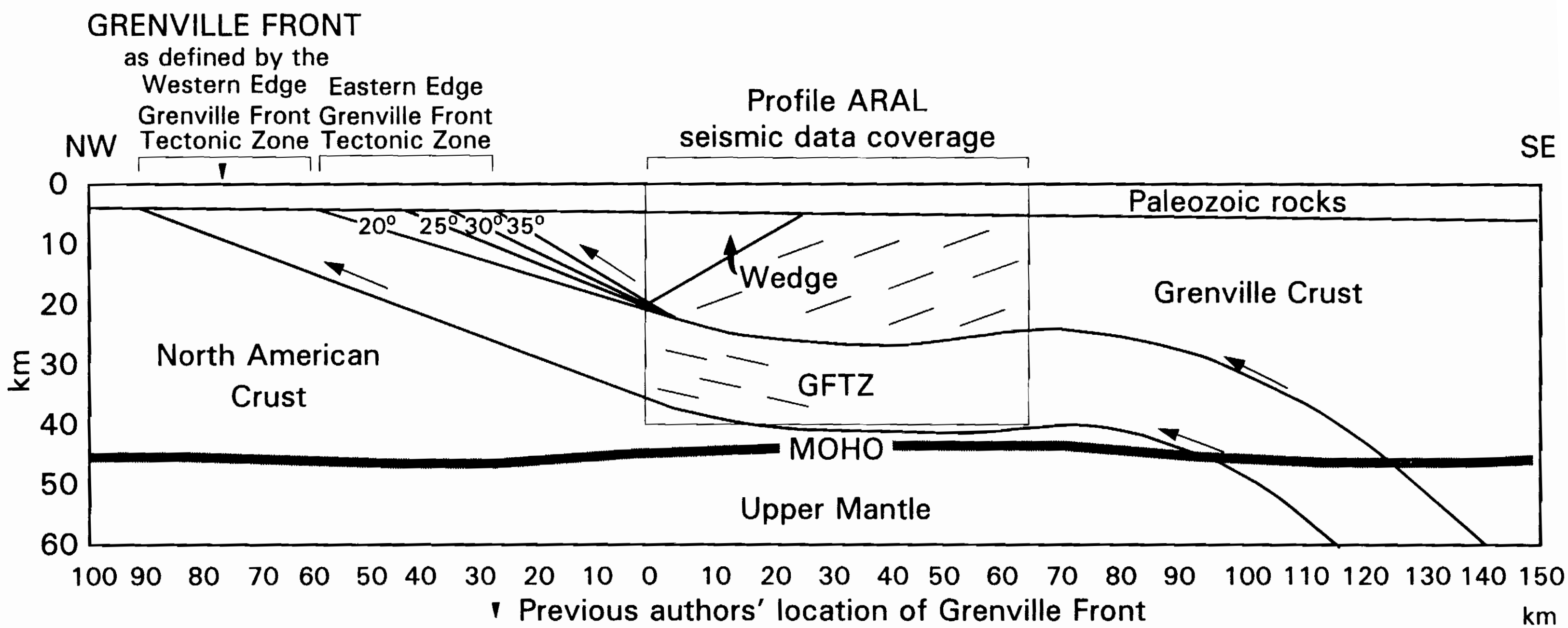


Figure 6.1. East-dipping reflections: Reflections with east dip are visible beneath the band of higher reflectivity in the middle crust. Extending the reflections to the surface for various dip angles suggests a range of possibilities for the position of the boundary between the reflective band and the region beneath it dominated by east dip. In addition, the east dip might correlate with the Grenville Front Tectonic Zone (GFTZ) on the GLIMPCE data of Green et al. (1988) (Figure 1.7). If so, the Grenville Front should be located between 60 and 90 km from the northeastern end of the seismic profile. The arrow shows the location of the Grenville Front as marked by previous authors on the basis of potential field signatures and well data. The fit between the positions is remarkable and encourages the interpretation of the east-dipping reflections as evidence for the presence of the Grenville Front Tectonic Zone, and suggests that the reflection data record the suture of the Grenville terrane with North America.

10 to 11.5 seconds (30 to 35 km) on preliminary INDEPTH (INternational DEep Profiling of Tibet and the Himalaya) cmp data that extends across the length of the 100 km line, and tentatively interpreted this feature to be the detachment between the Asian and Indian plates beneath southern Tibet. The band of higher reflectivity within the middle crust at about 9 seconds (approximately 25 to 30 km depth) on the eastern Tennessee seismic data has a similar seismic character, and could represent a detachment between the North American continental block and the Grenville terrane. To be consistent with the interpretation that the east-dipping reflections beneath the band have a relationship to the GFTZ, the band is interpreted as the top of a wide zone of deformation that characterizes eastern margin of the GFTZ. Potential field modeling does not suggest that the band separates crustal layers with substantially different densities and permits the interpretation of subhorizontal convergence of felsic continental blocks. The implication in this scenario is that the Granite-Rhyolite terrane extends eastward beneath the Grenville basement to an indeterminate distance below the base of reflection profile ARAL (Figure 2.2). This model for convergence suggests that continental suturing did not proceed in a subvertical orientation and involved substantial obduction of the Grenville crust.

Alternatively, the band might represent a compositional change in the middle crust. The depth of the band corresponds to the deeper velocity transition zone of Prodehl et al. (1984) where they reported a gradual increase in interval velocities from 6.8 km/sec to 7.1 km/sec over a depth of about 6 km ($\sim$1.7 sec). A similar transition is described by Holbrook et al. (1991) in Nevada. Holbrook et al. attributed a mid-crustal velocity discontinuity to the boundary between more silicic and more mafic crust. In addition, they ascribed the high reflectivity of the lower crust and upper mantle to the presence of ductile shearing and mantle-derived igneous material. Their model assumes considerable extension of continental crust, however, and might not be appropriate for eastern Tennessee. Furthermore, the necessity imposed by the gravity data that the crust in eastern Tennessee be essentially felsic throughout its thickness is problematic for their model, which involves the emplacement of a considerable volume of mafic material. Such a volume is not substantiated by the gravity data, which indicate the presence of thick, felsic crust.

The Wedge

The image and interpretation of the wedge-shaped body visible on the northwestern end of reflection profile ARAL is a major contribution of this study. The body can be distinguished from the strongly reflective crust to the southeast by its relatively low reflectivity and faint subhorizontal reflections. The contact between the wedge and the southeastern crust appears to dip at approximately 20° and is subparallel to the west-dipping reflections visible in the southeastern crust. That the wedge underwent uplift and erosion along with the adjacent basement is clear from the continuity of the reflections at the base of the Cumberland Plateau and Valley and Ridge provinces. Deposition of rift and drift strata occurred subhorizontally on top of both regions of the crust and indicates that uplift had ceased by late Precambrian time.

While the strike of contacts and units within the crystalline basement can not be determined with certainty, the linear extent of the New York-Alabama Magnetic Lineament suggests that crustal susceptibility changes trend approximately parallel to the strike of structures within the Valley and Ridge allochthon. Such evidence is critical in extending interpretations of the two dimensional reflection seismic profile away from the plane of the data and for justifying the use of two-dimensional potential field models to suggest possible density and susceptibility properties of bodies imaged on the seismic data.

The potential field modeling indicates that good fits to the recorded gravity and magnetic data can be obtained by assigning the wedge density and susceptibility values that are higher than those of the crust to the southeast, with a susceptibility value of 0.0520 for the wedge and 0.0450 for the southeastern crust characterized by west-dipping reflections. The density of the wedge models as 2.61 gm/cc, and the southeastern crust models as 2.56 gm/cc. Although these values are not strikingly different, they are suggestive of different compositions for the sections of the crust visible on profile ARAL.

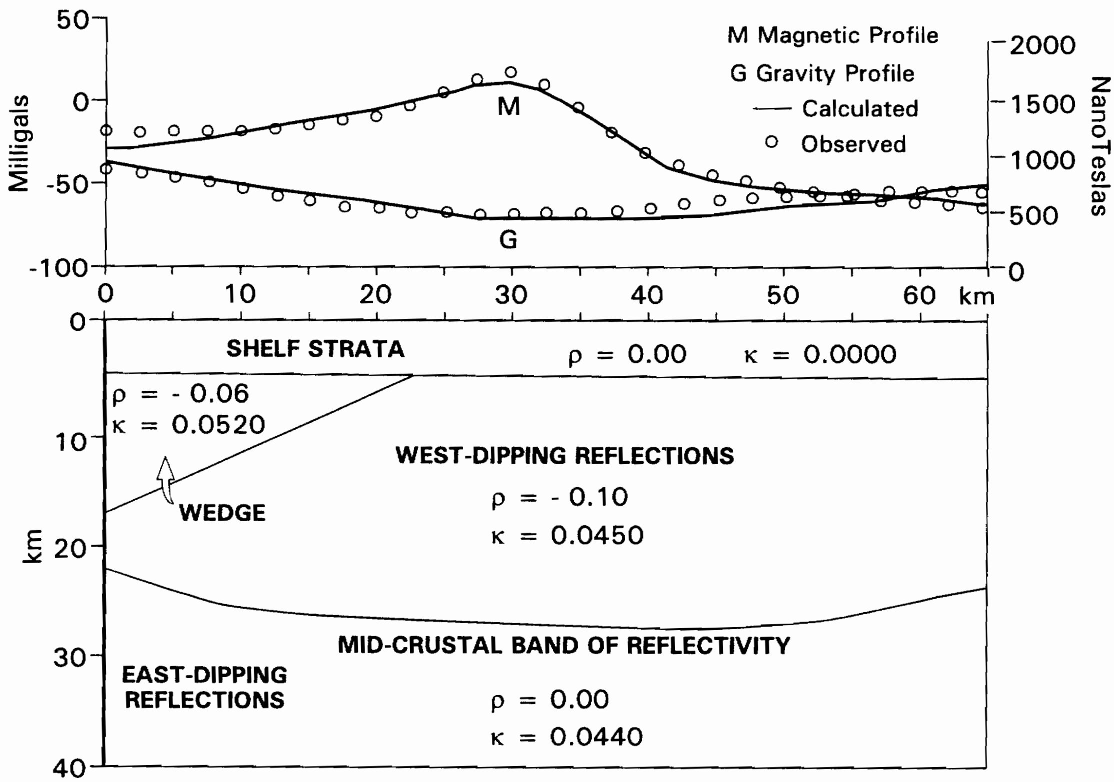


Figure 6.2. Two-dimensional model of profile ARAL: The gravity and magnetic values calculated for the model represent the best fit to the observed values, while the model is the best representation of the bodies on the reflection profile. The density, ρ, and susceptibility, κ, for the wedge are higher than those for the crust to the southeast, and are comparable to the values in the lower crust. The higher values are interpreted to be evidence of a compositional change with an associated difference in acoustic impedance across the boundary between the wedge and the crust to the southeast.

The absence of pervasive west-dipping reflections within the wedge is compelling evidence that the wedge was not affected by events that produced the reflections; however, the original dip angle at which the contact and the west-dipping reflections formed is not constrained. The possibility that these features have been rotated since their origins suggests additional age and emplacement relationships between the wedge and the southeastern crust. Relative to the southeastern crust, the wedge could be a younger intrusive body, a tectonically emplaced block with no clear relationship to the adjacent crust (e.g. a distinct terrane), a younger depositional unit, or an older, less layered unit.

The presence of the New York-Alabama Magnetic Lineament suggests that lithologies exist within the crust with distinctive susceptibilities, and the continuity of the gradient argues for preservation of relationships between crustal blocks over large distances to the northeast and southwest. Furthermore, the east-dipping reflections beneath the middle crustal band of higher reflectivity can be compared to the reflections in the Grenville Front Tectonic Zone in Canada. Therefore, the following models assume that the contact and west-dipping reflections remain in approximately the same orientations at which they formed. The models include casting the wedge as igneous in origin, as a tectonically emplaced crustal block, and as a basin.

The wedge as an anatectic melt

The wedge might be the result of anatectic melting following crustal thickening associated with continent-continent collision as shown in Figure 6.3. The Grenville event is widely accepted to be a major orogenic event with possible doubling of continental crust to 70 km (e.g. Wynne-Edwards, 1972); Dewey and Burke, 1973). The associated thermal event would result in regional anatexis along the axis of the thickest part of the crust. Assuming that the present day crustal thickness is a relic of Himalayan - scale tectonism produced during the Grenville event, the thermal regime along that axis should have produced considerable melting. The New York - Alabama Magnetic Lineament is sub-parallel to the maximum crustal thickness axis in eastern Tennessee reported by Taylor (1989) and James, et al. (1968), and could

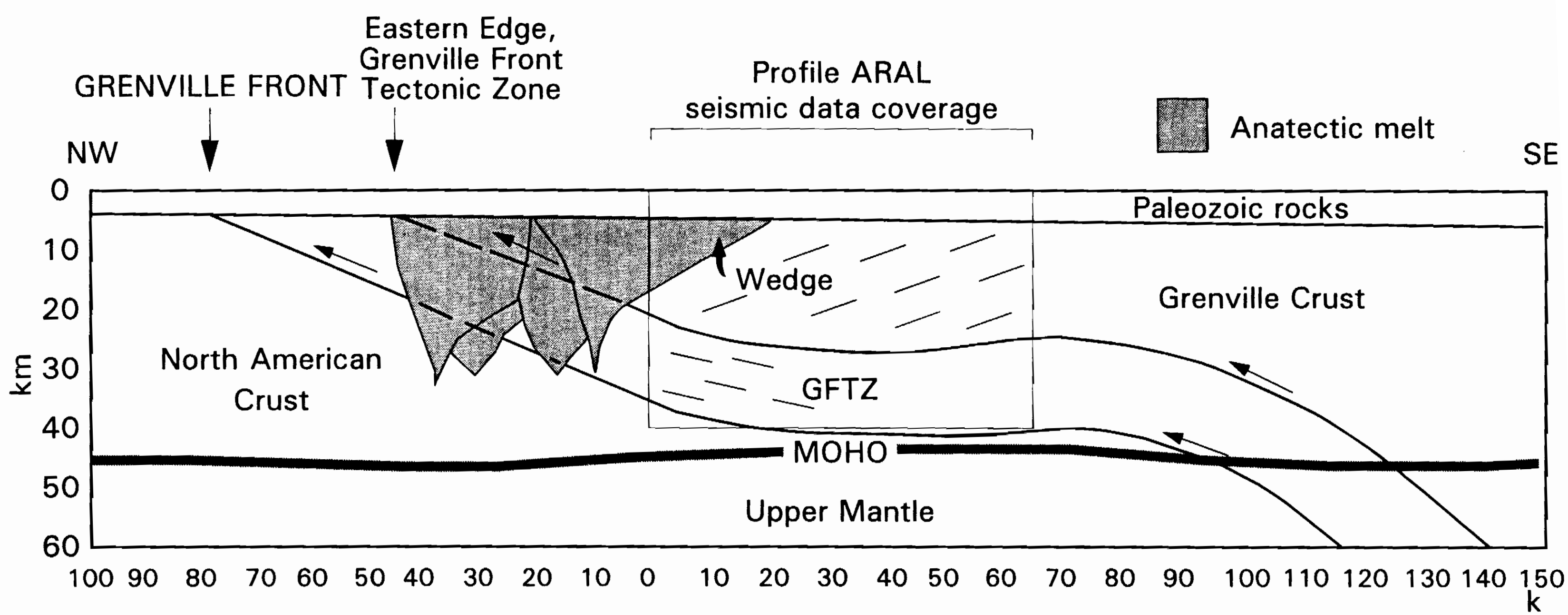


Figure 6.3. Anatectic melts: Anatectic melting in response to a thickened crust might have produced the wedge, shown as shaded masses. As younger bodies than the collision event, they would not contain evidence of deformation interpreted to produce the west-dipping reflections. Intrusion along west-dipping layers could explain the similarity between the dip of the wedge contact and the dip of the reflections in the crust to the southeast. The position of the GFTZ is based on the presence of east-dipping reflections visible below the wedge on ARAL. The location of the Grenville Front is based on the location proposed by previous authors and on the interpreted width of the GFTZ of about 30 km.

indicate the presence of such Precambrian intrusions along it's length.

Winkler (1979) indicated that anatexis can account for the production of substantial volumes of granites, granodiorites, and lesser amounts of tonalites and trondhjemites from precursors of paragneisses and quartzo-feldspathic mica schists. Parent rocks of these compositions can be expected along a continental margin, where pelitic rocks accumulate on the continental shelf. Water is necessary to generate quantities of melt and might be entrained along fractures in highly tectonized regions such as the Grenville Front Tectonic Zone.

If anatectic melting is responsible for the wedge, and thus for the NYAML as proposed here, then the linearity of the melt complex can be explained as occurring along the maximum thickness axis of the Grenville orogen and at positions where sufficient water was available in the rocks to favor the generation of large melt volumes.

The wedge, if responsible for the lineament, can be inferred to post-date the collision on the basis of the absence of high reflectivity. The presence of west-dipping reflections west of the Lineament in Ohio as well as southeast of the Lineament in Tennessee supports the interpretation that the wedge was emplaced into older, deformed crust. Hence, the NYAML can be interpreted to represent the axis of anatectic melting following collision.

The New York - Alabama Magnetic Lineament diverges from the location of the Grenville front north of the study area. Figure 6.4 diagrams the relationships among magnetic lineaments and the Grenville province, shows the interpreted position of the wedge along the NYAML, and indicates the positions of west-dipping reflections reported in Ohio and in the study area. The location of the NYAML proposed by King and Zietz (1978) places it about 650 km east of the Grenville Front along the eastern boundary of the Adirondack Mountains. The Amish anomaly location proposed by Culotta et al. (1990) places it 400 km east of the front on the western side of the Adiron-

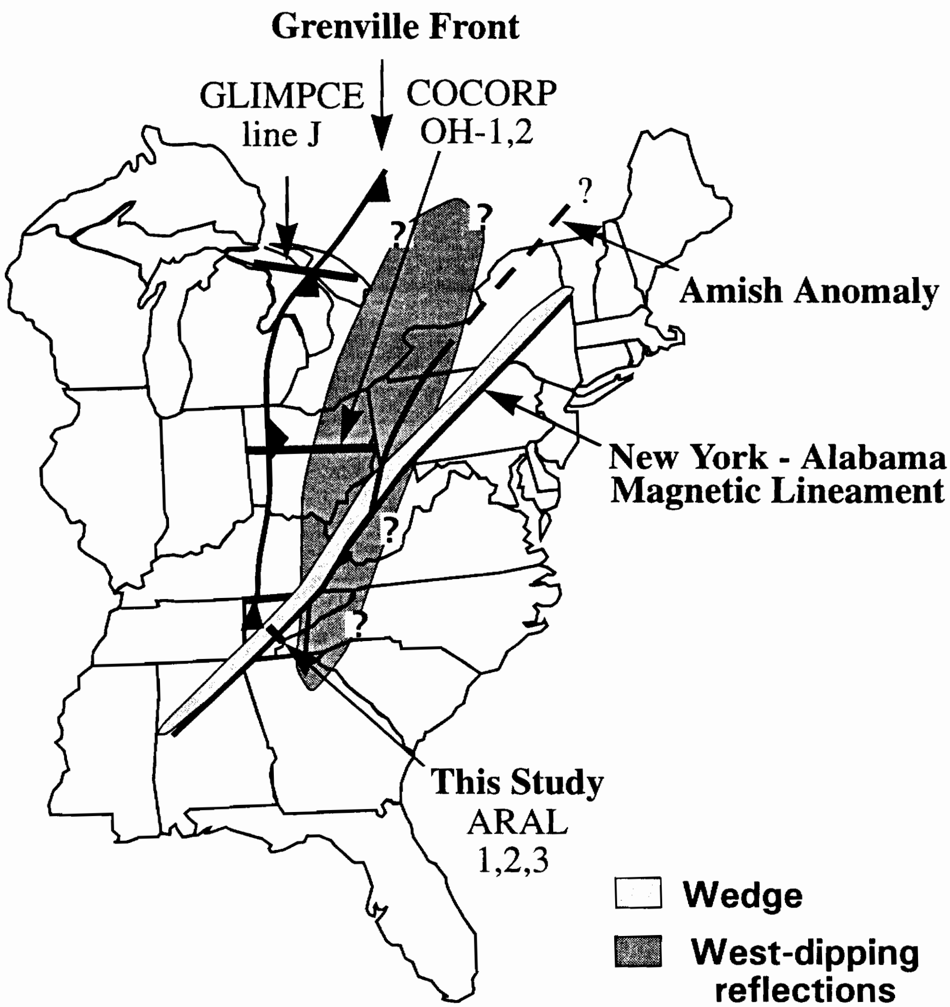


Figure 6.4. Crustal and geophysical features: The wedge, interpreted to result from anatectic melting, is shown extending along the NYAML, approximately parallel to the axis of maximum crustal thickness proposed by Taylor (1989) (Figure 1.9). The thick crust is interpreted to be related to Grenville orogenesis. The west-dipping reflections southeast of the wedge in eastern Tennessee are interpreted to be older than the wedge and might be related to those reported in Ohio by Pratt et al. (1989) (Figure 1.10). The distance between the Grenville Front and the thickest part of the crust increases northward and might indicate either a widening collisional zone from Tennessee to Canada or result from more extensive erosion to deeper crustal levels in Tennessee. The former proposal is more likely as the crustal thickness remains greater in Tennessee than in the northern U.S. and Canada.

dack Mountains. Assuming that the NYAML, which falls along the axis of maximum crustal thickness, is produced by a contrast between the wedge and the adjacent crust, and that the wedge is the product of anatectic melting along the axis of maximum thickness of the Grenville orogen, then the thickest part of the orogen formed farther away from the Grenville Front in Canada than it did in the southern United States.

A major limitation of this model is the availability of water to permit the formation of large volumes of melt at high grade conditions in the crust. The discovery of flowing water by Kozlovsky (1982) at 11 km in the deep Kola wells might help to reduce this objection.

The wedge as an intrusion

The orientation of the contact between the wedge and the crust to the southeast is approximately parallel to the west-dipping reflections, and might indicate that intrusions comprising the wedge were emplaced along preexisting boundaries in the southeastern crust. Intrusive complexes of Grenville age are reported along the eastern edge of the Valley and Ridge Province in Virginia within the Blue Ridge Province by Sinha and Bartholomew (1984). They describe the Pedlar River Charnockite Suite as intrusive into the Lady's Slipper Granulite Gneiss and support the interpretation with U/Pb age dates of 1075 m.y. and 1130 m.y., respectively, for the rocks. As potential field data required the crystalline basement to be essentially felsic throughout, the quartz-bearing charnockite and granulite gneisses exposed in Virginia are likely compositions for the wedge in eastern Tennessee. If the density and magnetic susceptibility contrast between the wedge and the adjacent crust are responsible for the New York-Alabama Magnetic Lineament, the intrusive complex is required to be batholithic in scale.

The wedge might have developed as part of a volcanic arc complex formed on the Grenville margin in response to the subduction of the North American plate as shown in Figure 6.5. The modest increase in density indi-

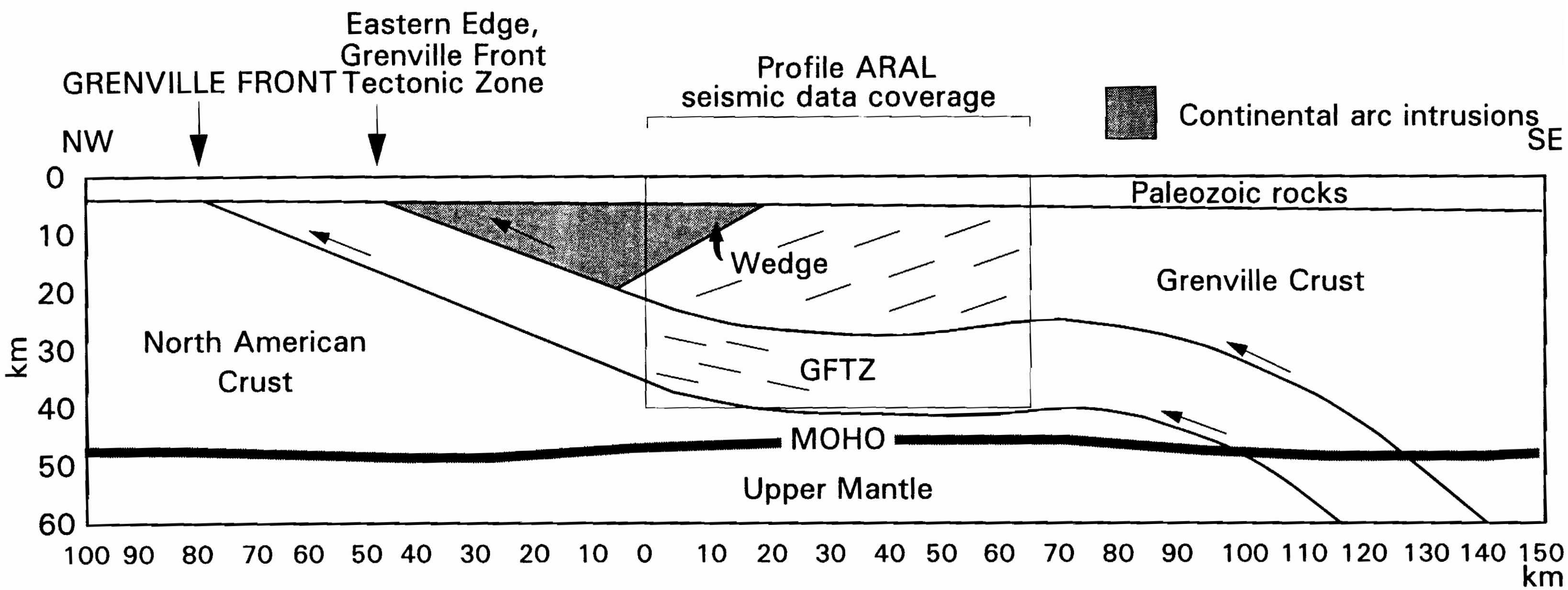


Figure 6.5. Continental arc intrusions: Formation of an arc complex on the Grenville margin might have occurred in response to the subduction of oceanic crust on the leading edge of the North American plate. The arc complex would have been caught up in collision between the continental blocks. The difference in susceptibility and density between the arc complex and the adjacent crust on both sides might be explained with this model; however, the absence of west-dipping reflections within the wedge suggests that they are older features. The preservation of layering that might be imaged by the west-dipping reflections through a Himalayan scale orogeny is more problematic than relating the reflections to structures produced during orogenesis.

cated in the gravity model suggests that the wedge can be modeled as a lithology of more intermediate composition than the Grenville crust and permits the interpretation that it originated as a continental arc.

A limitation imposed on the wedge as an arc complex emplaced prior to continent-continent collision is the lack of west-dipping reflections within the wedge. If the collisional event can be used to explain the west-dipping reflections as responding to the presence of a pervasive, crustal scale fabric, then the deformation features responsible for the reflections should extend into the wedge if the wedge is composed of rocks emplaced before collision. It is possible that the lithology of the wedge differs from that of the crust with the west-dipping reflections such that the impedance contrasts are not sufficient in the wedge to produce reflections even though the west-dipping fabric is present.

Alternatively, the west-dipping fabric might predate the formation of the wedge and the collision event. Arguments against the last possibility are the same as those for the unlikelihood of preserving original compositional layering. The probability of preservation of such features through a crustal doubling event with accompanying metamorphism to at least amphibolite grade is low; however, the possibility exists that the metamorphic grade is lower in eastern Tennessee. If so, the layering is more likely to be preserved.

The wedge as a tectonically emplaced block

If the wedge is viewed as a discrete block, it might have been emplaced as a disparate terrane on the margin of either North America or the Grenville plate before suturing between North America and the Grenville terrane (Figure 6.6). Such a terrane might have been caught up in a zone of strike slip or dip slip deformation and emplaced between the colliding continental blocks. The strongest evidence for interpretation of the wedge as a tectonically emplaced block is the extent and linearity of the New York-Alabama Magnetic Lineament.

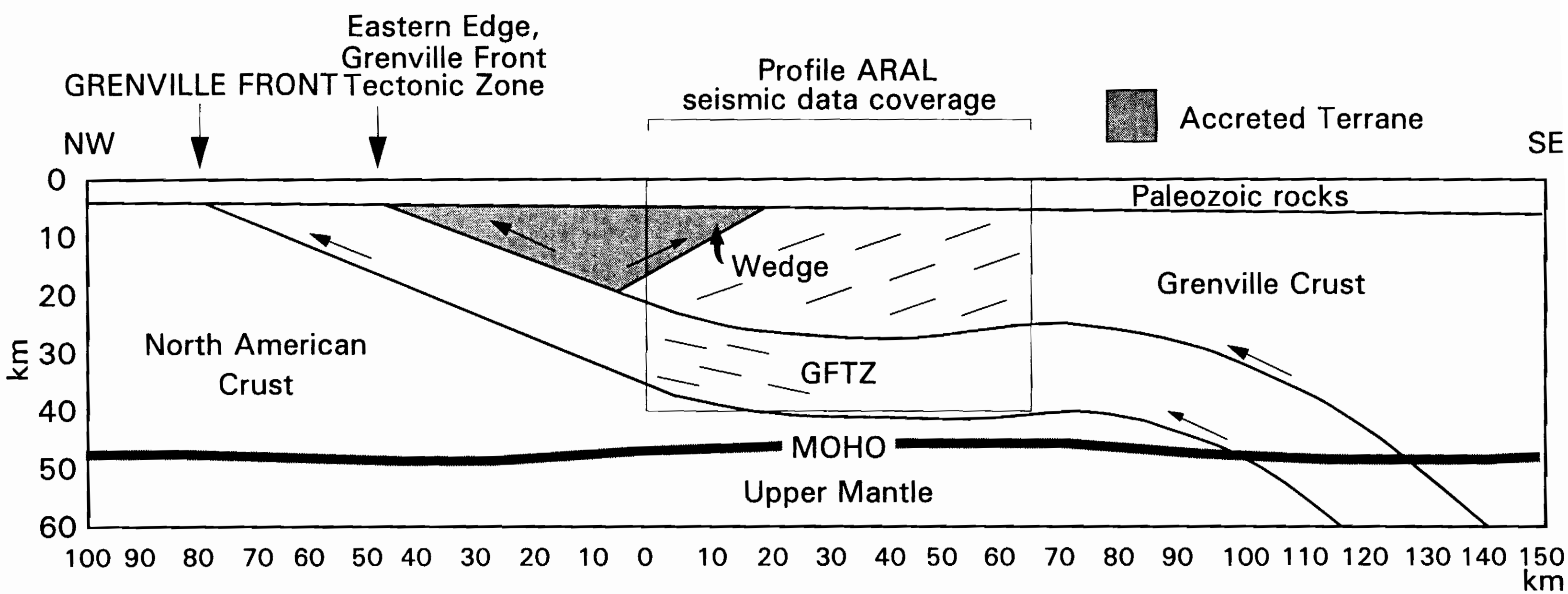


Figure 6.6. Accreted terrane: The accretion of a disparate terrane to the Grenville margin before the Grenville terrane collided with North America can explain the length and linearity of the New York - Alabama Magnetic Lineament. The contact between the wedge and the southeastern crust can be interpreted as the suture between the terranes. Like the arc complex, the terrane would have been caught up in collision between the continental blocks and should show evidence of that deformation. An alternative hypothesis is to model the wedge as an escaping block, where the accreted terrane is squeezed upward on both boundary faults. In this scenario, the block comprising the wedge might have decoupled from the surrounding basement and avoided development of the west-dipping structures thought to be responsible for the reflections in the southeastern crust.

Strike-slip mechanisms have been suggested to account for the New York-Alabama Magnetic Lineament (e.g. King and Zietz, 1978), and provide an explanation for the length and linearity of the anomaly. The San Andreas Fault in California is an example of the juxtaposition of crustal blocks along a strike-slip margin and is approximately equivalent in scale to the New York-Alabama Magnetic Lineament; however, little evidence exists in the reflection data in eastern Tennessee to support tectonic-scale strike-slip deformation. As imaged on profile ARAL, the crust is dominated by relatively low dipping reflections of substantial continuity. Regions dominated by strike-slip deformation are usually typified by short, disjointed reflections terminating in subvertical faults that can not be imaged on conventional reflection seismic data (e.g. D'Onfro and Glagola, 1983; Roberts, 1983). While the possibility can not be eliminated that originally subvertical deformation features have been rotated into their present low angle orientation and can now be imaged, the continuity of the reflections observed in the middle crust in Tennessee suggests that penetrative deformation is not present as would be expected following the emplacement of a block the length of the New York-Alabama Magnetic Lineament. Therefore, the seismic reflection data herein are interpreted to record an essentially dip-slip tectonic event.

The wedge could be older but show no west-dipping deformation if it behaved as an escaping block. In this scenario, the wedge is interpreted to move upward on boundary faults shown as the eastern margin of the Grenville Front Tectonic Zone and as the contact between the wedge and the crust to the southeast. Such a deformation style might decouple the wedge from the adjacent rocks. This model could account for the density of the wedge by moving it from depth along the GFTZ contact.

The wedge as a younger depositional unit

The wedge might be composed of sedimentary or volcanic material deposited above rocks comprising the more reflective crust to the southeast

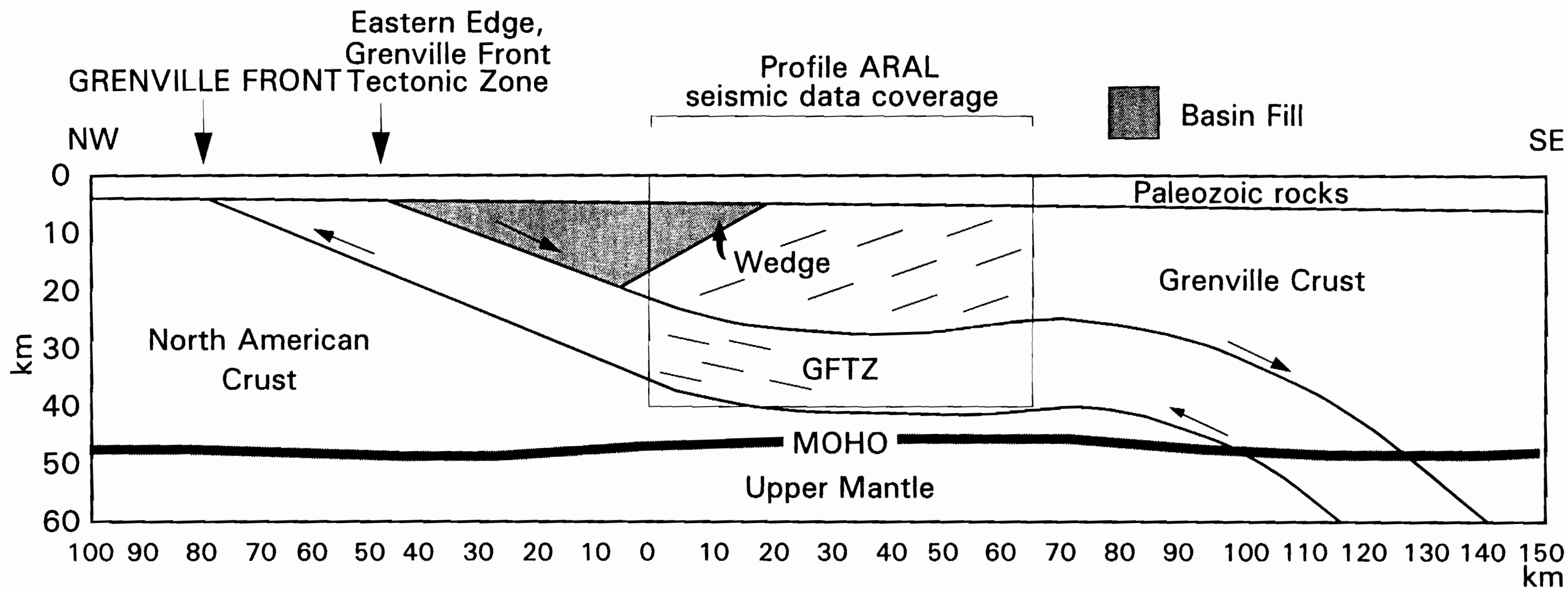


Figure 6.7. Basin fill: Reactivation of the eastern margin of the Grenville Front Tectonic Zone might have produced a basin at the base of the shelf strata. The systematic filling of the depocenter as the detachment surface moved could account for the difference in layer orientation between the wedge and the crust to the southeast, where the wedge exhibits subhorizontal reflections and the crust to the southeast is characterized by strong west-dipping reflections. A basin large enough to account for the size of the wedge is unlikely, and the linearity of the New York - Alabama Magnetic Lineament does not support the interpretation of its source as a basin. Furthermore, the density of the wedge is greater than the density of the adjacent crust, suggesting that the wedge is either the more intermediate in composition or the more indurated of the two crustal blocks.

(Figure 6.7). In this scenario, the wedge is clearly younger (e.g. Iapetan) than the crust to the southeast. Density and susceptibility contrasts between the crustal units are small enough to permit modeling both blocks as depositional. Robinson et al. (1985) and Strangway (1981) discussed such magnetically susceptible units, and indicated that the rocks could have sufficient susceptibility contrasts with adjacent units to produce magnetic anomalies. The seismic reflectivity of the lower unit could originate from impedance contrasts between layers of different composition. This model provides for the parallel orientation of the wedge contact and the reflections; however, the preservation of compositional layering in rocks that have been at amphibolite or granulite metamorphic grades is unlikely.

The wedge as an older unit

An alternative model is that the contact has been completely overturned from its original orientation. In this case, the wedge would have underlain the reflective crust and, unless intrusive, be older. Presence of the higher reflectivity can be explained by compositional layering as in the previous model. This scenario involves considerably more rotation of the deep crust over a significant regional distance in order to maintain the long, smooth magnetic gradient of the New York-Alabama Magnetic Lineament. In addition, the model is weakened by the problem of retention of original layering as expressed above.

Arguments against either part of the crust preserving original depositional layering include the thickness of the wedge. The wedge thickens along a distance of 27 km from about 3 km near the center of profile ARAL to over 12 km along the northwestern end of the profile. As the thickening shows no sign of diminishing, the thickness of 12 km is a minimum. Furthermore, the crust including the wedge has been uplifted and eroded to the base of the Cumberland plateau, requiring the original thickness of the wedge to have been substantially greater. Therefore, the base of the wedge on the northwestern end of profile ARAL must have formed at a minimum depth well over 12 km. Furthermore, during deposition the crust would be isostatically compensated by sinking under the weight of the incoming sediment. Foundering of the crust with depositional loading to a depth of over 12 km across a distance of only 25 km is unlikely for the strong, thick crust of eastern Tennessee. Hence, the wedge probably did not form in a depositional setting.

West-dipping reflections

High reflectivity in the middle crust has been attributed to features developed in compressional and extensional tectonic environments. Several authors have demonstrated that reflectivity can be produced in response to impedance contrasts generated by the presence of mylonite zones (e.g. Fountain et al., 1984, 1987; Passchier, 1986), sills and dikes, basin fill, and volcanic layering. Reflectivity has been attributed to the presence of relatively young magma bodies (Serpa et al., 1988; De Voogd et al., 1988). High reflectivity persists across most of the long profile ARAL in eastern Tennessee, where the west-dipping reflection packages appear to bound coherent geometric blocks. The assumption that the crust was metamorphosed to upper amphibolite to granulite grade during orogenesis suggests that original compositional layering was not preserved. Therefore, the possibility that reflectivity is generated by the mechanisms that involve depositional or pre-metamorphic layering is unlikely; however, the continuity and organization of the west-dipping reflections on the seismic data suggest a common mechanism for the production of the reflective interfaces.

Seismic reflection data in western Canada obtained as part of the Lithoprobe project exhibit striking similarities to the data in this study (Ross et al., 1995). Strong crustal reflectivity on the Lithoprobe data was interpreted as evidence for a 1.8 by old collisional event that produced crustal scale ductile thrusting and imbrication. The Moho shows evidence of deformation that the authors related to orogenic development. The similarity of reflection profile ARAL to the Lithoprobe data strongly suggest that the crustal thickening mechanism for the Grenville orogeny in eastern Tennessee was similar to that interpreted for western Canada. The shortening most likely resulted from ductile folding and thrusting that produced and moved along mylonite zones dipping to the northwest as evidenced by the strong west-dipping reflections on profile ARAL. The deformation could explain the topography observed on the Moho in eastern Tennessee.

West dip has been reported on reflection data in Ohio and alluded to on unpublished COCORP data south of this study area on the Tennessee-Georgia border (Culotta et al., 1990). The west-dipping reflections might record a regional scale deformation event affecting the Grenville crust throughout much of the eastern United States. The continuity of the west-dipping reflections can be interpreted as evidence that the crust was strained as a coherent block throughout its thickness, and might have formed as the ductile equivalent of back thrusting.

On the seismic data, the thickness of the crust dominated by west-dipping events is at least 20 km. Assuming that the crust was deformed before it was uplifted and eroded to the late Precambrian unconformity at the base of the shelf strata, as is suggested by the lack of large scale deformation on that surface, then at least the base of the block experienced deformation under ductile conditions.

Mylonites and ductile shear zones as reflectors

Deformation in a ductile region can be produced by a number of mechanisms, but often involves a significant amount of shear (e.g. Hobbs et al., 1976). The formation of shear zones does not require the presence of favorably oriented compositional or bedding surfaces, permitting them to develop in high grade metamorphic rocks where such original layering is rarely preserved. Shear within the deeper crust has been related to the formation of mylonite zones (Sibson, 1977; Reston, 1990). Because the geometries of blocks enclosed by the west-dipping reflections on profile ARAL resemble duplex structures (Boyer and Elliott, 1982), the interpretation that the west-dipping reflections are bounding mylonite zones is reasonable. Mylonite zones with large, planar areal extents have been described by Francis and Sibson (1973) and Bell (1978).

Originally defined as a brittle phenomenon by Lapworth (1885), the term mylonite is now used for ductile features that retain evidence of dynamic recrystallization. Defining features of mylonites were described by a number of authors including Passchier (1986), Fountain et al. (1984), and Jones and Nur (1982), but are summarized by Hobbs et al. (1976) as "narrow planar regions in which deformation is intense relative to that of the adjacent rocks". They indicated that mylonites are fine-grained equivalents of the adjacent rock in which lineations are defined by elongate mineral grains. Mylonites are described as compositionally layered, containing minerals with preferred orientations resulting from crystal plastic flow, exhibiting flow folds, and maintaining a monoclinic fabric element parallel to the stretching lineation and normal to compositional layering (Passchier, 1986).

If the fabrics interpreted from the west-dipping reflections formed under ductile conditions at or below middle crustal depths, then crustal scale ductile shear zones, or mylonites, are good candidates for the style of deformation suggested by the reflections. Remineralization and crystal plastic deformation along such zones provide the necessary mechanisms for introducing p-wave velocity anisotropy into the crust, leading to the formation of impedance contrasts between the deformed zones and the parent rock without significantly changing the bulk composition of the crust.

Seismic anisotropy has been extensively studied in mylonites by numerous authors. Fountain and Christensen (1989) list the p-wave anisotropy of two amphibolite facies tonalitic gneisses in Connecticut as 0.3% and 5.2% of the p-wave velocities ranging from 5.85 and 5.97, respectively, at

100 MPa to 6.33 and 6.35, respectively, at 1000 MPa. The difference in p-wave anisotropy between these two lithologically similar rocks most likely is due to an difference in foliation development or to a difference in the number of anisotropic minerals between the gneisses. Jones and Nur (1982) found strong velocity anisotropy in two of nine mylonites at low confining pressures up to 1 kb, comparable to pressures found in the upper crust. Siegesmund et al. (1991), using model rocks without microcracks, calculated a range of p-wave velocity anisotropy values from 2% in granitoid rocks to 11% in ultramylonites. Mainprice et al. (1990), using samples and data from the Swiss Alps, calculated large reflectivity coefficients for calcite mylonites found along the Glarus overthrust and Morcles nappe, interpreted to have formed in an upper crustal tectonic environment. They also determined reflectivity for quartz mylonites collected from the Simplon and Insubric Lines formed at middle to lower crustal depths and found that, although the fabric produced a more isotropic p-wave velocity character, anisotropy did exist and could be enhanced by the growth of phyllosilicates. Fountain et al. (1984) suggested that retrograde effects, more common along highly deformed zones than in undeformed rock nearby, might enhance seismic reflectivity by remineralization as well.

Thin bed tuning and reflectivity

A mechanism that can account for the strength of the west-dipping reflections is thin bed tuning. Amplitudes of wavelets reflected from thin beds of optimum thickness can be increased up to twice that of a reflected wavelet from a single velocity interface at a given impedance contrast. Optimum tuning thickness is dependent on the frequency content of the incoming wavelet (e.g. Sengbush et al., 1961). Because mylonite zones range in thickness from millimeters to hundreds of meters (Chroston and Max, 1988; Schmid et al.,1987), they provide for tuning of a wide range of input frequencies. Fountain et al. (1984) modeled a number of exposed mylonites and showed amplitude increases by factors of two for the appropriate tuning frequencies.

In this study, vibroseis source frequencies fell between 14 and 56 hertz, with transmission effects and the extended correlation technique resulting in the loss of higher frequencies with depth. High frequency loss due to the extended correlation technique occurs below 3 seconds at the rate of 3.4 hz/sec for profiles ARAL, KR, and LE-1 and LE-2, and at the rate of 1.9 hz/sec for profiles ARDU and FM. Because the west dipping reflections are imaged mainly below 3 seconds, the thicknesses of the tuned layers can be expected to be greater with depth in accordance with the net lower frequency content of the data. While a clear seismic amplitude comparison between the tuned layer and the source wavelet is necessary to quantitatively describe tuning thicknesses, a high amplitude return from crustal depths in regions expected to be of essentially the same bulk composition suggests the presence of tuned layers. Assuming an average crustal velocity of 6500 m/s two-way travel-time, the thin, tuned layers can be expected to range in thickness from about 30 meters for 56 hertz (above 3 seconds) to about 110 meters for 14 hertz (possible throughout the reflection data). Therefore, mylonites ranging from 30 to 110 meters thick could be responsible for the west-dipping reflections with thinner layers more likely shallower in the section.

Subvertical low reflectivity regions

The subvertical zones of lower reflectivity might indicate the presence of dike swarms related to the post-Alleghanian intrusions mapped on the surface in eastern Tennessee, Kentucky, and Pennsylvania. Such intrusions might be imaged on reflection seismic data as subvertical regions of low reflectivity similar in character of dike swarms interpreted by Çoruh et al. (1988) in Virginia; however, the absence of potential field anomalies over the low reflective regions on seismic data does not encourage the interpretation of the zones as mafic intrusive complexes. For example, the Norris Lake kimberlite is near the northwestern edge of the study area and has a pronounced magnetic signature (Johnson, 1961). In addition, low magnetic susceptibilities and densities on gravity and magnetic models (this study) indicate felsic compositions. Furthermore, the values are more similar to those assigned to the southeastern part of the crust than they are to the values assigned to the wedge, suggesting a stronger compositional affinity between the low reflective zones and the southeastern crustal block.

Paleozoic and younger deformation

The crystalline basement has experienced considerable regional uplift since the deposition of the rift and drift rocks associated with the development of the Iapetan Ocean without significant rotation as evidenced by the low angle of dip on the detachment surface. The absence of rotation on that surface implies that major crustal deformation in eastern Tennessee had ceased by the Cambrian and that subsequent movement of the basement was limited to regional vertical uplift. Such uplift could produce compensatory fracturing, but the continuity of the reflection near the base of the Valley and Ridge and Cumberland Plateau at 1.5 seconds (4.5 km) indicates that offset along such fractures is small. Compressional tectonic activity in eastern Tennessee during the Alleghanian orogeny appears to be restricted to the emplacement of thin-skinned thrust sheets in the Valley and Ridge Province. The structural front, visible at cmp 625 on line ARAL-3 on the northeastern end of profile ARAL, defines the westernmost extent of deformation exposed at the surface that is associated with the late Paleozoic thrusting. The continuity of reflections above the bright reflection near the base of the Paleozoic shelf strata suggests little internal deformation of the strata within the thrust sheets and that many of the sheets sole into the bright reflection. The rocks associated with the bright reflection apparently acted as a regional detachment surface for many of the thrusts.

Recent earthquake activity

The trend of epicenters in the Eastern Tennessee Seismic Zone runs parallel to and near the New York - Alabama Magnetic Lineament. This correlation has been postulated to be an indication of a significant crustal change by King and Zietz (1978) and by Johnston et al. (1985). Johnston et al. proposed that the earthquakes occur within a part of the crystalline basement termed the Ocoee block between the New York - Alabama Magnetic Lineament and Clingman Lineament identified by Nelson and Zietz (1983). From this study, the majority of the foci are found to be located within the region of the crust dominated by west-dipping reflections, with few foci in the wedge or above the detachment surface in the shelf strata. The results of this study are consistent with that interpretation except that the contact between the seismogenic crust and the non-seismogenic block to the northwest (the wedge) does not have a vertical orientation. The integration of the dipping contact into crustal velocity models and inversions used in regional earthquake investigations is likely to improve hypocenter locations.

The possibility that a seismogenic crustal block might be delineated by the magnetic anomalies in this region is provocative; however, the seismicity is not everywhere present between the anomalies, nor does seismicity appear to fall in all locations where west-dipping reflections are reported (e.g. Ohio, Pratt et al., 1989). Furthermore, earthquake focal mechanisms indicate that most of the failure mechanisms within the Eastern Tennessee Seismic Zone are strike-slip on subvertical failure planes oriented N-S or E-W. While these orientations have been employed in the determination of the regional stress field for the eastern United States reported by Zoback and Zoback (1989), they are not compatible with failure oriented along the west-dipping reflections. The formation of the west-dipping reflections probably occurred during or associated with Grenville orogenesis, and the boundaries the reflections record are not being reactivated in the modern stress field.

Favorably oriented fractures could be reactivated in a stress field in which the minimum and maximum compressive stresses were horizontal as indicated by the orientations of the p-axes and nodal planes of the earthquake focal mechanism solutions. Uplift of a crust from confining pressures of hundreds of MPa might permit the opening of subvertical fractures with little noticeable displacement, as would broad, regional arching or the crustal block. Such small fractures could be reactivated in the modern stress field of the eastern United States.

The N45°E trend of the New York - Alabama Magnetic Lineament follows the crustal thickness maps of Taylor (1989) and James et al. (1968). The Eastern Tennessee Seismic Zone and the Giles County Seismic Zone fall to the southeast of the Lineament and along the axis of maximum crustal thickness. As the rocks of the Alleghanian allochthon are removed by erosion, the thick crust, if felsic, should isostatically compensate for the loss by rising. Significant uplift rates have been proposed by Prowell (1983) for the Appalachian orogen and by Sasowsky (1992) for eastern Tennessee. Differential erosion rates and delay in rebound could account for the scattered seismicity along the New York - Alabama Magnetic Lineament and along the region underlain by thickened crust; however, the strike slip orientation of the nodal planes along the Eastern Tennessee Seismic Zone and Giles County Seismic Zone in Virginia does not support significant vertical stresses as would be expected in an isostatic compensation situation. Using earthquake focal mechanism solutions, Bollinger et al. (1991) suggested that the maximum compressional stress in eastern Tennessee is oriented N50°E. If the erosion rates vary significantly along the Appalachian orogen, uplift rates should also vary. Such variation might place part of the orogen in locally distinct stress fields and explain the presence or absence of earthquake activity along the axis of thick crust. Furthermore, stress differentials imposed by uplift variations might explain the local compressional and strike-slip mechanism of the earthquakes in the southeastern United States.

The relationship between the eastern Tennessee seismic zone and the

NYAML is uncertain. If the gradient of the NYAML is produced by the contrast between the wedge and the crust to the southeast (in which the earthquakes fall), then the 30^{o} NW dipping boundary between the crustal blocks should be important in the location of the foci. At present, most authors (e.g. Johnston et al., 1985) have noted the alignment of the earthquakes along the mapped position of the gradient. The foci might fall within the crust characterized by west-dipping reflections and be constrained by the contact between the wedge and the reflections. If so, then the seismogenic crust in eastern Tennessee has a distinct reflection character that might be useful in delineating regions more susceptible to seismic activity. Better velocity models developed using the boundaries visible on the reflection data will aid in more accurate locations of the foci and help constrain this interpretation.

The basement to the southeast of the wedge is more seismogenic than the crustal block comprising the wedge. If the wedge absorbs strain energy that is sufficient to produce brittle failure in the crust to the southeast, the wedge might act as a stress concentrator relative to the adjacent rocks. As strain energy is released in the basement adjacent to it, a stress differential between the wedge and the adjacent crust will develop. There exists a possibility that such a stress differential might produce a significant earthquake. Such an event could be expected to have a long recurrence interval to allow for differential stress to redevelop following the compensating seismic event. The most likely location for a stress differential to form is in a region where discrete crustal blocks behave in seismically distinct ways. The eastern Tennessee seismic zone represents such a region.

The present location of the Eastern Tennessee Seismic Zone might reflect the favorable combination of (1) the build-up of a stress differential due to the presence seismogenically different crustal blocks, (2) a higher isostatic rebound rate for the deep crust (delayed from but produced by higher erosion rates reported for eastern Tennessee), and (3) the presence of an appropriate triggering mechanism (e.g. Costain et al, 1987a,b). The use of reflection data to verify the presence of seismogenic crustal blocks and to refine velocity models used in earthquake hypocenter locations is indicated to explore these relationships in eastern Tennessee.

Conclusions

The wedge is interpreted to be a distinct crustal block on reflection seismic data, gravity models, and magnetic models. Although the coverage provided by the reflection data is limited (the longest profile is 70 km), the New York-Alabama Magnetic Lineament (NYAML) can be reproduced by modeling a susceptibility contrast between the wedge-shaped body on the northwest end of profile ARAL and the crust to the southeast. Two-dimensional gravity and magnetic modeling support the interpretation that the wedge is compositionally distinct from the crust to the southeast. The models indicate that the wedge is more dense and susceptible than the adjacent crust, and that the wedge is the most likely source of the NYAML. A possible explanation for the formation of the wedge is as an anatectic melt complex of batholithic scale. Such a complex could have been generated during Grenville thermal events along the axis of thickest crust.

Strong reflectivity in the southeastern part of the profile can be attributed to the presence of west-dipping mylonitic fabrics. The fabric could represent deformation produced during continent-continent collision. The absence of the fabric in the wedge suggests that the emplacement of the wedge postdates the formation of the fabric. This interpretation is consistent with anatectic melting of the thick crust following collision.

A band of higher reflectivity is imaged on all reflection profiles at approximately 9 seconds depth. On profile ARAL, the band has a broad synformal geometry and forms the base of the region dominated by west-dipping reflections. Below the band, east-dipping reflections can be seen along the northwestern end of the profile. The band and east-dipping reflections are interpreted as the southern United States extension of the Grenville Front Tectonic Zone (GFTZ). Extrapolation of the zone to the surface positions the Grenville Front from 60 to 90 km northwest of the northwestern end of profile ARAL. This location brackets the location of the Grenville Front interpreted by previous authors 80 km from the end of the profile.

The deep profiles FM and ARDU appear to image the Moho at about 14 seconds. The higher crustal velocity used to convert the times to depths justified by the high velocity shelf strata and the well indurated crystalline basement indicate the depth to the Moho at about 45 km. The depth is consistent with that proposed by previous authors. The interpreted Moho reflections form a band of higher reflectivity extending for approximately 1 second. The reflection data corroborates the refraction data used by Prodehl, et al. (1984) and suggests that the Moho beneath eastern Tennessee is transitional. Profile FM indicates that the crust thickens to the northwest by about 3 km. The sloping Moho might be evidence supporting the presence of thinner crust beneath the northeastern part of the study area.

Subvertical dike swarms of granitic or granodioritic compositions are interpreted to permeate the crust. The dike swarms appear to be older than the base of the shelf strata and might be related to Iapetan extension.

The absence of large-scale relief on the bright reflections at the base of the Alleghanian allochthon and beneath the Cumberland plateau indicates that crustal scale deformation had essentially ceased by the time these Late Precambrian shelf rocks were deposited. The crust beneath the shelf strata had been uplifted and eroded to considerable depth prior to deposition of the shelf strata as evidenced by the strong subhorizontal reflection at the base of the Valley and Ridge and Cumberland Plateau. Many of the thrusts within the allochthon sole into the bright reflection suggesting that the weak units forming the detachment surface remained essentially undeformed until the Alleghanian orogeny.

Pseudomagnetic field investigations indicate that the region northwest of the steep gradient of the NYAML is intruded by subvertical bodies of mafic composition. The sources have high densities in comparison with the source of the NYAML, which appears to be felsic. The observed magnetic signature can be reproduced over the region northwest of the NYAML by the pseudomagnetic technique, but can not be replicated along the Lineament where Poisson's relation is not satisfied. Removal of a regional gravity field determined from crustal thickness data does not affect the pseudomagnetic field character. The residual field indicates that the overprint of a thick crust has not obscured the presence of a signature that could satisfy Poisson's relation and produce a pseudomagnetic field comparable to the observed magnetic field. Hence, the source of the New York - Alabama Magnetic Lineament is interpreted to be the boundary between more dense and magnetically susceptible rocks to the northwest and less dense, less susceptible crust to the southeast.

The relationship of the eastern Tennessee seismic zone (ETSZ) and the NYAML is unclear. The mapped position of the NYAML refers to a dipping boundary between different crustal types and not to a vertical contact. The locations of the foci in the ETSZ, if related to the NYAML, should fall within the southeastern crustal block, which extends beneath the wedge to the end of reflection profile ARAL. Earthquakes are not found along the length of the NYAML and are not reported wherever reflection data records west-dipping reflections in the crust (e.g. Ohio). More work is needed to combine reflection data with earthquake locations in the ETSZ to explore these relationships.

Earthquake foci in the Eastern Tennessee Seismic Zone indicate mainly strike-slip motion. The west-dipping reflections dominant in the seismogenic region of the crust are not favorably oriented for reactivation along subvertical fractures as required by the focal mechanism solutions, but can be interpreted as evidence that more deformation occurred in that part of the crust. Subvertical fractures can not be imaged on these reflection data but could have been produced by uplift in response to erosion and isostatic adjustment of the thick felsic crust. Differential erosion and uplift rates along the Appalachian orogen as reported by Prowell (1983) and Sasowsky (1992) could produce local stress fields with different orientations than the regional field and account for the focal mechanism solutions determined for the regional earthquakes. Continuity of the west-dipping reflections suggest that little vertical or horizontal offset across the fractures has occurred and can be interpreted as evidence of localized strain release. Furthermore, the contact between the wedge and the adjacent crust might be such that a stress differential could develop across the boundary as strain is released in the crust to the southeast. A stress differential might take centuries to build up and could induce large earthquakes with long recurrence intervals.

The observations and interpretations set forth in this manuscript relate the Grenville Front and the NYAML tectonically and provide an explanation for the location of earthquakes in the eastern Tennessee seismic zone. The identification of the wedge-shaped body within the Grenville basement has provided an explanation for the New York - Alabama Magnetic Lineament. The presence of east-dipping refections that can be attributed to the Grenville Front Tectonic Zone constrains the location of the Grenville Front in eastern Tennessee, and the presence of west dip in the crust to the southeast of the wedge might relate to crust with similar reflections in Ohio. The complicated crustal structure found in eastern Tennessee should be integrated into velocity models used for the accurate location of earthquake hypocenters. Additional reflection data and modeling will extend the interpretations suggested here and provide more insight into the crustal structure of the eastern United States.

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Appendix 1: Seismic Data Acquisition

The nine seismic profiles used in this research were collected by industry for the purpose of illuminating structures within the Valley and Ridge tectonic province of eastern Tennessee. The data acquisition parameters are shown in Table A1.1 for each line. The seismic lines were acquired along roadways oriented perpendicular to the strike of structures in the Valley and Ridge wherever possible; the strike of most structures ranges from N50°E to N65°E, hence the lines trend nominally N30°W with the exception of LE-2, which trends N10°W in response to more easterly striking structures in that area. Eastern Tennessee is a mountainous region where available roadways are necessarily crooked due to the considerable topographic relief. Thus, the seismic lines vary locally in their orientation to structural strike, and, because they are crooked and structures are dipping, in the subsurface coverage they provide.

All data were acquired using 3 to 5 P-wave source vibrators arrayed in-line. Sources were located midway between surveyed stations and arranged bumper to bumper with pads spaced 10 to 15 m apart. The latter spacing was employed if units were down and the shooting was progressing with fewer vibrators. From 6 to 20 sweeps were summed depending on the number of operational vibrators, with fewer sweeps summed for more vibrators. Drive levels were set at the maximum levels possible without decoupling the pad from the ground; cultural conditions occasionally required reductions in drive level. Skips in vibration points occurred on all lines in response to cultural conditions such as bridges and hospitals, or at the landowner's request.

Vibroseis upsweeps of 14 to 56 hertz were used for all lines. Such data can be extended to depths beyond the full correlation listening time. Upsweep data does not have the problem with correlation ghosts that can accompany extended correlation of downsweep data (Waters, 1987). Traces were recorded for 16 seconds with 13 second sweeps (ARAL-1, ARAL-2, ARAL-3, KR-2, LE-1, LE-2) or for 23 seconds with 20 second sweeps (ARDU-1, ARDU-2, FM-1) for a full correlation time of 3 seconds on all lines. All data were sampled at 4 ms.

Table 1: Acquisition Parameters

	ARAL 1	ARAL 2	ARAL 3	FM 1	KR 2	ARDU 1	ARDU 2	LE 1	LE 2
Year acquired	1982	1982	1982	1983	1983	1982	1982	1982	1982
Station spacing (m)	50	50	50	67	67	67	67	67	67
Split spread *	S	S	S	A	S	A	A	A	A
Near offset (m)	226	226	226	436	302	369	369	436	**
Far offset (m)	2591	2591	2591	1174 2784	1845	1174 2716	1174 2716	1174 2784	***
Sweep freq (hz)	14-56	14-56	14-56	14-56	14-56	14-56	14-56	14-56	14-56
Sweep length (sec)	13	13	13	20	13	20	20	13	13
Record time (sec)	16	16	16	16	23	23	23	16	16
Vibrators (#)	3-4	4	4	5	5	4-5	5	4-5	4-5
Recorder	DFS-V/ FT-1	DFS-V/ FT-1	DFS-V/ FT-1	DFS-V/ FT-1	DFS-V/ FT-1	CFS/ DFS-1V	CFS/ DFS-1V	CFS/ DFS-1V	CFS/ DFS-1V
Channels	96	96	96	48	48	48	48	48	48
Phones / trace	48	48	48	36	36	36	36	36	36
Geophone type	L10A	L10A	L10A	20-D	20-D	20-D	20-D	20-D	20-D
Sample rate (ms)	4	4	4	4	4	4	4	4	4
CDP fold	48	48	48	24	24	24	24	24	24

* A = asymmetric split spread, S = symmetric split spread.
** 302 m from VP1a toVP110b, 436 m from VP111a to line end.
*** 1040 m and 2649 m from VP1a to VP110b, 1174 m and 2784 m from VP111a to end.

Appendix 2: Seismic Data Processing

Seismic data were reprocessed from field tapes or demultiplexed tapes on a VAX 11/785 mainframe computer with Cogniseis DISCO software versions 8.0 and 8.1.

Preprocessing steps

Demultiplexing, diversity stack, and spectral whitening

Where available, field tapes were used. Field tapes were demultiplexed for lines ARAL-1, ARAL-2, ARAL-3, FM-1, AND KR-2. Lines ARDU-1, ARDU-2, LE-1, and LE-2 were received as demultiplexed tapes. Because the data were acquired using a diversity stack option, recovery of true reflection strength was impossible. The diversity stack process (Embree, 1968) weights components in inverse proportion to the average power over selected intervals. A discussion of the diversity stack option with comparisons to other trace summing methods can be found in Gimlin and Smith (1980). While useful for improving the signal to noise ratio, the true amplitude information is lost, and a spectral whitening effect is imparted to the data. Hence, spectral whitening methods such as vibroseis whitening (Çoruh and Costain, 1983) did not enhance the data and were not performed at this time.

Extended correlation

The selection of vibroseis upsweep data for analysis was deliberate: the extended correlation technique permitted the imaging of the deep crust without the destructive correlation ghosting associated with extending the correlation of vibroseis downsweep data (Waters, 1987). The procedure is explained by Pratt (1986) and Okaya and Jarchow (1988). The lengths of the sweep and recording time determine the depth to which the extended correlation can be performed and retain a full octave of frequency content. seismic lines ARAL-1, ARAL-2, ARAL-3, KR-2, LE-1, and LE-2 were acquired with 13 second sweep lengths and 16 second record lengths. The full correlation window was 3 seconds, within which the full spectrum of frequencies (14 to 56 hertz) could be recovered. The records were padded to permit correlation to 11.25 seconds, with high frequency losses of 3.4 hertz/second. At 11.25 seconds, the frequency content of the data was one octave (14 to 28 hertz). Seismic lines ARDU-1, ARDU-2, and FM-1 were acquired using 20 second sweep lengths and 23 second record lengths. These records could be extended to 18 seconds with a full octave of frequency remaining. The frequency drop-off rate for the lines was 1.9 hertz/second below 3 seconds.

Correlation tests on ARAL-1 revealed that the use of auxiliary trace recordings of the summed vibroseis sweeps yielded variable results due to the amplitude variation in the auxiliary sweep records. Hence, a synthetic sweep of 14 to 56 hertz was chosen for correlation and was used for all lines. No attempt was made in these processing efforts to accommodate the numerous possible phase problems elucidated by Newman (1994). Newman admonishes that serious phase distortion occurs in the acquisition and processing of vibroseis data that result in variable wavelet shapes. Such phase changes often result in the data not tying to impulsive-source data or to vibroseis surveys acquired and processed differently. The use of an unfiltered synthetic sweep partially alleviates the problem of correlating with a phase-shifted field filtered pilot sweep, but does not pointedly address the problem of unknown phase of the input data. The high propagation quality (Q) of the eastern United States crust probably argues most strongly for the use of traditional processing methods with less phase distortion than would be expected in a region dominated by low Q. In this study, no attempt is made to tie these data to each other or to other types of data. In addition, the use of these data is restricted to structural interpretations and thereby avoids errors that would be problematic in stratigraphic analyses that rely on consistent waveform characteristics among lines.

Line geometry

All lines were acquired along crooked mountain roads with in-line arrays of sources and receivers. Details of individual line geometries can be found in the acquisition section. Station spacings on lines ARAL-1, ARAL-2, and ARAL-3 are 50 m while the remaining lines FM-1, KR-2, ARDU-1, and ARDU-2 have station spacings of 67 m. Lines LE--1 and LE--2 were acquired with redundant shots. As single shot signal-to-noise ratio was excellent on these lines, vertical stacking of the redundant shots was deemed unnecessary. Instead, the repeated shots were assigned to locations between the original stations at 1/2 the station spacing to double the subsurface coverage. The procedure is justified by the size of the Fresnel zone and the inaccuracy of locating the true positions of common depth points in regions comprised of dipping layers. The procedure results in a new line geometry for the two lines in which the station spacing is 16 m.

Datum statics

Most of the terrane over which the data were acquired is mountainous and structurally deformed. The geologic map of eastern Tennessee reveals the presence of imbricate stacks of thrust sheets within which are numerous carbonate layers. Datum statics were applied to all lines to normalize the topographic datum: lines ARAL-1, ARAL-2, ARAL-3, ARDU-1, ARDU-2, and KR-2 are hung on an average elevation of 300 m above sea level, and lines FM-1, LE-1, and LE-2 are hung on an average elevation of 350 m above sea level. Datum statics corrections were estimated from the normal moveout of selected shots for each line.

Processing

Sort, stack, and residual statics

Shots were edited before sort. All lines were acquired along crooked roads. To improve homogeneity of fold, the lines were assigned straightened positions crossing the center of groups of common mid points before sorting

(Figure A2.1). The presence of dipping layers and the size of the Fresnel zone justify this approximation to the positioning of the lines as the true positions of the cmps are not known in the absence of complete subsurface information. The procedure resulted in nominal folds of 48 for lines ARAL-1, ARAL-2, and ARAL-3, and folds of 24 for lines KR-2, FM-1, ARDU-1, ARDU-1, LE-1, and LE-2.

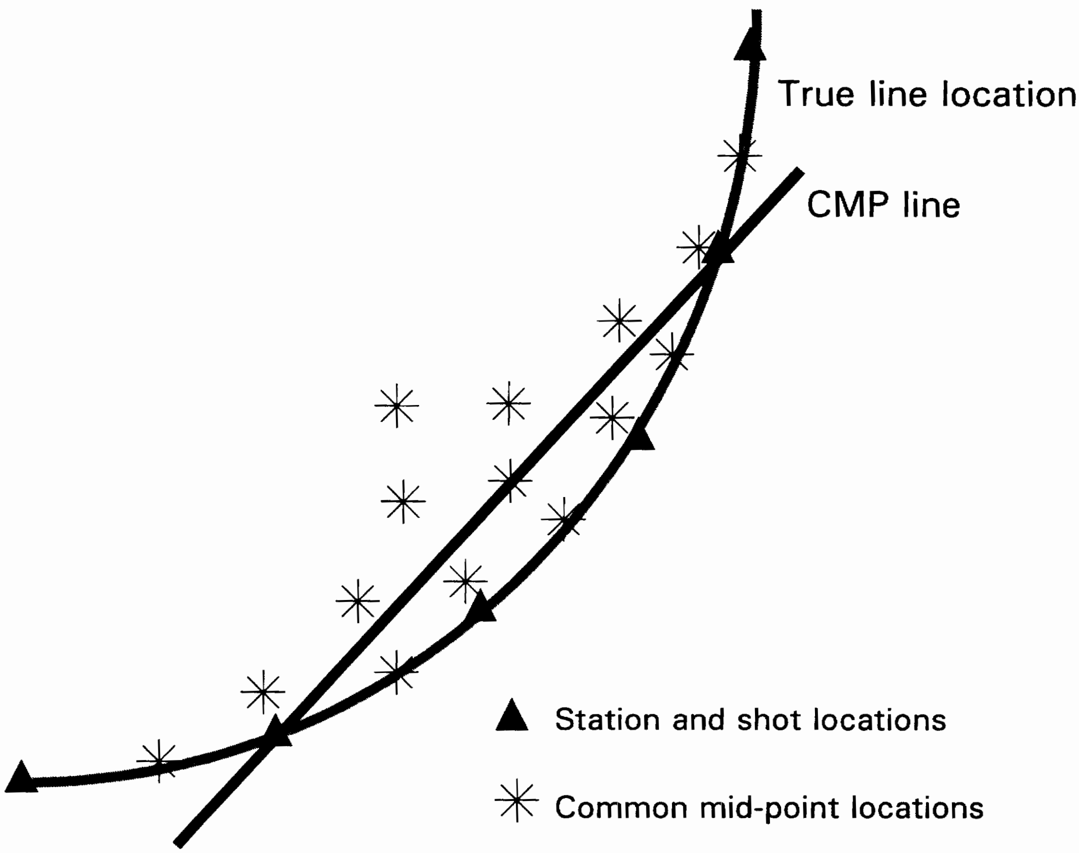

Figure A2.1. CMP scatter plot for a crooked line: The solid dark line indicates the position of the CMP line established between known station locations to more closely approximate the presumed locations of the mid points. The resulting CMP line results in more homogeneous fold.

Brute stacks imaged a strong subhorizontal event on all lines at approximately 1.5 seconds. This reflection provided the basis for the preparation input for the residual statics runs. At least four residual statics were run for each line with 4 iterations within each run. At minimum of four constant velocity analyses were performed for each line, and stretch mutes were applied to correct for NMO-correction stretch.

Deconvolution

The spectra of the data were whitened using pre-stack prediction error deconvolution. Along with the removal of multiples, the compressed wavelet permits easier velocity picking. Operator length and lag tests were run to assure removal of long period multiples. The presence of high velocity carbonates in the near surface produced multiples with periods of varying lengths, and operators were chosen for each line to suppress the energy observed in the autocorrelations. Enough points were allowed in the operator to account for long period multiples if they were present. Lags were chosen at the first zero crossing near a full sample (Yilmaz, 1987). Post-stack deconvolution tests were run and deconvolution applied to stacked data to eliminate remaining multiple energy and to whiten the data after the NMO correction process. Tests for operator length and lag were run on stacked data to optimize the deconvolution parameters.

Final processing and displays

Iteration of velocity analyses and statics were performed until satisfactory stacks were obtained. These stacks were used to generate automatic line drawings for interpretation. Automatic line drawings are an unbiased approach to estimating reflector continuity by determining the continuity of signal on adjacent traces within a user-specified dip range. The dip ranges were sufficient to accommodate the steep reflections known to exist in the allochthon. The angle used in the signal coherency algorithm was allowed to be high to eliminate inadvertent dip filtering.

Migration

Although considerable dip was present in reflection data in this study, the lines do not have sufficiently long recording times or offsets to migrate the data properly for the depths of interest. Relatively poor signal to noise ratio and limited velocity control combine to reduce the accuracy of migration of the deeper data. Because the data are dip lines and are not constrained by strike line ties, the interpretation could be affected by out-of-the-plane sources (e.g. Serpa and Dokka, 1992). To address the possibility, shot gathers were examined. The relatively weak signal-to-noise present at the depths of interest obscured much of the primary reflection energy, but out-of-the-plane sources could be detected only on gathers for profile FM, which occasionally recorded side-swipe between 1.5 and 6 seconds. The lines comprising profile ARAL appeared to be unaffected by out-of-the-plane energy.

Lynn and Deregowski (1981) discussed dip limitations on migrations as functions of line length and recording time. They illustrated the problems encountered in their Figure 5, modified here as Figure A2.2. The diagram, based on a 60 km long line with a 20 second record length, shows perpendiculars to layers dipping to the left as do most of the reflections imaged on the data in this study. Points along the appropriate dip perpendicular can be projected vertically and horizontally to indicate the needed offset and two-way migrated recording time necessary to properly migrate the data. The figure represents the longest line in the study, line ARAL-1 in profile ARAL, for which the line length is 52 km and the record length is 16 seconds. On line ARAL-1, a reflector at 12 seconds dipping 30^{o} could be imaged at a minimum offset of 20 km. Measuring from each end of the line, such dips could only be imaged over the center 12 km of line ARAL-1. Shallower dipping layers could be migrated over longer distances, but the structures of interest in this study extend from 1.4 seconds to over 10 seconds. Furthermore, the shorter lines included in profile ARAL could not be migrated with the same result as the long line, and the resulting migrated composite would be discontinuous at each line intersection. Therefore, the data are not migrated and the possibility of the presence of diffractions and other artifacts must be considered in their interpretation.

The unmigrated data image the events in the deeper crust reasonably well for the relatively shallow dips present; the dips do not appear to be steep enough to migrate great distances. The angle of 20^{o} at the contact between the wedge and the southeastern part of the crust would migrate to a dip of only 21^{o} using the "Migrator's equation" of $\sin\theta = \tan\phi$, where θ is the angle of dip of the migrated layer and ϕ is the angle of dip of the unmigrated layer. The approximation is only appropriate for ideal situations where no outside effects come into play, but suggests that the unmigrated data are imaging the subsurface fairly well.

Additional constraints on the migration of crustal data were discussed by Warner (1987), where he pointed out that deep crustal data is affected by near surface discontinuities that act to produce spurious reflection points in the deeper crust. When migrated, these points smear into migration smiles, destroying the information on the remainder of the data.

The reflection data used in this study are of excellent quality for crustal data and can be reliably interpreted in their unmigrated form as long as the interpreter considers their limitations. The data can not be processed for true reflection strength due to the diversity stack option employed in the field, and the data should not be taken to indicate true strike of structures without the addition of out-of-the-plane control. The use of additional and independent geophysical tools such as potential field and earthquake data is indicated to accurately assess the information on these reflection profiles.

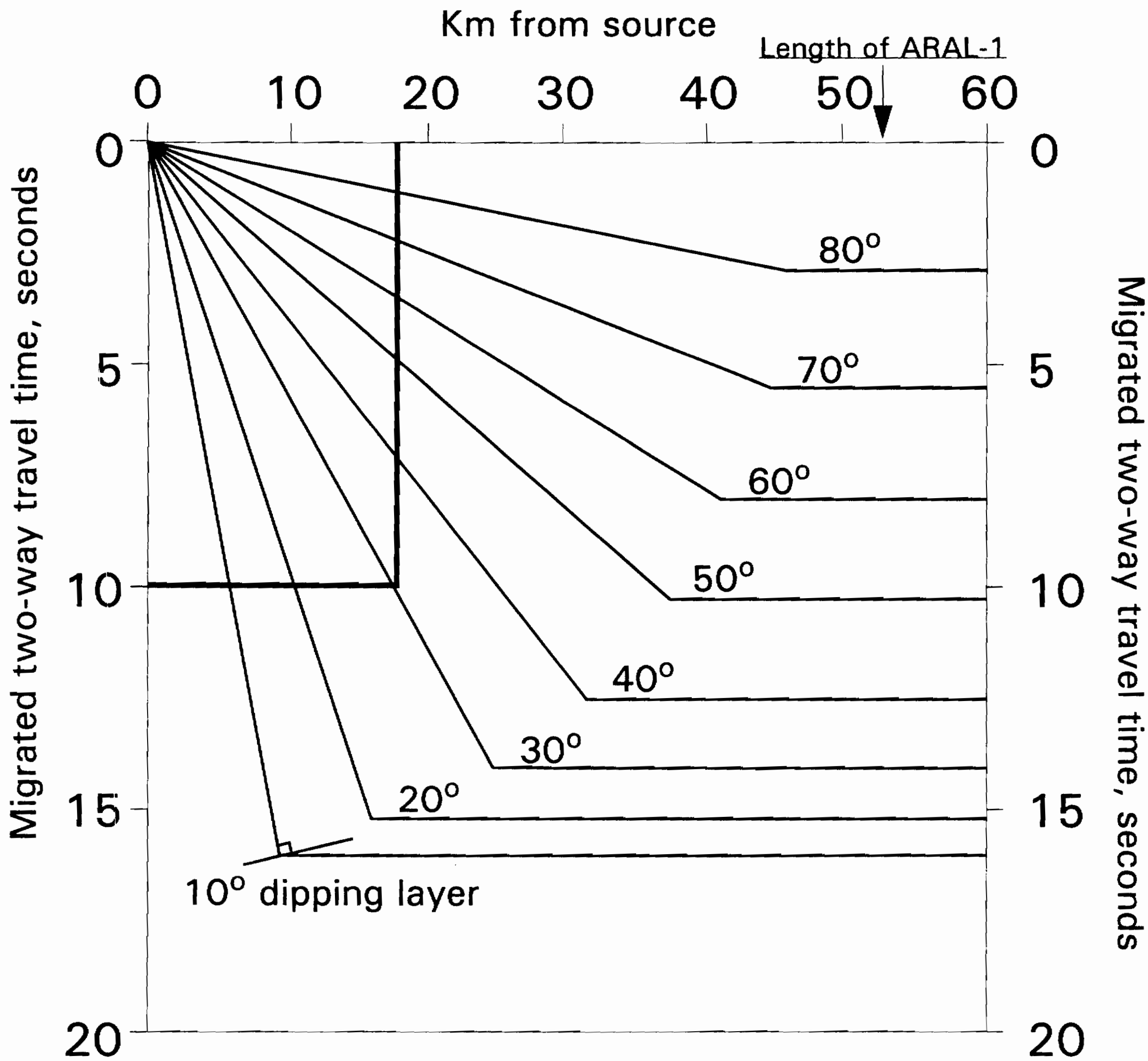

Figure A2.2. Offset vs. time migration limitations: The lines originating from the origin are the perpendiculars to the dipping layers observed on the data. To properly migrate 30° dipping events at 10 seconds, the maximum depth of the west-dipping reflections on profile ARAL, the minimum offset is 17 km (dark line). The length of the longest seismic line in the survey, line ARAL-1, is shown at 52 km. Dips in the study area are predominantly in one direction, but to accommodate regions containing opposite dips, the graph must be duplicated across the left vertical axis and appropriate offsets for the depths of interest must be present in both directions. Hence, most of the data in the survey are too short to be migrated for crustal events. Modified from Lynn and Deregowski, 1981.

Vita

Debbie L. Hopkins was born in Pocatello, Idaho, to Ben Browning Hopkins and Audre Dewey Hopkins. A deep appreciation of the natural and physical sciences sprang from her childhood in Oregon and Utah, and was nurtured by her grandparents, Thomas and Lorraine Dewey. Following graduation from Highland High School, she attended the University of Utah. Deb majored in Biological Science while she worked variously as a laboratory assistant in Anthropology, Paleontology, and Entomology, and as a ski instructor.

Upon gradation, she became supervisor of a medical research laboratory for the Department of Anatomy at the University of Utah Medical Center. She performed cell cycle and RNA extraction experiments to characterize the genetic disease cystic fibrosis. After leaving the University, she worked as a medical technologist and waitress before obtaining a position with Phillips Petroleum's Woods Cross Refinery as a manual laborer. She worked her way up the union ranks to Instrument-Electrical Helper before she left to return to school.

Deb completed a Master of Science degree in Geology at the University of Utah. Her major emphasis was in Structural Geology with Ronald Bruhn, and her minor emphasis was in Earthquake Seismology with Robert Smith. Following completion of her M.S. degree she worked for Conoco, Incorporated, in Houston, Texas, as a geophysicist.

Following her departure from Conoco, Deb chose to pursue a Ph.D. degree in Geophysics with John K. Costain at Virginia Tech. She plans to work in industry or academe following the completion of her degree.