Gravity and Magnetic Modeling of Basement Beneath Alabama Gulf Coastal Plain

by

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Abstract

The southeastern United States has experienced two complete successions of Wilson cycles: (1) the assembly and break up of Rodinia and the opening of the Iapetus ocean; and (2) the closing of Iapetus ocean, the assembly of supercontinent Pangaea and its subsequent break up, and the opening of modern Atlantic Ocean. Evidence of these supercontinent cycles are recorded in the rocks of Alabama and adjacent areas, but in the southern portion of the state, these rocks are covered by as much as 7 km of Coastal Plain sediments. The Grenville-aged basement rocks beneath the Plateaus, the Valley and Ridge, and most of the Piedmont provinces have Laurentian origin, but the Uchee terrane in the Southern Piedmont province to the south is interpreted to have exotic Peri-Gondwanan origin. Also to the south, rocks of the Suwannee terrane are thought to have a Gondwanan origin based on faunal assemblages found in well cores. Triassic to Upper Jurassic sedimentary rocks of South Georgia basin onlap the northern limit of Gondwanan-affiliated Suwannee terrane rocks and obscure the suture between Laurentia and Gondwana.

In this study, I use airborne gridded gravity and magnetic data to develop crustal models along three transects that cross major tectonic structures, geophysical anomalies, and the ancient North American (Laurentian) margin. Models derived from gravity and magnetic data are constrained by well-log information, geologic mapping, and previous geophysical studies in Alabama and nearby areas. Results show that a pronounced east-west trending gravity low observed in southern Alabama can be interpreted as the suture between relict Gondwanan crust and Peri-Gondwanan/Laurentian crust. The denser crystalline rocks of the Piedmont and Valley and Ridge provinces correspond to minor gravity highs. Based on its distinctive gravity and magnetic properties, the Wiggins terrane in southwestern Alabama is interpreted as a unique tectonic terrane. The eastern boundary of Wiggins terrane with Gondwanan crust is delineated by another prominent gravity low. Based on its magnetic expression, Laurentian crust is thought to continue beneath the Coastal Plain sediments until it is truncated by the tectonic suture with Gondwanan-affiliated crust. This truncation is marked by the Brunswick magnetic anomaly (BMA). Sharp magnetic gradients and long-wavelength gravity gradients along faults such as the Towaliga fault, Alexander City fault, and Bartletts Ferry fault suggest these structures are major, crust-penetrating features.

Gravity and magnetic modeling reveal the thickening of crust from south to north, and a change in crustal thickness near the suture zone. Results suggest that the crust beneath the Wiggins Arch, in western to southwestern Alabama, is similar to that of Mississippi Gulf coast and most closely resembles a transform margin. Unlike the Texas Gulf coast, however, there is no evidence of a volcanic rifted margin in Alabama. The gravity and magnetic data are consistent with the presence of Mesozoic rift basins.

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ACF = Alexander City fault

BF = Brevard fault

BFF = Bartletts Ferry fault

BMA = Brunswick magnetic anomaly

COCORP = Consortium for continental reflection profiling

GEF = Goodwater-Enitachopco fault

GRF = Goat Rock fault

HLF = Hollins Line fault

NY-AL lineament = New York-Alabama lineament

PMW = Pine Mountain window

SLF = Stonewall Line fault

TBF = Terrane Boundary fault

TCF = Talladega-Cartersville fault

TF = Towaliga fault

Introduction

The Late Paleozoic collision of Gondwana with Laurentia, the Appalachian orogenic episode, and the subsequent Mesozoic rifting that led to the opening of the modern Atlantic Ocean and the Gulf of Mexico are key tectonic events that shaped the southeastern U. S. The Paleozoic Alleghanian orogeny is the latest episode in the continent-continent collision of Laurentia and Gondwana (Hatcher, 2005). During the Mesozoic, Pangaea broke apart and plutons and dikes associated with rifting were emplaced nearly perpendicular to the strike of deformed metamorphic rocks. This Early Mesozoic intracratonic rifting created the graben system, known as South Georgia basin (Chowns and Williams, 1983; McBride and Nelson 1987; Thomas, 1988; McBride et al., 1989). Later, Triassic to Jurassic sediments filled the basin, overlapping the northern boundary of the Gondwana-related rocks and obscuring the Alleghanian suture. Crustal rocks containing the remnants of this continental suture zone are now covered by as much as 7 km of sediments of the Alabama Gulf Coastal Plain. Although regional-scale magnetic and gravity data exist, the nearest seismic profiles are from the 1980's COCORP campaign in adjacent Georgia (e.g., Cook et al., 1979, 1981; Nelson et al., 1985a, 1985b; McBride and Nelson, 1988, 1991). Thus, the detailed geophysical models of the deep crust beneath the Alabama Coastal Plain sediments are lacking.

The primary objective of this study is to develop new models of the crustal structure beneath the Alabama Gulf Coastal Plain using gravity and magnetic data along three transects that cross the postulated suture between Laurentian and Peri-Gondwanan/Gondwanan crust. Other objectives are to constrain the deep structure beneath major geological features in Alabama (e.g., Black Warrior Basin, Wiggins Arch, and the Brevard zone); to prepare a crustal model that explains major geophysical features (e.g., New York Alabama (NY-AL) lineament and Brunswick magnetic anomaly (BMA) in Alabama); and to compare the results with models derived from the COCORP transects in adjacent Georgia (e.g., Cook et al., 1979, 1981; McBride and Nelson, 1988, 1991; McBride et al., 2005).

This study aims to illuminate the details of the crust beneath Alabama that have not been revealed by drilling and seismic profiles. These details include the location of boundaries between different types of crust and the subsurface expression of major faults. In a broader context, the study will explore how models proposed for adjacent Georgia and other Gulf Coast margins relate to the crustal models derived for Alabama. Taken together, all these models provide insight into the tectonic processes that modified the southeastern margin of ancient Laurentia.

Background

Tectonic History

The southeastern continental margin and coastal plain of the United States found through two complete cycles of supercontinent formation and breakup (Pindell and Dewey, 1982; Hatcher, 1978, 1987; Salvador, 1991; Thomas, 1991). These cycles involved the two complete successions of Wilson cycles: (1) the assembly and break up of Rodinia and the opening of the Iapetus ocean; and (2) the closing of Iapetus ocean, the assembly of supercontinent Pangaea and its subsequent break up, and the opening of the modern Atlantic Ocean (Thomas, 2006). The second of the continental events resulted in the development of the Appalachian orogen. Three distinct major continent-continent collisions are recorded in the Appalachian orogenic belt: the Middle Ordovician Taconic orogeny, the Devonian Acadian orogeny, and the Pennsylvanian-Permian Alleghanian orogeny (Figure 1; See Hatcher, 2010 for summary). Deformation associated with the Alleghanian suture zone are covered by coastal plain sediments in the southern part of Alabama and Georgia.

The Taconic orogeny is the earliest of the Appalachian mountain building stages. This orogeny led to the amalgamation of Laurentia with an arc terrane (Drake et al., 1989; Horton et al., 1989). Although clastic sedimentary units associated with Taconic orogeny are found in part of the Valley and Ridge province of Alabama and Georgia,

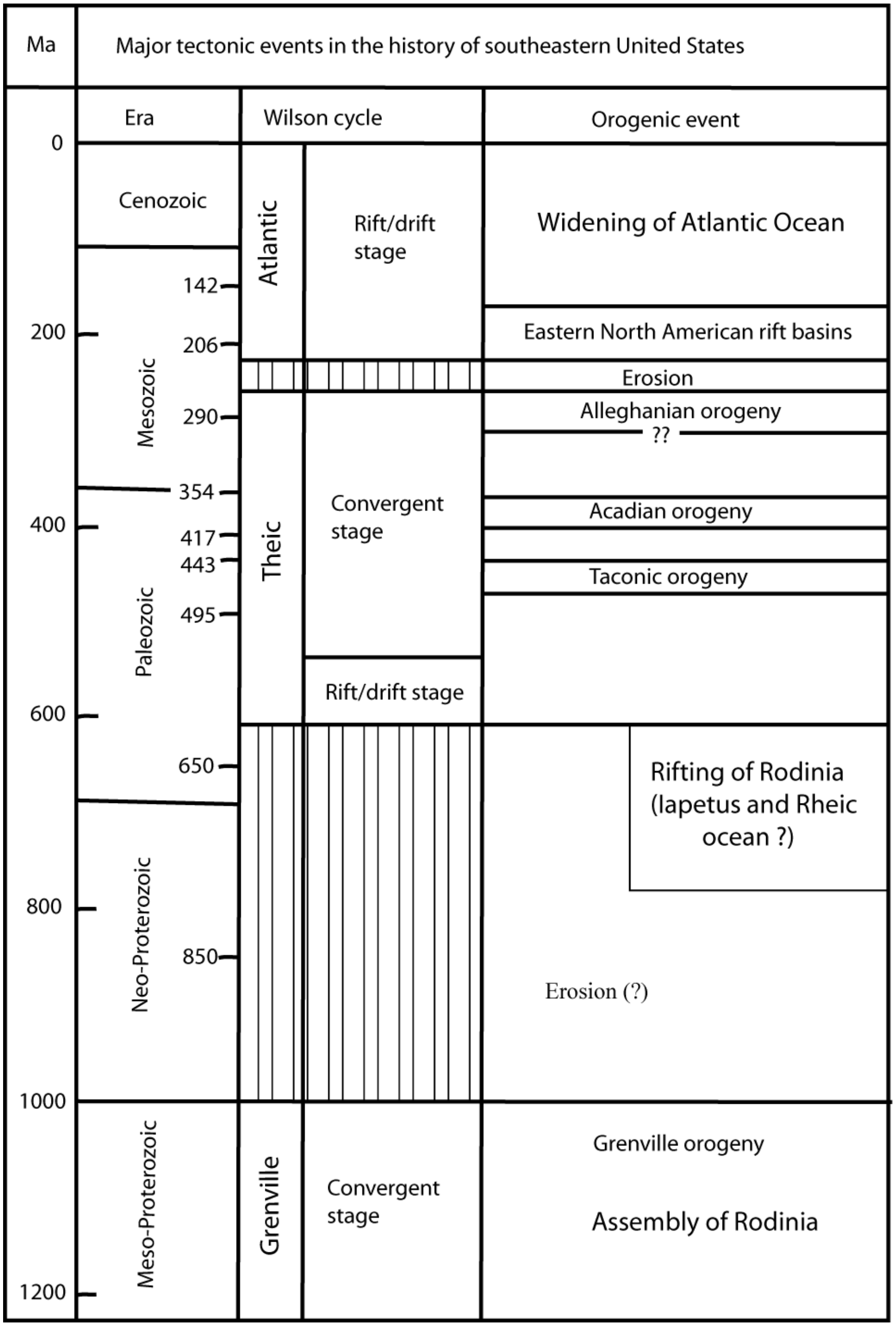


Figure 1. Major orogenic events forming the Appalachian mountain belt (modified after Hatcher, 1987; Faill, 1997).

evidence of the orogeny in Piedmont rocks is sparse (Steltenpohl, 2005). Eclogite and granulite facies Taconic rocks are found in New England, however, dominating the collision of the arc terrane (Hatcher, 2010).

The Late Silurian to Early Mississippian Acadian orogeny resulted in the accretion of Late Neoproterozoic to Cambrian volcanic arc terranes of the Peri-Gondwanan Carolina superterrane to Laurentia (Horton et al., 1989; Osberg et al., 1989; Steltenpohl et al., 2008; Hatcher, 2010). This orogenic event also added smaller terranes to parts of the northern Appalachians (Faill, 1997; Hatcher, 2007). Most of the Piedmont, including the Talladega Slate belt of central Alabama and Georgia, is thought to have experienced Acadian metamorphism as a result of this event (Tull, 1980; Glover et al., 1983; Hatcher, 2010).

The Alleghanian orogeny includes the closing of Iapetus ocean by the continent-continent collision of Laurentia with Gondwana, which aided in the consolidation of the supercontinent Pangaea (Hatcher, 2005, 2010). The Carboniferous to Early Permian Alleghanian orogeny reactivated some major faults, such as Brevard, Goat Rock and Bartlett’s Ferry faults in the Southern Piedmont. Some of faults were initially formed in the thrust-related transportation of the previously docked terranes of the Taconian and Acadian orogenies (Hatcher, 2005, 2010). Evidence of the Alleghanian orogeny is widespread throughout the central and southern Appalachians, but it is less evident in the northernmost maritime Appalachians (Hatcher et al., 1989).

During Mesozoic, Pangaea began to rift apart and Africa and South America began to drift away from North America south of the original Alleghanian suture. Mesozoic rifting created a NW-trending set of diabase dikes that cut across the

Appalachian structural grain at high angles. Rift basins later developed into the modern Atlantic Ocean and Gulf of Mexico. This Early Mesozoic intracratonic rifting created the largest graben system in the southeastern United States, the South Georgia basin. Triassic to Jurassic basin sediments onlap the northern limit of the Gondwanan crust and obscure the Alleghanian suture between it and Laurentia. Evidence for the Wiggins-Suwannee suture, and the Suwannee terrane and the Mesozoic rift basins is only found in drill cores and geophysical data.

Geologic Background

Based on topographic relief, rock type and geologic structure, Alabama is broadly divided into five major physiographic provinces. From north to south these provinces are the Interior Low Plateau, the Appalachian/Cumberland Plateau, the Valley and Ridge, the Piedmont, and the Gulf Coastal Plain (Figure 2) (Sapp and Emplaincourt, 1975). The Interior Low Plateau of Alabama contains mostly sedimentary rocks deposited during the Mississippian. The Appalachian/Cumberland Plateau, which includes the Black Warrior Basin, consists predominantly of Carboniferous coal-bearing siliciclastic sedimentary rocks. The Black Warrior Basin, a foreland basin, is bordered on the southwest by the northwest-striking Ouachita thrust belt and on the southeast by northeast-striking Appalachian thrust (Guthrie and Raymond, 1992; Thomas, 2004).

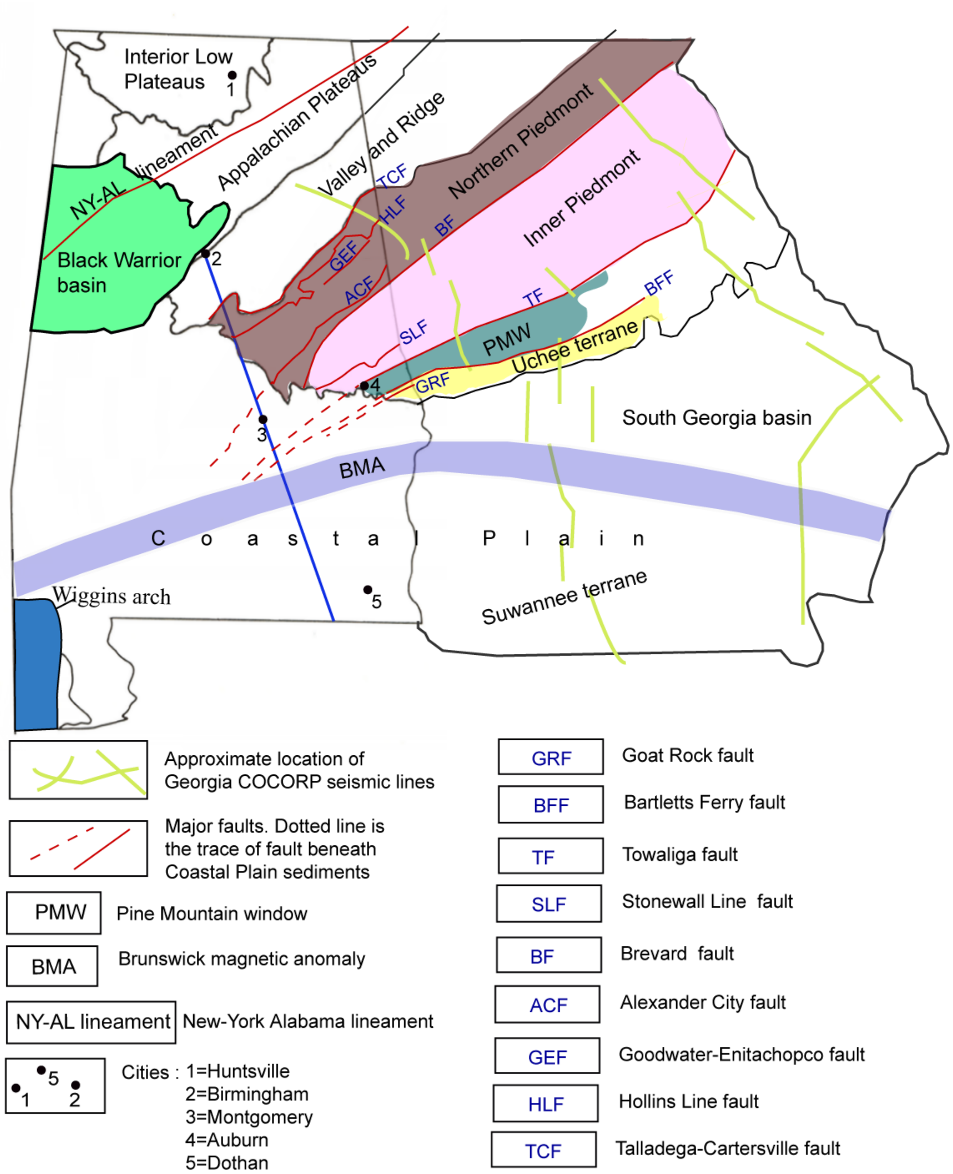


Figure 2. Map showing the major tectonic structures and physiographic provinces in Alabama and adjacent Georgia area (modified from Sapp and Emplaincourt, 1975; Hopson and Hatcher, 1988). Location of the Georgia COCORP seismic transects used by McBride and Nelson (1988, 1991). Red dotted line represents the tracing of faults beneath Coastal Plain sediment based on magnetic data (after Steltenpohl et al., submitted). Blue line from Gulf coast to Birmingham is the profile location of gravity modeling by Savrda (2008). Location of NY-AL lineament (after Steltenpohl et al., 2010).

The Valley and Ridge province consists of arrays of northeast and southwest trending ridges and valleys. Valleys are generally formed by shale and carbonate, whereas ridges consist of sandstone and chert (Raymond et al., 1988). Rocks of the Valley and Ridge are sedimentary in origin and were deposited during Cambrian to Pennsylvanian, then later were folded and faulted during the Late Paleozoic Appalachian collision between Laurentia and Gondwana. The thickness of preserved Paleozoic rocks in synclines, determined by the depth to the pre-Cambrian crystalline basement, is less than 7 km (Thomas, 1982; Neathery and Thomas, 1983; Raymond et al., 1988).

Piedmont Province

The Piedmont province of Alabama is structurally complex and is bounded by the Talladega-Cartersville fault system to the northwest and the Gulf/Atlantic Coastal Plain onlap to the southeast (Figure 2). The crystalline rocks of Piedmont province show an increase in metamorphic grade from northeast (low-grade greenschist facies) to southeast (high-grade amphibolite facies) (Tull, 1980; Glover, 1983). The Piedmont is divided into three lithotectonic provinces: the Northern Piedmont, the Inner Piedmont, and the Southern Piedmont (Raymond et al., 1988).

The Northern Piedmont is separated from the Valley and Ridge province by the Talladega-Cartersville fault system and from the Inner Piedmont by the Brevard fault zone. From northwest to southeast, the Northern Piedmont is structurally divided into Talladega, Coosa and Tallapoosa blocks, separated by the Hollins Line fault and the Goodwater-Enitachopco fault (Thomas and Neathery, 1980; Raymond et al., 1988). The Talladega block contains low-grade metasedimentary and metavolcanic rocks. Major lithologies of Talladega block are phyllite, marble, slate, and greenstone. Rocks of Coosa block are mostly metasedimentary (graphitic schist, garnet mica schist with quartzite), with a significant amount of amphibolite (Stow et al., 1984; Raymond et al., 1988). Amphibolite sequences are also common within the rocks of Coosa block. The southernmost block of northern Piedmont, the Tallapoosa block, is composed of high-grade metasedimentary and meta-plutonic rocks.

The Inner Piedmont is bounded by the Brevard fault zone in the north and the Towaliga fault in the south. The Brevard zone consists of highly sheared rocks. This shear zone is rooted in the main sole thrust beneath the Northern Piedmont and Inner Piedmont (Hatcher., 1971; Cook et al., 1979). The Inner Piedmont is divided into the Dadeville and the Opelika Complexes, separated by the Stonewall line fault (Bentley and Neathery, 1970; Osborne et al., 1988). The Dadeville Complex lies within a major synformal structure, the northeast-plunging Tallassee synform. The lithology is dominated by meta-igneous and metavolcanic rocks (amphibolite and granitic gneisses) (Steltenpohl et al., 1990a). Southeast of the Stonewall Line fault, the meta-sedimentary rocks (biotite gneiss, mica schist, quartzite, and rare amphibolite) of the Opelika complex are intruded by Ordovician granitic plutons (Steltenpohl et al., 1990a, Steltenpohl, 2005). Based on structure, the Inner Piedmont is interpreted as a group of several allochthonous blocks, which contain nappes or large thrust sheet (Hatcher, 1978; Hatcher and Williams, 1986).

The Southern Piedmont, separated from Inner Piedmont by Towaliga fault, consists of two blocks: the Pine Mountain window and the Uchee terrane. These blocks are separated by Bartletts Ferry/Goat Rock faults. The Pine Mountain window contains a metamorphosed basement complex and a younger cover sequence of quartzite, marble, and schist. The basement rocks of Pine Mountain window are dated at 1.1 Ga (Odom et al., 1973, 1985; Steltenpohl et al., 1990b, 2004).The southern boundary of Pine Mountain window is formed by Bartletts Ferry/ Goat Rock faults, which are prominent southeast-dipping mylonite zones up to 3 km wide. The Uchee terrane is the southernmost block of the Piedmont and consists of coarsely crystalline biotite gneiss with schist and amphibolite (Steltenpohl et al., 2008). To the south, the Appalachian rocks are covered by the younger sediments of Coastal Plain.

Coastal Plain Sediments

Sedimentary rock units of Coastal Plain strike east-west in the eastern part of Alabama and it changes to northwest then north in western Alabama. The thickness of this sedimentary rock package increases from 15 to 300 m in northwestern Alabama to more than 7 km near the coast (Raymond et al., 1988). The rocks range in age from Late Triassic to Early Cretaceous (Figure 3).

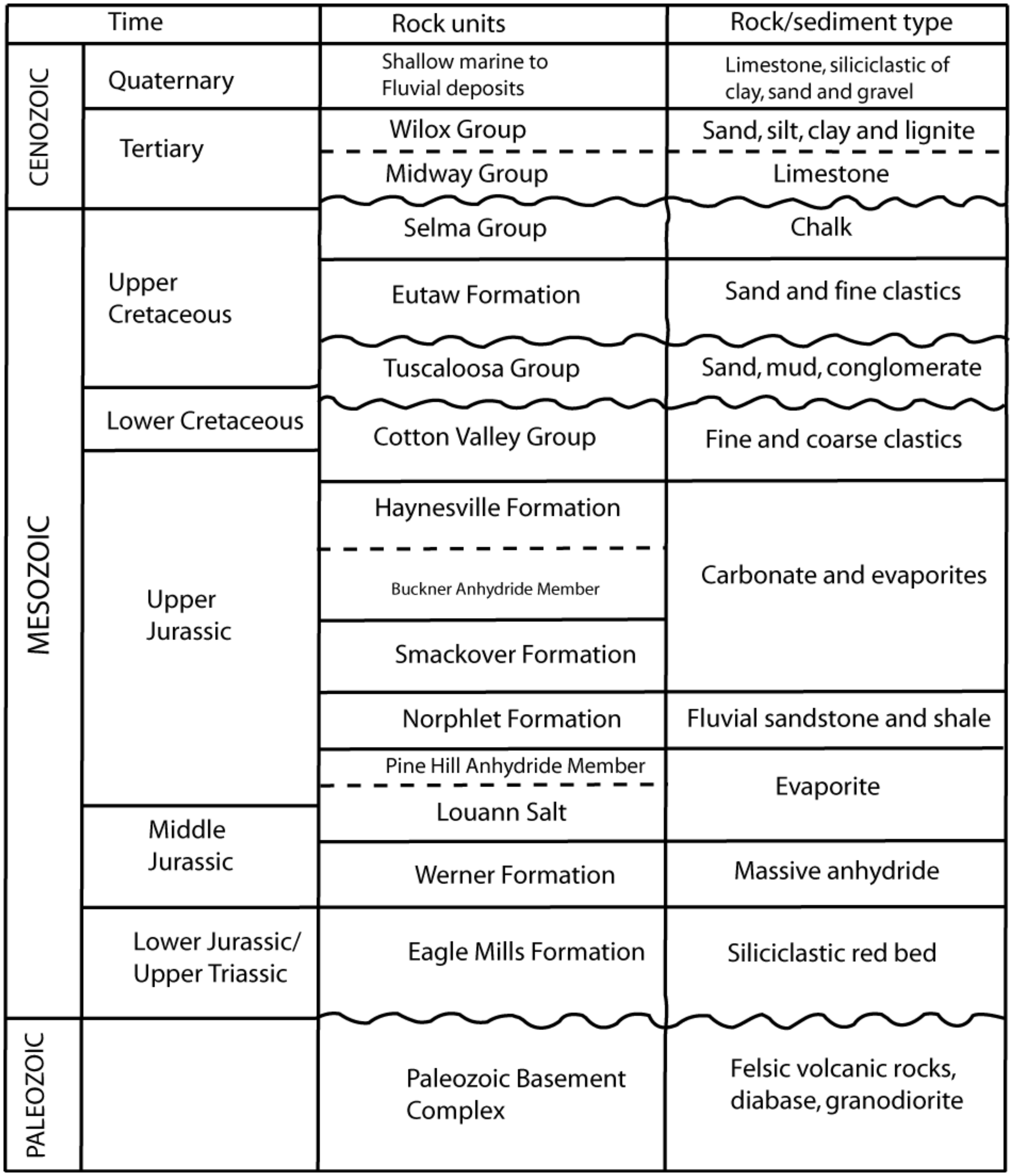


Figure 3. Generalized stratigraphic column of Coastal Plain sediments (modified after Tew et al., 1993; Geological Survey of Alabama, 2006)

Rocks Beneath the Coastal Plain Sediments

Sedimentary rocks of South Georgia basin and Suwannee terrane provide basement for the Coastal Plain sediment. Triassic to Lower Jurassic rocks of South Georgia basin are described as sandstone facies (sandstone, conglomerate, schist with clastic fragment of granite, gneiss, basalt and rhyolite) and mudstone facies (mudstone, shale with rhyolite and tuff clasts) (Chowns and Williams, 1983; Daniels et al., 1983; Guthrie and Raymond et al., 1992). Suwannee terrane rocks have a Gondwanan origin based on faunal assemblages found in drill cores (Applin, 1951; Barnett, 1975; Pojeta et al., 1976; Wilson, 1966). The rocks associated with Suwannee terrane in southern Alabama are divided into three units from bottom to top: (1) Paleozoic sedimentary rock; (2) granodiorite; and (3) a felsic volcanic units containing rhyolite, trachyte, felsic tuff (Dallmeyer, 1989b; Thomas et al., 1989; Guthrie and Raymond, 1992). Based on similar lithological and paleontological properties in the rocks of Suwannee terrane in Georgia and Florida, the Paleozoic sedimentary rock within the Suwannee terrane of Alabama is interpreted to unconformably overlie the felsic volcanic rock and high-grade metamorphic rocks (Chowns and Williams, 1983; Dallmeyer, 1989b; Thomas et al., 1989).

Also beneath the Coastal Plain sediments is the Wiggins Arch, a tectonic terrane located in southernmost Alabama and Mississippi. The Wiggins Arch in Alabama consists of low-grade metamorphic rocks. Three different, low-grade metamorphic rock units (chlorite-sericite-quartz phyllite, sericite-quartz phyllonite, and granite mylonite) were identified from the wells that penetrated Wiggins terrane in southern Alabama (Guthrie and Raymond, 1992).

Major Geophysical Anomalies in Alabama

Several of the magnetic terranes discussed by Higgins and Zietz (1983) are present in Alabama. A prominent magnetic low sub parallel to perpendicular to the East Coast magnetic anomaly is the BMA (Figures 2, 4 and 6) (Pickering et al., 1977; Klitgord and Behrendt, 1979; Daniels et al., 1983). This magnetic anomaly is same as the Altamaha anomaly of Higgins and Zietz (1983). COCORP seismic profiles, acquired in 1983 in Georgia, support the hypothesis that BMA characterizes a Late Paleozoic Suwannee suture rather than Mesozoic rift basins (Nelson et al., 1985b).

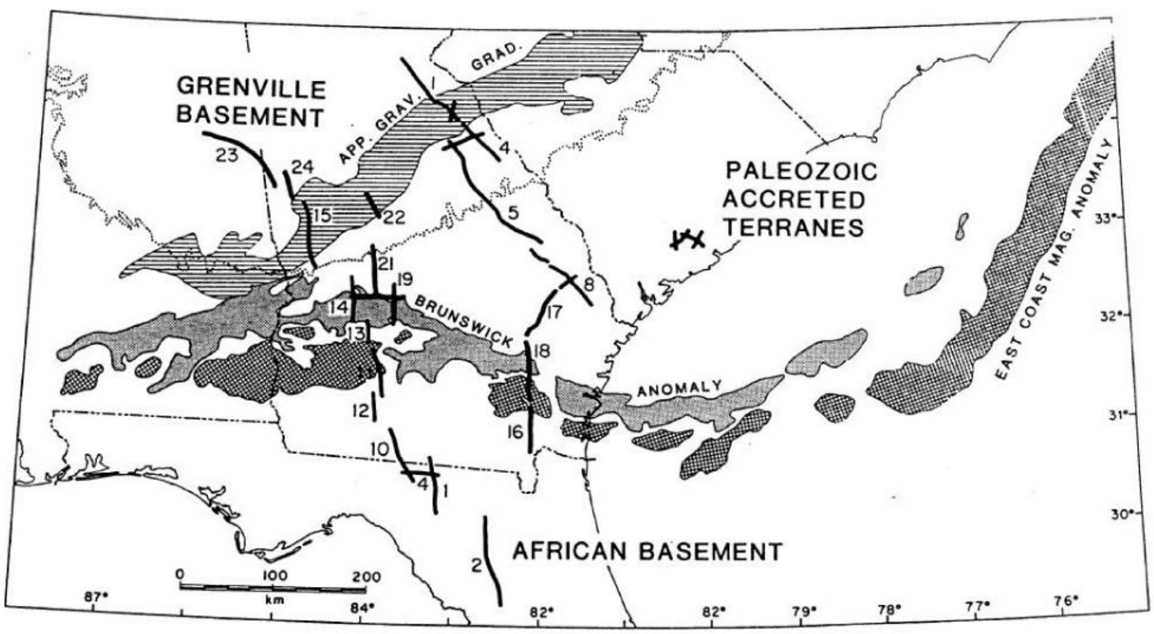


Figure 4. Location map showing COCORP survey lines, crustal terranes and magnetic anomalies (after McBride and Nelson, 1991).

Another prominent magnetic anomaly observed in Alabama is the southern end of the northeast trending NY-AL lineament (Figures 2, 5, 6 and 7). Steltenpohl and others (2010) suggest that this geophysical anomaly delineates a major crustal boundary beneath the Appalachian basin. In northern Alabama, this lineament separates a magnetic-high/gravity-low southeastern block from a magnetic-low/gravity-high northwestern block (Figures 7A and 7B). The sources of this magnetic anomaly are not exposed and no drilling has revealed its source (Steltenpohl et al., 2010).

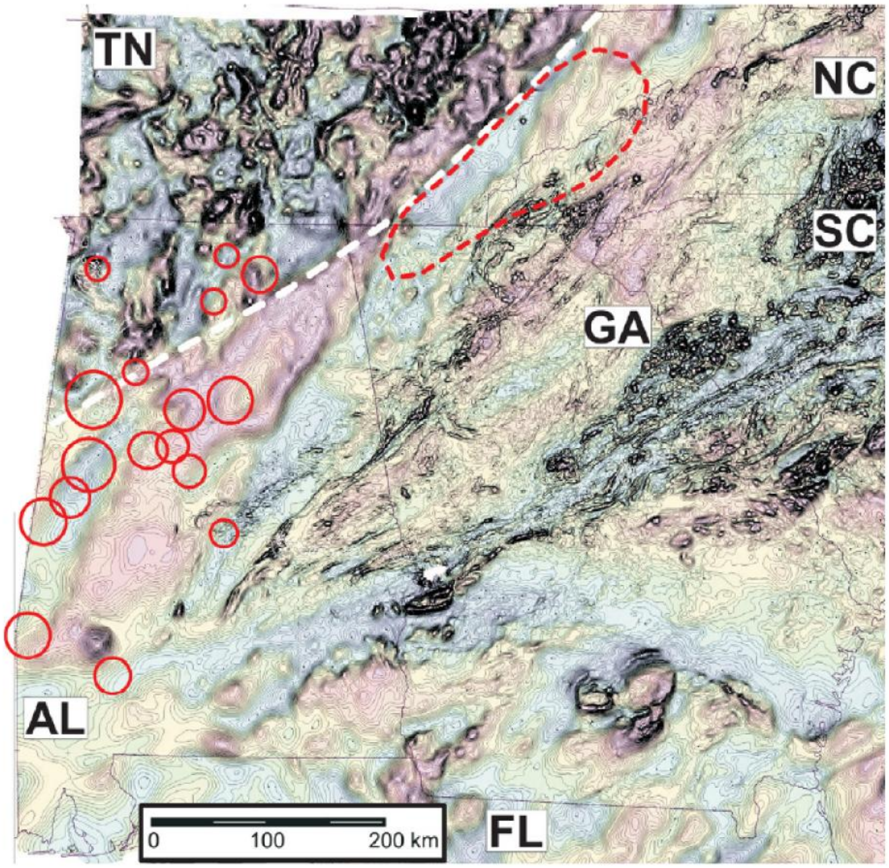


Figure 5. Location of NY-AL lineament (white dashed line) on contoured magnetic map of several states (after Steltenpohl et al., 2010). State abbreviations AL-Alabama, TN-Tennessee, FL-Florida, GA-Georgia, SC-South Carolina, and NC-North Carolina.

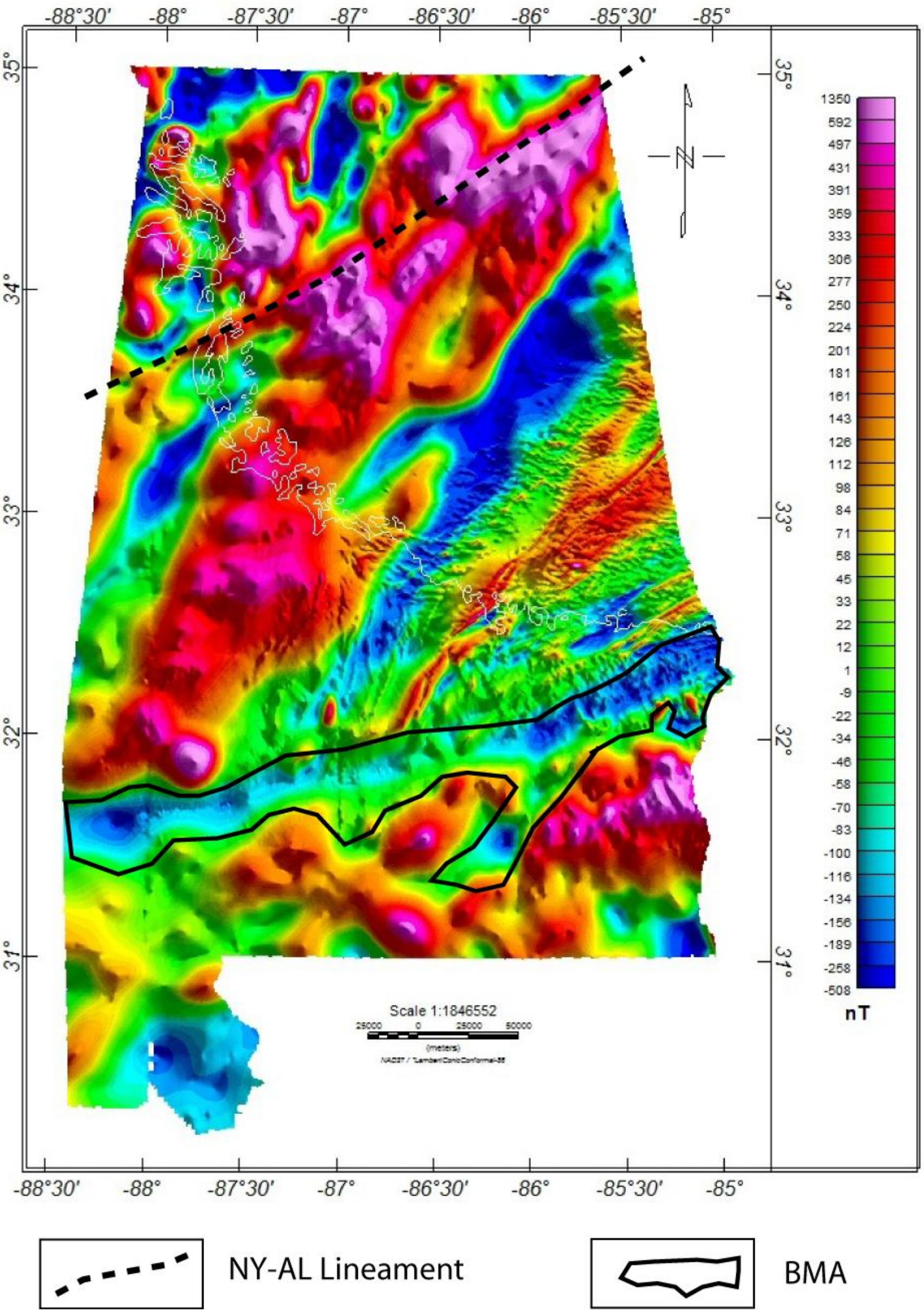


Figure 6. Approximate location of BMA and NY-AL lineament on magnetic map of Alabama. Magnetic map (after D. Daniels pers. communication, 2007). Location of NY-AL lineament (after Steltenpohl et al., 2010). Location of BMA (after McBride and Nelson, 1991).

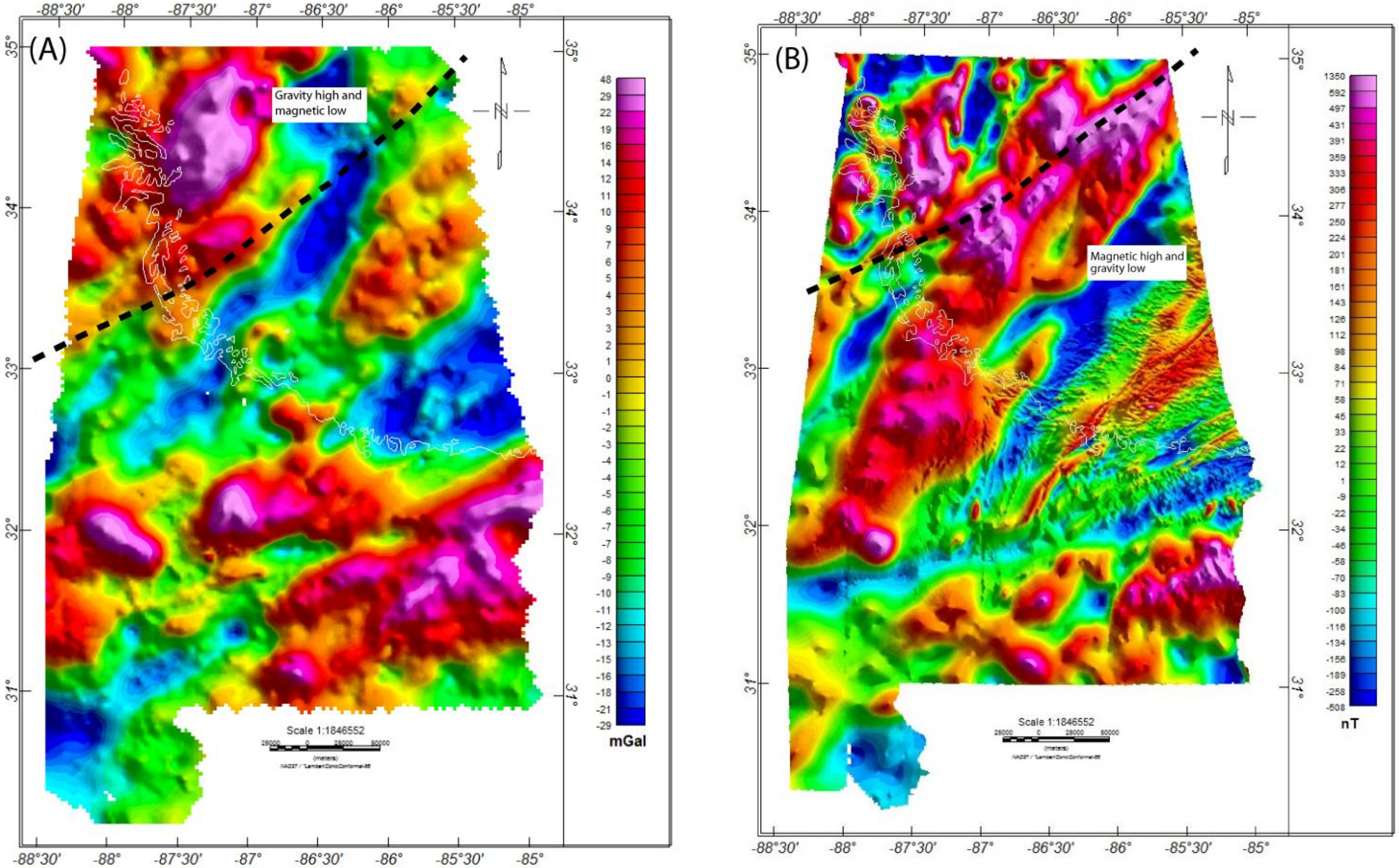


Figure 7. Approximate location of NY-AL lineament on (A) gravity map of Alabama, and (B) magnetic map of Alabama. In Alabama, the lineament separates a gravity high and magnetic low northeastern block from magnetic high and gravity low southeastern block. Gravity and magnetic map (after D. Daniels pers. communication, 2007). Location of NY-AL lineament (after Steltenpohl et al., 2010).

Previous Work

Little work has been aimed at developing a deep crustal model beneath Alabama; however, some details of crustal structure beneath Mississippi and Georgia are available. Harry and Londono (2004) used forward gravity modeling to explore the crust beneath Mississippi. They suggested that the load on the southern Laurentian margin produced flexural subsidence in the Black Warrior Basin during Late Paleozoic (Harry and Londono, 2004). The Consortium for Continental Reflection Profiling (COCORP) project has also provided insight into the crustal structure beneath Georgia through the acquisition of deep seismic data.

Depth of Moho

Estimation of total thickness of the crust plays an important role for constructing crustal scale gravity models because of the large density difference between crust and mantle rock. In the seismic reflection data from the Georgia COCORP profile, the deepest reflections seen are interpreted as the Moho (Nelson et al., 1985a). McBride and Nelson (1991) identified depth to the Moho in the southern Appalachians and reported that crustal thickness is greater in northern Alabama/Georgia than in the south (Figure 8). They estimate that the thickness of crust is about 37 km beneath the South Georgia basin and the Suwannee terrane, and it reaches about 43 km beneath Brevard zone

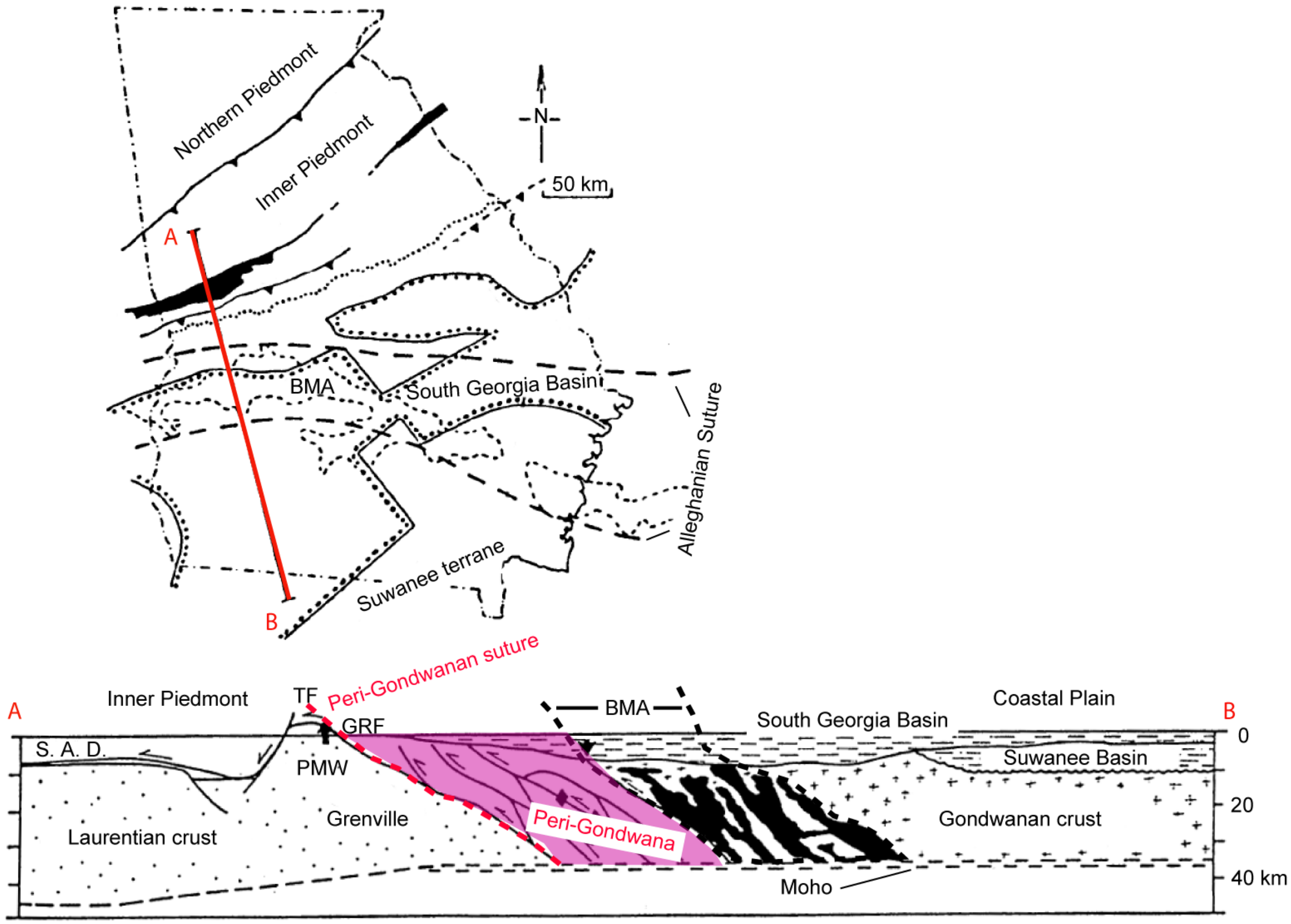


Figure 8. Schematic geologic interpretation of southern Appalachians from COCORP seismic reflection data (after McBride and Nelson, 1991). The cross section AB (north to south) shows the depth of Moho is greater in the north. Abbreviations: S.A.D. = Southern Appalachian Decollement, TF=Towaliga Fault, GRF= Goat Rock fault, PMW=Pine Mountain window. Peri-Gondwanan rocks (purple) and suture (red dashes) are modifications after Steltenpohl et al. (2008).

(McBride and Nelson, 1988, 1991). They observed that Moho depth is shallow near the suture zone compared to the Piedmont.

Southern Appalachian Subsurface Geology

Many researchers have worked on the surface and subsurface rocks and structures in the southern Appalachian and it is well known that the Grenville basement rocks beneath the Plateaus, the Valley and Ridge, and most of the Piedmont have Laurentian origin (e. g., Hatcher, 1972, 2004; Neathery and Thomas, 1976; Cook et al., 1979, 1981; Nelson et al., 1985a, 1985b; McBride and Nelson, 1988, 1991; Guthrie and Raymond, 1992; Thomas, 2006; Steltenpohl et al., 2008, 2010). From the detailed study of the COCORP seismic survey line that crosses the Brevard zone, Cook and others (1979) postulated that a relatively horizontal, layered sequence of sedimentary rock underlies a 6 to 15 km-thick section of Piedmont rocks. These interpretations are similar to the suggestion of Hatcher (1971, 1978) and Clark and others (1978) that the sedimentary rocks of the Valley and Ridge extend beneath the Brevard zone. Cook and others (1979) also suggested that the Brevard fault does not extend beyond 6 to 9 km depth (2 to 3 s TWT). They estimated offsets on the Brevard fault to be about 8 km in throw and 30 km in heave.

The Pine Mountain window exposes the Grenville basement and its metasedimentary cover at the Earth’s surface (Clarke, 1952; Hooper et al., 1988; Steltenpohl et al., 2004). Nelson and others (1987) interpreted a prominent reflection in COCORP data to be the southern Appalachian detachment. They noted that the reflector was not continuous to the southeast beneath Pine Mountain belt. These terminated reflections north of the Pine Mountain window suggest that the Piedmont detachment does not continue beneath the window. They also found that the Towaliga fault, which separates Pine Mountain window from Inner Piedmont, has average dip of 54° and offsets Grenville basement about 9 km (Figure 9).

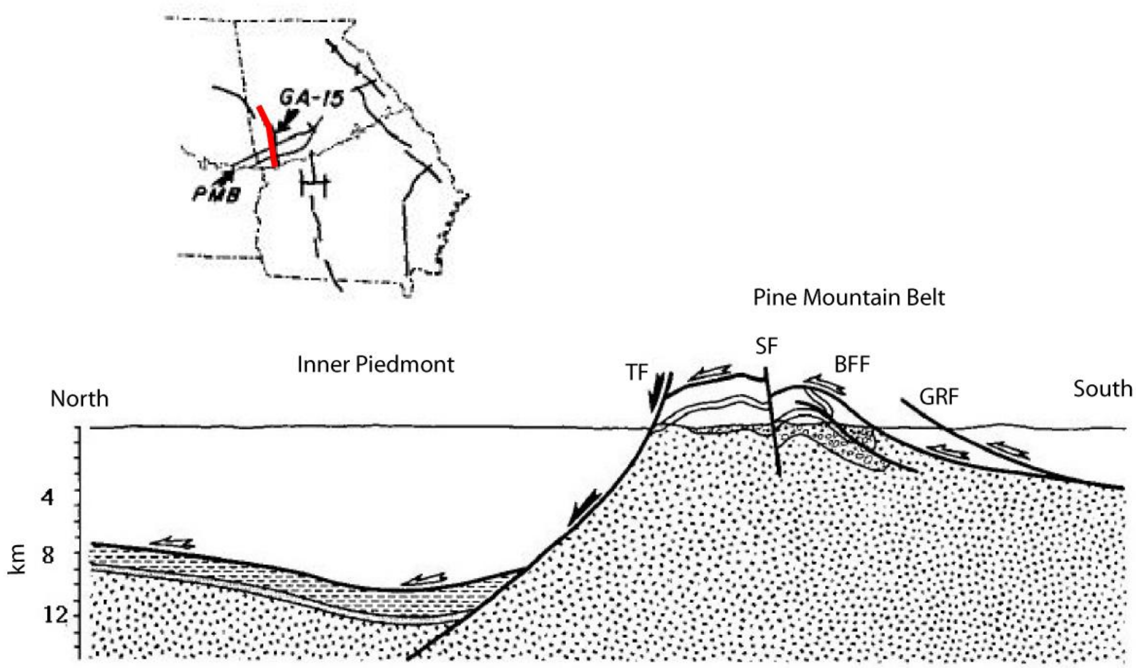


Figure 9. Cross-section showing geologic interpretation of Georgia COCORP line-15 (red line) (modified after Nelson et al., 1987). About 9 km of normal-slip offset along Towaliga fault is seen in the cross section. Abbreviations: TF = Towaliga fault, SF= Shiloh fault, BFF= Bartletts Ferry fault, and GRF= Goat Rock fault.

Origin of Rocks Beneath Coastal Plain

Lower Jurassic rocks beneath the Coastal Plain sediments are divided into three different affinities: Laurentian crust (Chowns and Williams, 1983; Thomas et al., 1989); the Gondwanan Suwannee terrane; and the Wiggins terrane. Wilson (1966) recognized lower to middle Paleozoic strata in cores drilled through the Coastal Plain in northern

Florida and developed the hypothesis that northern Florida was underlain by a fragment of Gondwanan crust that was sutured onto Laurentian crust during late Paleozoic. Even though Wilson's hypothesis was generally accepted by later researchers (e.g., Neathery and Thomas, 1975; Pindell and Dewey, 1982; William and Hatcher, 1983; Chowns and Williams, 1983; Guthrie and Raymond, 1992), the exact location of the Paleozoic suture was enigmatic until the COCORP data were acquired in the mid-1980's. The Georgia COCORP profiles served as a main source for information on the crustal configuration of southeastern United States. In their analysis of Georgia COCORP data McBride and Nelson (1988) associated the Alleghanian suture zone with a series of prominent southeast-dipping reflectors.

Laurentian crust beneath the Coastal Plain includes the southwestern continuation of the Black Warrior basin, the Valley and Ridge, and the Piedmont rocks. The Valley and Ridge and the Piedmont provinces extend 50-60 km southeast of Coastal Plain unconformity, as revealed by deep wells (Neathery and Thomas, 1976; Dallmeyer, 1987; Guthrie and Raymond, 1992). The rock units and fault zones in the Southern Piedmont can also be traced beneath the Alabama Coastal Plain using magnetic data because they are associated with a sharp magnetic gradient (Neathery et al., 1976; Horton et al., 1984; Steltenpohl et al., submitted).

The origin of the Wiggins terrane is not clear. Thomas and others (1989) proposed that the rocks of the Wiggins terrane may be (1) metamorphosed equivalents of rocks in the Suwannee terrane, or (2) amalgamation of the Suwannee terrane with another terrane, or (3) an entirely different terrane. Dallmeyer (1989a) noted that the $^{40}Ar/^{39}Ar$ biotite cooling ages from basement rocks of the Wiggins Arch range from 262 to 320 Ma, similar to the rocks in the Piedmont. He suggested that the Wiggins Arch originated in the Piedmont and later was transported to its present position during the Late Paleozoic Alleghanian orogeny.

Although studies have provided unprecedented insight into the deep structure of Georgia, details of how crustal geometries continue beneath the Alabama Coastal Plain is lacking. Savrda (2008) reconciled the crustal models of Alabama Gulf Coastal Plain established by Wilson (1966) with updated geological and geophysical data. She constructed a crustal model using gravity data along a single profile from the Gulf of Mexico to Birmingham. In her model, she delineated the possible location of the suture between continental crust of ancient North America (Laurentian) and Peri-Gondwana rocks is Bartletts Ferry/Goat Rock fault as reported by Steltenpohl and others (2008) (Figure 10). Even though this work represents an important first step, fully integrated gravity and magnetic models for Alabama have not yet been formulated. This project uses gravity and magnetic data to construct crustal models for Alabama.

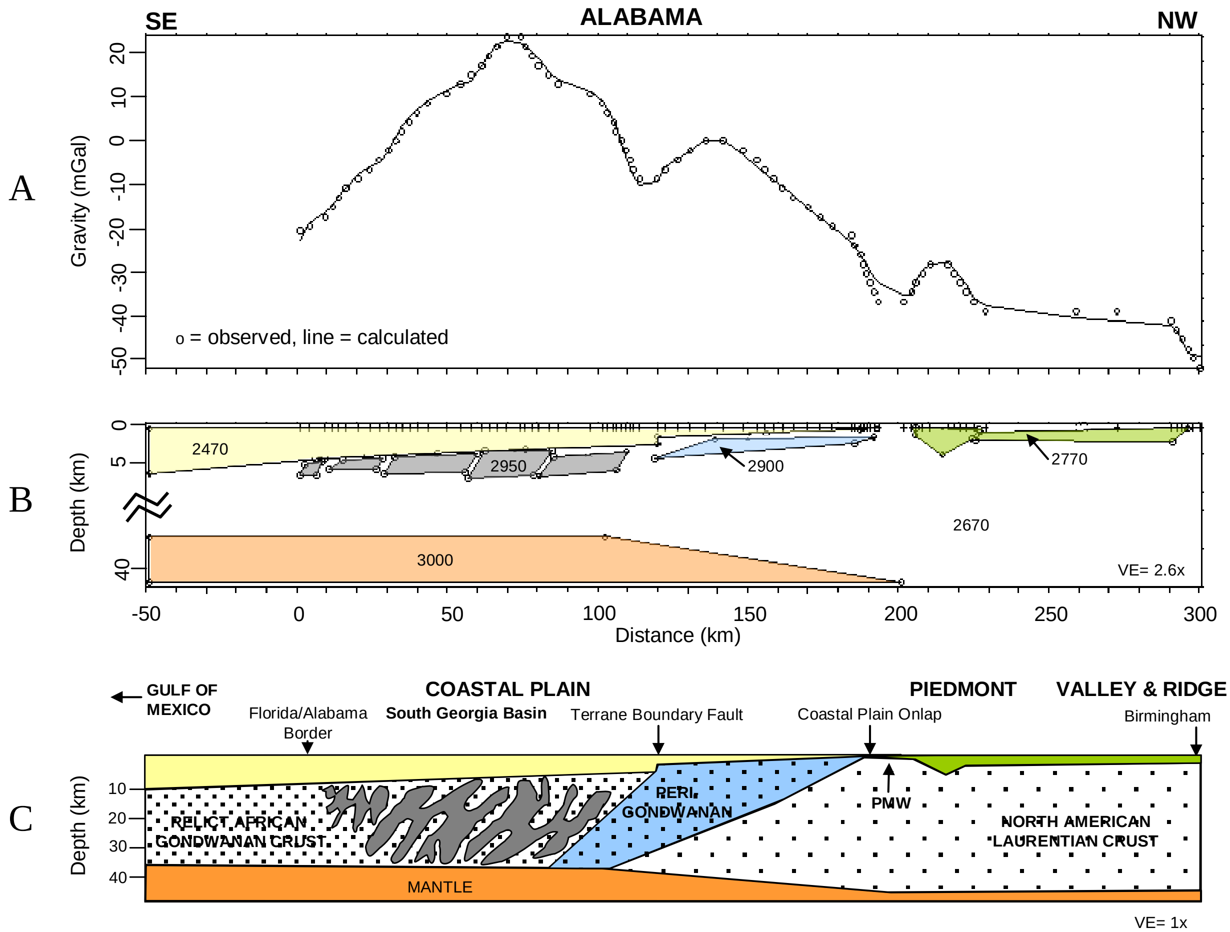


Figure 10. (A) Gravity data (observed and calculated) along profile from the Gulf of Mexico to Birmingham, Alabama. (B) Density cross-section. (C) Geologic interpretation of density model shown in B (After Savrda, 2008). See figure 2 for profile location.

Methodology

This section explains the details of the data collection and the data processing methods of the study. Data collection involved obtaining existing gravity and magnetic data, gathering well-log data, and accumulating other supporting data. Data processing included analyzing gravity data, magnetic data, and other ancillary data.

Data Collection

This study uses Bouguer-corrected gravity data (Figure 11) and magnetic data (Figure 12) collected by the Defense Mapping Agency. These data are accessible from the U. S. Geological Survey and National Geophysical Data Center (http://irpsrvgis00.utep.edu/repositorywebsite/). Both the magnetic and the gravity data have been reprocessed at a 2500-m grid interval (D. Daniels, pers. communication, 2007) to allow better correlation with surface features. The original magnetic map (Godson, 1986) was created by using digitized contours of the Composite magnetic anomaly map of the United States (U. S. Geological Survey, 1982). Other data used in this study are from well logs, geological cross sections, and published literature.

Gravity Data

Measured gravity on the surface of the Earth must be corrected in many ways before an anomaly map can be produced. Free-air correction accounts for the variation in elevation of different gravity stations and a Bouguer correction accounts for deficits or

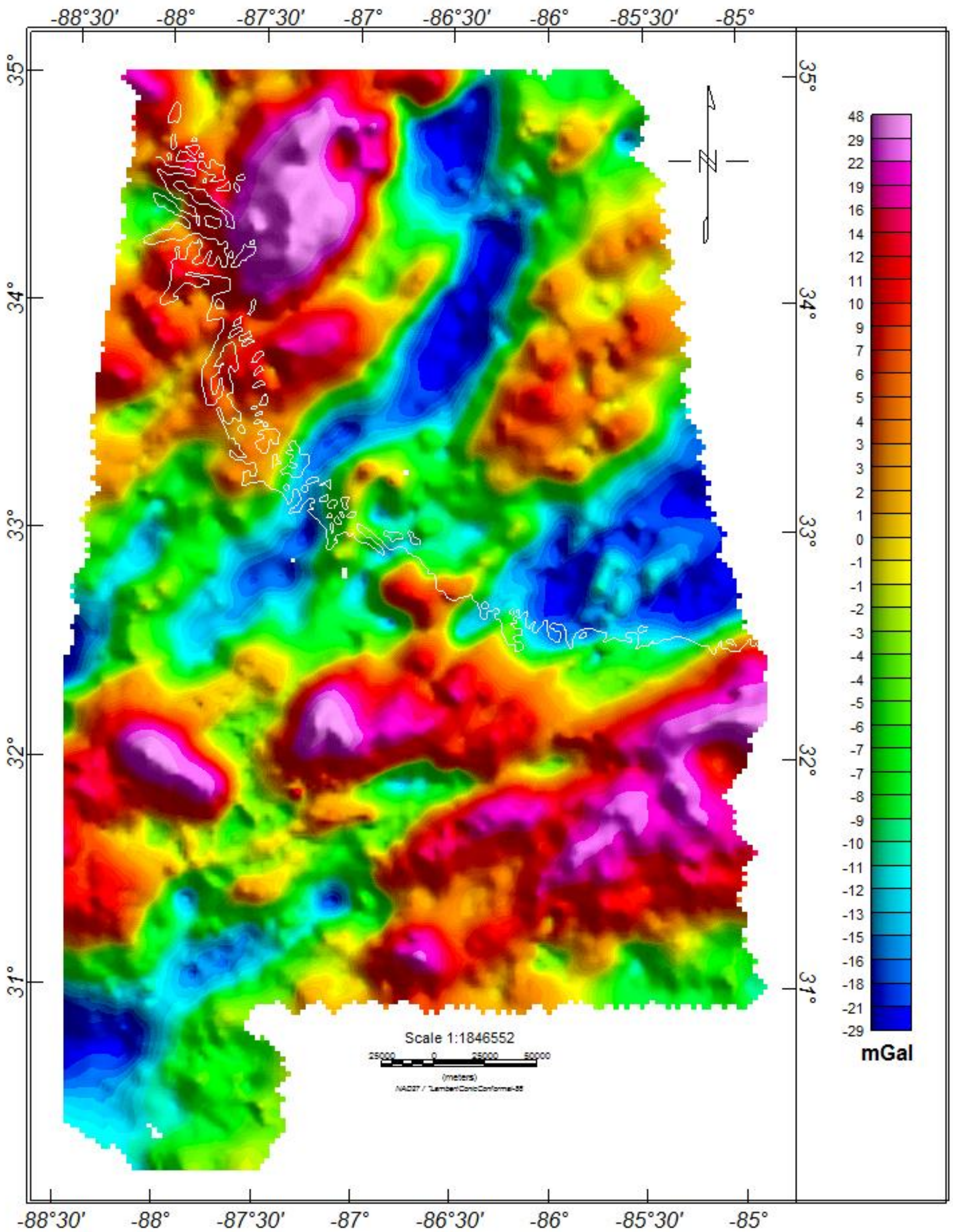


Figure 11. Bouguer-corrected gravity map of Alabama. Original data are reprocessed with a 2500-m grid interval to allow better correlation with surface features (Godson and Scheibe, 1982; Godson, 1985; D. Daniels pers. communication, 2007). Fall line is shown in white.

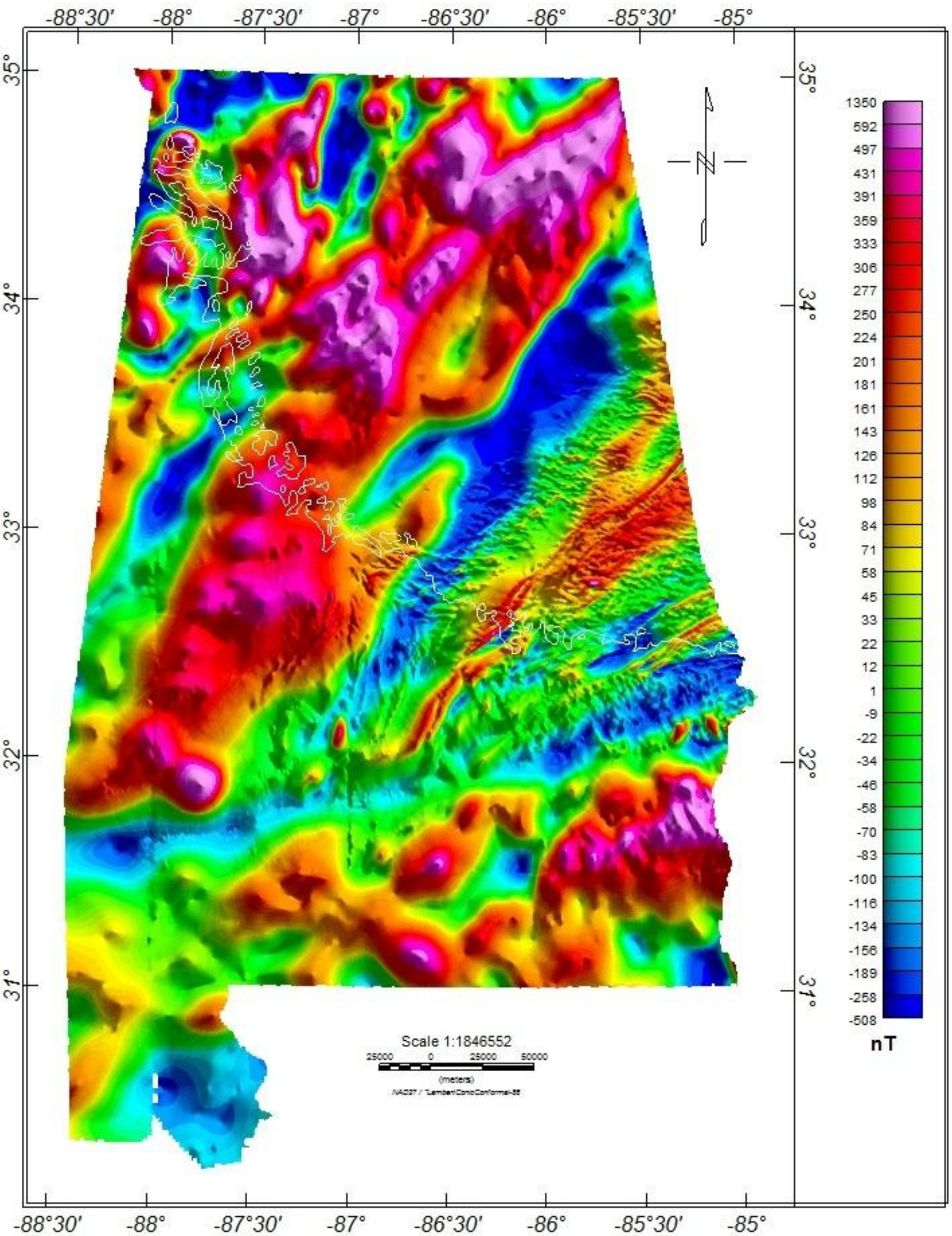


Figure 12. Magnetic map of Alabama. Original data are reprocessed using 2500-m grid interval (U. S. Geological survey, 1982; Godson, 1986; D. Daniels pers. communication, 2007). Fall line is shown in white.

excess mass between the measuring station and sea level. The Bouguer-corrected gravity data used in this study assume a reduction density of 2.67 g/cm^3. Terrain corrections have been applied in the areas of steep elevation gradients (Phillips et al., 1993).

Since the Bouguer gravity anomaly correlates with the lateral variation of density of the crustal rocks, a positive or a negative anomaly is created whenever there is a change in rock density (Phillips et al., 1993). In this study, an abrupt density change or high gravity gradient is used to delineate the boundaries between different tectonic terranes and rock units.

Magnetic Data

Magnetic anomalies usually result from the distribution of magnetic minerals, such as magnetite, hematite and pyrrhotite. Sedimentary rocks are typically nonmagnetic, but igneous and metamorphic rocks can be strongly magnetic. In Alabama, the Alexander City fault zone, the Brevard fault zone and the Bartletts Ferry/Goat Rock fault zones correlate with strong magnetic gradients. The magnetic map of Alabama was used in this study to delineate major surface geologic features, such as faults, and infer their significance for the tectonic history of southeastern United States.

Other Data

In addition to gravity and magnetic data, other supporting data, such as well-log data, geologic maps, COCORP seismic data from adjacent Georgia, and empirical data on the density and magnetic susceptibility of different rock types were assembled for the current study. Supporting data are an integral part of this study, as these data were used to help characterize the study area for gravity and magnetic modeling. Well-log data were used to constrain the lithologies of the upper few kilometers, and seismic data were used to constrain the deeper crust. Geologic maps were utilized to locate major geological boundaries along the modeled profile lines. Physical properties of the subsurface rocks were used to produce the gravity and magnetic starting models. Sources of data included published data sets, published literature, and personal communications (Table 1).

Table 1. Types of data and their sources

Type of data	Data Source
Bouguer gravity anomaly grid (.GRD)	Personal communication (Daniels, 2007)
magnetic data grid (.GRD)	Personal communication (Daniels, 2007)
Well-log information	Neathery and Thomas (1975); Guthrie and Raymond (1992)
Seismic reflection data	Cook et al., 1979, 1981); Nelson et al. (1985a,b); McBride and Nelson (1988, 1991); McBride et al. (2005)
Geologic cross section	Guthrie and Raymond (1992); Steltenpohl (2008)
Density and magnetic susceptibility of rocks	Telford et al. (1990); Dobrin and Savit (1988)

Data Processing

Gravity and magnetic data were digitally processed to prepare the data for modeling. Geosoft's Oasis Montaj software package was used for data processing. This software has different options for gridding and filtering of gravity and magnetic data. Oasis Montaj provides dynamic linking between images, maps, profiles, and graphs. Geosoft's GM-SYS software was used for profile modeling. GM-SYS is an interactive gravity and magnetic profile modeling software, which provides some constraints for modeling variables such as depth, density and magnetic susceptibility of rock bodies.

Gravity Data Processing

For the purpose of this study, the gravity data were processed using the Upward Continuation Residual (UCR) filter of the Oasis Montaj software to simulate the measurement surface at a higher level (in this case 40 km) above the Earth's surface. The UCR method provides frequency separation in potential field data that is more geologically interpretable than a fixed frequency or band-pass filter (Jacobson, 1987). This essentially low-pass (low spatial frequency) filter attenuates high spatial frequencies, including noise. The UCR filter transfers the anomalies measured on one surface to ones that would have been measured on a parallel surface (Blakely, 1995).

To perform the UCR filtering, the gravity grid was first prepared for Fast Fourier Transformation (FFT). In this process the existing grid was expanded to be square and dummy areas were replaced by interpolated values so that grid becomes smoothly periodic (Figure 13). This smooth square grid was transferred to the frequency (wave number) domain using a FFT. The next step was to apply UCR filter to the transformed grid. The upward continuation of 40 km was then applied to the wavenumber grid. Finally, the wavenumber grid was transformed to the space domain, and the grid size was reduced to its original size (G_2, Figure 14). It is necessary to subtract the upward continued grid from original grid to access the useful high frequency signal. Grid Math is used to do this calculation.

Resultant grid $\mathbf{(G_0) = G_1\text{-}G_2}$

Where,

G_0 = High-frequency gravity residual

G_1 = Original gravity

G_2 = Upward continued gravity

The resultant grid (G_0, Figure 14) effectively measures the anomalies from shallower sources by removing the dominant effect of deeper crust. Since basins and upper crustal density features are boosted in this grid, it is used for upper crustal modeling in this study. Similarly, the upward continued gravity data captures the anomalies that can be associated with long-wavelength features and are used to study large-scale characteristics of the crust.

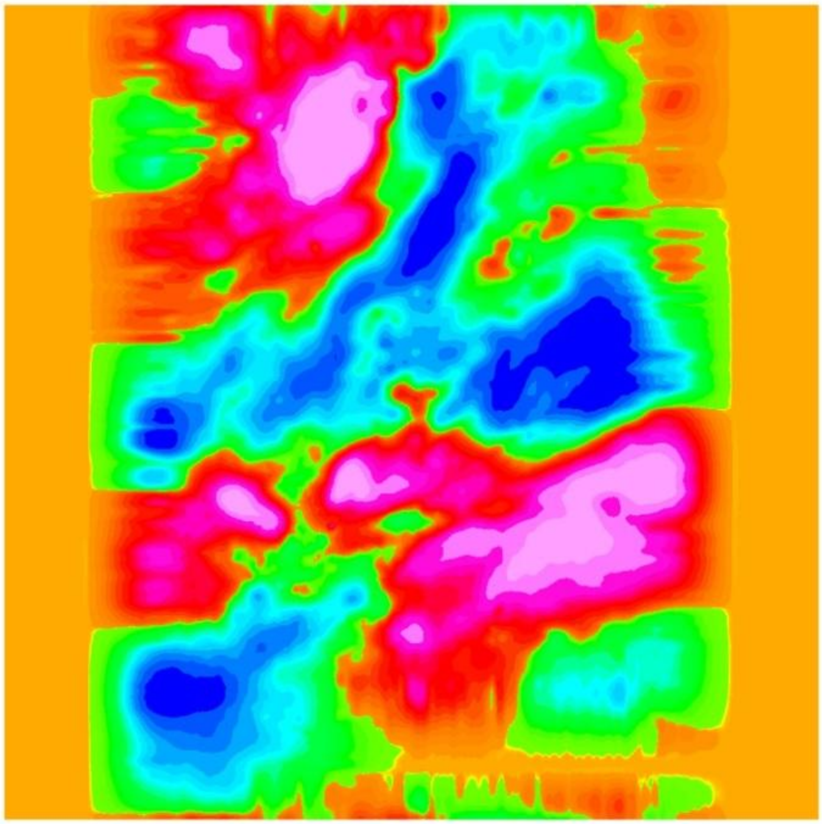


Figure 13. Gravity grid used for Fast Fourier Transformation (FFT). This grid is obtained after original grid (G_1) is expanded to be square and dummy areas are replaced by interpolated values.

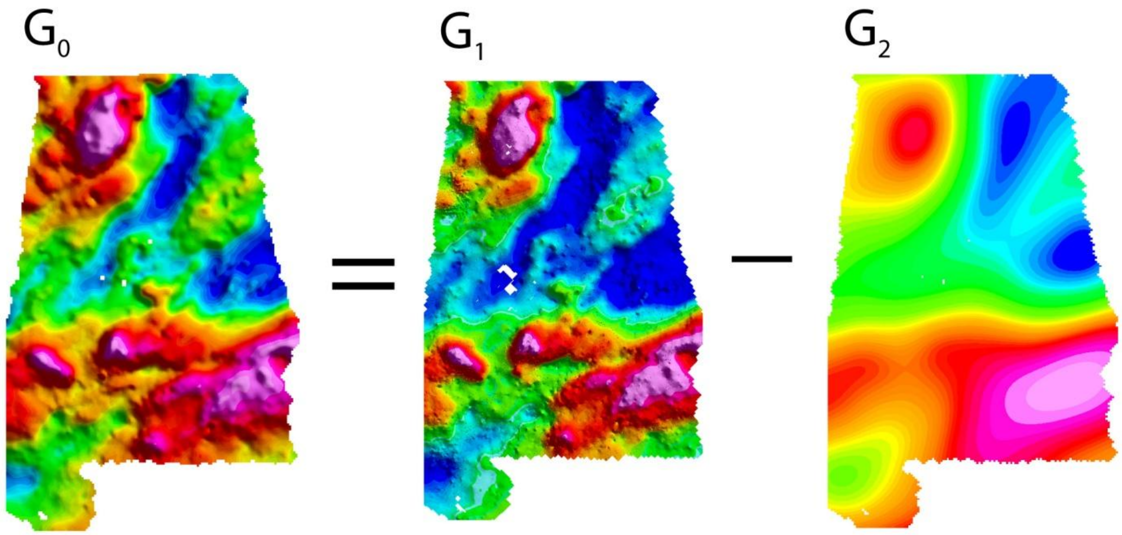
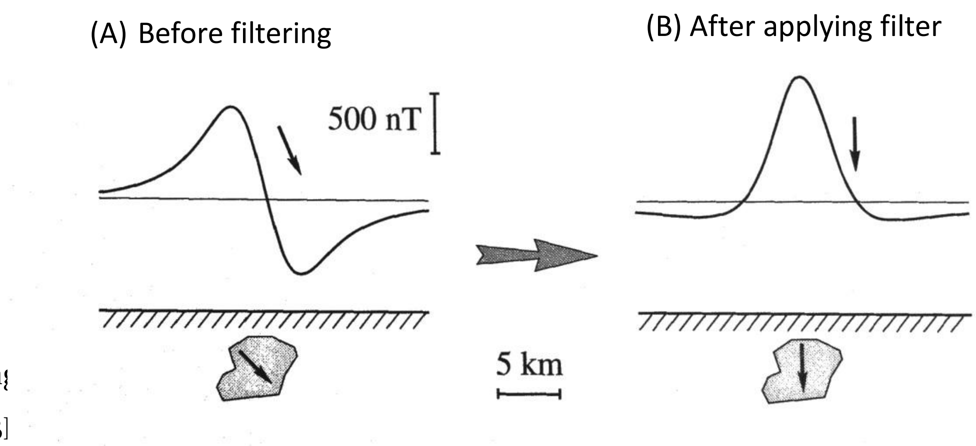


Figure 14. Gravity data processing with UCR filter. G_0 is the resultant grid obtained by subtracting the upward continued grid (G_2) from the original data (G_1).

Magnetic Data Processing

If both of the magnetization and the ambient field are not vertical, which is true in our case, symmetrical magnetic substances will produce asymmetrical anomalies (Blakely, 1995). To overcome this effect, magnetic data were processed for reduction to the magnetic pole (D. Daniels, pers. communication, 2007). This processing step removes the inclination effect from total-field magnetic anomalies by transforming the total field anomaly into the vertical component of the field produced as if the source were at the North magnetic pole (90° inclination).This step assumes that magnetization vectors are roughly parallel to the Earth's ambient field direction. The filter removes a major interpretation complexity because after magnetic data are reduced to pole, symmetrical bodies will produce symmetrical anomalies (Figure 15). The procedure for reduction to magnetic pole filtering is similar to the UCR filtering in gravity. Inclination and declination of the center part of the study area was calculated using the IGRF.omn subroutine in Oasis montaj and this value was used for the reduction to pole.



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Auxiliary Data Processing

Auxiliary data processing involved gathering of well-log data and geological cross-sections, and digitizing them to constrain the modeling. Well logs and existing geologic cross-sections used for the constructed gravity and magnetic models come from Guthrie and Raymond (1992); Hatcher (1990); Steltenpohl (2005); Steltenpohl et al. (2008). Collected well-log data contains the latitude, longitude and elevation of wells, as well as a core description and depth to basement rocks. The well-log information was used to create a shape file in ArcCatalog, and then later imported in ArcMap as a separate layer. The shape file was projected in the same coordinate system (NAD_1927_Lambert Conic Conformal-86) as that of the geological map of Alabama (Geological Survey of Alabama, 2006). Once the gravity map, magnetic map, and structural map of Alabama were georeferenced, the well-location layer was overlain on the geologic map (Figure 16) and other maps to correlate rock types and other geological structures along the modeled profiles.

Well logs and cross-sections (Guthrie and Raymond, 1992; Hatcher, 1990; Steltenpohl, 2005) were mainly used to estimate features of the upper crust (e.g., depth to stratigraphic boundaries, projected subsurface positions of faults and structures, rock types, and associated densities and magnetic susceptibilities). Since most published models do not extend deeper than a few kilometers, cross-section from Steltenpohl et al. (2008) and projection along geologic strike of the COCORP seismic reflection models developed for western Georgia (e.g., Nelson et al., 1985a, 1985b; Nelson et al., 1987; McBride and Nelson, 1988; 1991; McBride et al., 2005) were used to help construct starting models for the deeper crust.

Using geologic information on the mapped rock types, a range of densities and magnetic susceptibilities for different rock units was estimated for construction of the initial gravity and magnetic models (Table 2). The density of crustal blocks used by Savrda (2008) in her gravity model was also taken in to account when estimating the density of different blocks for this study.

Table 2. Estimated density and magnetic susceptibility of different crustal blocks and geological units (Dobrin and Savit, 1988; Telford et al., 1990).

<table>
<tr><th>Rock Type</th><th>Density (kg/m^3)</th><th colspan="2">Magnetic susceptibility (SI unit)</th><th>Lithology</th></tr>
<tr><td>Coastal Plain sediment</td><td>2000-2500</td><td colspan="2">0.00010-0.00025</td><td>Sediments</td></tr>
<tr><td>Pre-middle Jurassic rocks</td><td>2300-2800</td><td colspan="2">0.002-0.08</td><td>Phyllite, sandstone, shale, salt, basalt, granophyre, diabase</td></tr>
<tr><td>Rift related mafic volcanic rock</td><td>2800-3200</td><td colspan="2">0.03-0.09</td><td>Basic igneous rock</td></tr>
<tr><td>Gondwanan crust</td><td>2600-2900</td><td colspan="2">0.002-0.0025</td><td>Felsic volcanic rocks, and few mafic rocks</td></tr>
<tr><td>Top of Uchee terrane</td><td>2400-2600</td><td colspan="2">0.001-0.009</td><td>Mylonite</td></tr>
<tr><td>Uchee terrane</td><td>2500-3000</td><td colspan="2">0.0001-0.0009</td><td>biotite gneiss, schist and migmatite</td></tr>
<tr><td>Metasedimentary, top of Pine Mountain window</td><td>2400-2800</td><td colspan="2">0.002-0.006</td><td>Meta-sedimentary rock</td></tr>
<tr><td>Grenville (Laurentian crust)</td><td>2500-3000</td><td colspan="2">0.0001-0.0007</td><td>Metamorphic rock</td></tr>
<tr><td>Inner Piedmont (Dadeville complex)</td><td>2700-3200</td><td>0.01-0.09</td><td rowspan="2">Granitic Plutons</td><td>Amphibolite, schist, and gneiss</td></tr>
<tr><td>Inner Piedmont (Opelika complex)</td><td>2500-2800</td><td>0.001-0.009</td><td>Biotite gneiss, mica schist, quartzite, rare amphibolite</td></tr>
<tr><td>Northern Piedmont (Tallapoosa)</td><td>2500-3000</td><td colspan="2">0.01-0.05</td><td>high grade metasedimentary rocks, and plutonic rocks</td></tr>
<tr><td>Northern Piedmont (Coosa)</td><td>2500-3000</td><td colspan="2">0.002-0.008</td><td>Schist with quartzite, amphibolite</td></tr>
<tr><td>Northern Piedmont (Talladega)</td><td>2500-3000</td><td colspan="2">0.002-0.008</td><td>phyllite, marble, slate, and greenstone</td></tr>
<tr><td>Valley and Ridge</td><td>2500-3000</td><td colspan="2">0.0001-0.00065</td><td rowspan="3">Limestone, shale, dolomite, Chert, sandy dolomite, phyllite, greywacke, conglomerate</td></tr>
<tr><td>Valley and Ridge</td><td>2500-3000</td><td colspan="2">0.0001-0.00065</td></tr>
<tr><td>The Interior Low Plateau (sedimentary rock)</td><td>2500-2700</td><td colspan="2">0.0001-0.00065</td></tr>
<tr><td>Grenville basement (Mafic intrusive bodies)</td><td>2800-3200</td><td colspan="2">? ?</td><td>Basic igneous rock</td></tr>
<tr><td>Pottsville Formation</td><td>2200-2500</td><td colspan="2">0.0001-0.00065</td><td>Sandstone, siltstone, clay, shale, and coal</td></tr>
<tr><td>Mantle</td><td>3300</td><td colspan="2">0.003</td><td>Igneous rock</td></tr>
</table>

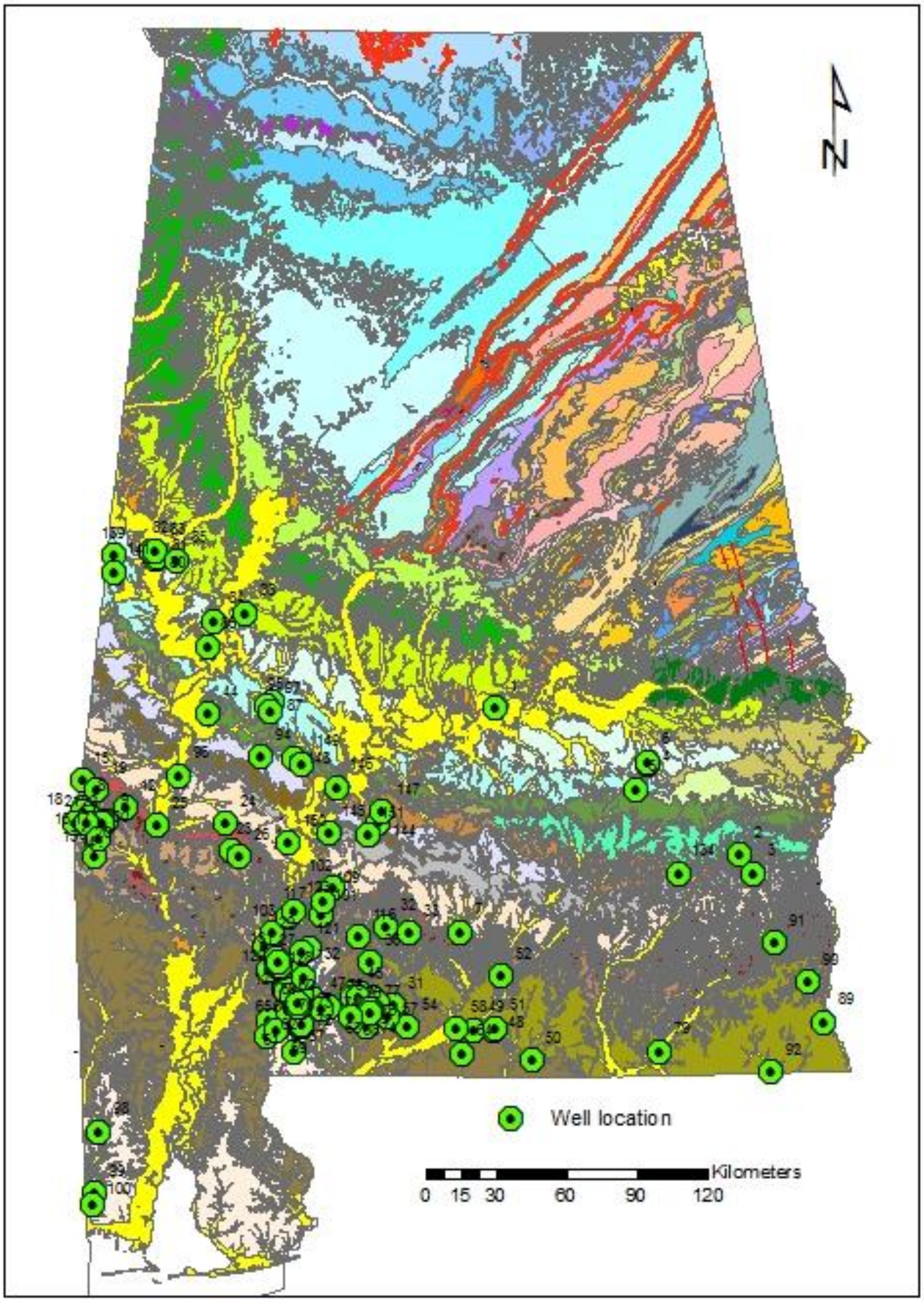

Figure 16. Well locations (Neathery and Thomas, 1975; Guthrie and Raymond, 1992) overlain on Geological Map of Alabama (modified after Geological Survey of Alabama, 2006).

Gravity and Magnetic Modeling in GM-SYS

To perform the gravity and magnetic modeling, three transects ($\overline{\mathrm{AB}}$, $\overline{\mathrm{CD}}$, and $\overline{\mathrm{EF}}$) were extracted from the regional data set and crustal models were built along those transects (Figures 17 and 18). Transects were chosen such that they cross the strike of regional geological and geophysical trends. Transects were selected to cross key tectonic features, such as the South Georgia basin and its associated rift sediments and mafic rocks, the BMA, the NY-AL magnetic lineament, the Pine Mountain window, the Wiggins Arch, the Alexander City fault, the Brevard fault, the Uchee terrane, and the Suwannee terrane.

Crustal models along the three transects were developed using Geosoft's Oasis montaj and GM-SYS software. The GM-SYS software calculates the gravity and the magnetic responses for the same geologic model. Forward modeling, which is used for this study, provides an option for joint inversion of the gravity and the magnetic data. Two-dimensional modeling only considers the gravity effect produced by the bodies directly below the line, whereas $2\frac{3}{4}$ dimensional modeling expands the theoretical calculation to include the effects of the bodies perpendicular to the line.

Key horizons for the model, such as the mantle layer, the Piedmont rocks, and the Coastal Plain sediments were assumed to extend well beyond the model limits in order to reduce the edge effects. Initial models assumed simple parallel layers, which were later modified to create polygons of different shapes and sizes.

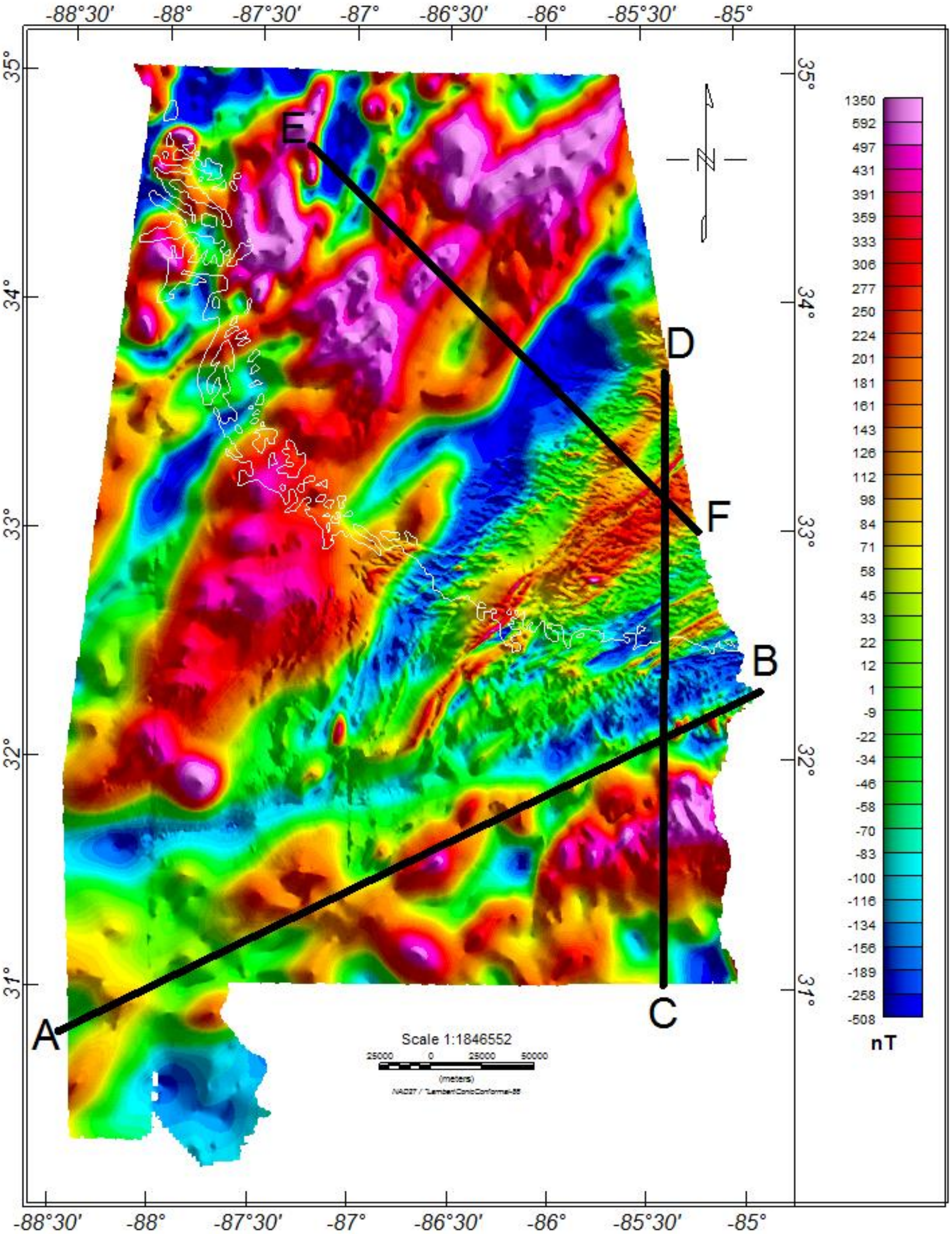

Figure 17. Location of modeled transects on magnetic map.

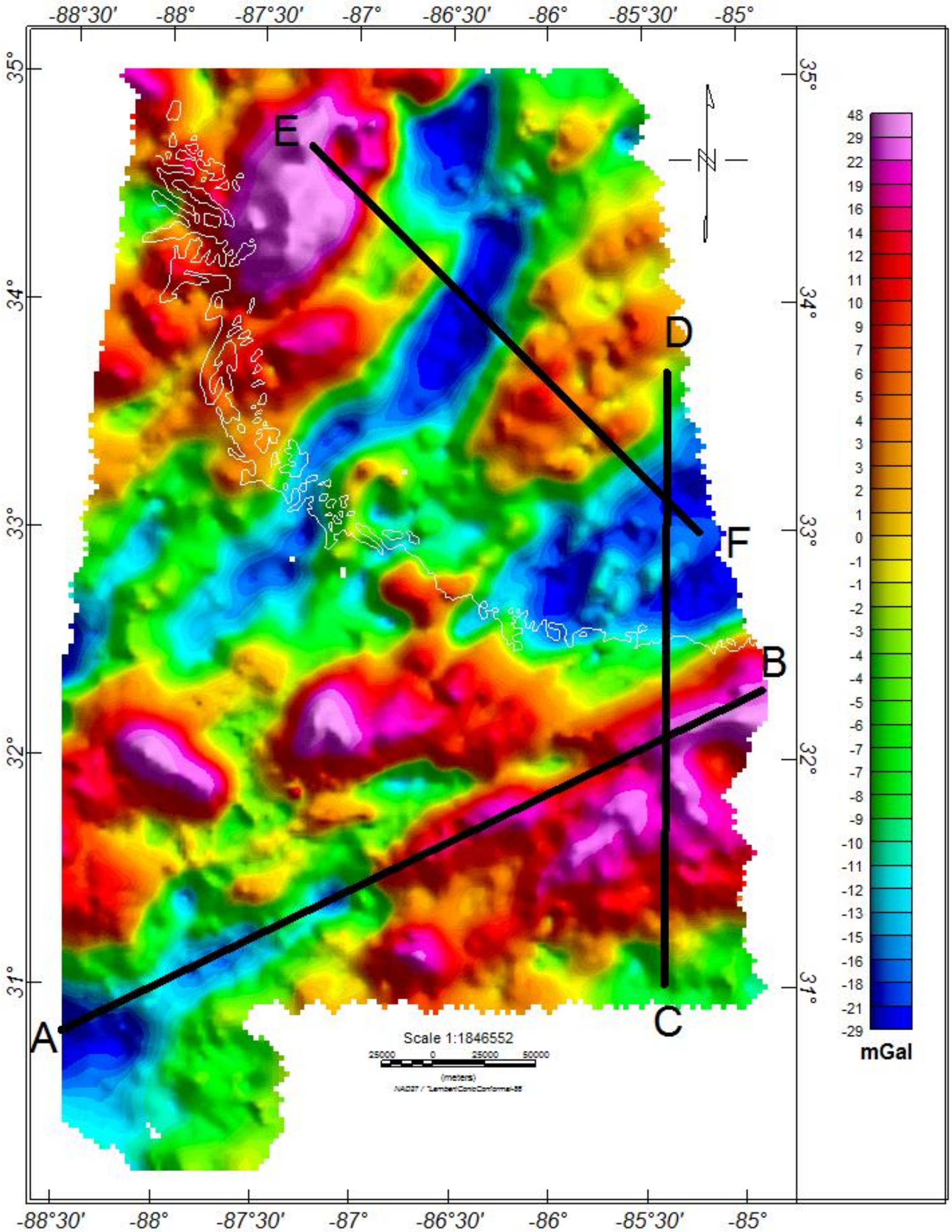

Figure 18. Location of modeled transects on gravity map.

Each polygon used in the models was assigned density and magnetic susceptibility values according to Table 2 and available geological data and borehole data. Locations of structural features on each profile were noted. Each model version was modified until a satisfactory match between calculated gravity and magnetic values and the observed gravity and magnetic values was achieved. Long-wavelength anomalies were matched first, followed by smaller wavelength features of the observed data curves. Depth, geometry, density and magnetic susceptibility were varied within 15 % of initial values in order to obtain a good match between the observed and the calculated curve. Since the inverse modeling process can provide multiple solutions for the same gravity and magnetic anomaly, the nonuniqueness of crustal interpretation is helped by the use of well-log data, geophysical data, geological cross sections, and other available data.

Results

Three magnetic and gravity profiles (AB, CD and EF) are modeled for the study area (Figures 17,18 and 19). Line AB is 380 km long and crosses the state of Alabama from west to east. In the NAD_1927_Lambert Conic Conformal-86 coordinate system, the end coordinates of point A are (-232383.70, 3720817.35) m, and the coordinates of point B are (103443.27, 3883235.48) m. Line CD runs from south to north in eastern Alabama and is 295 km long. End coordinates of points C and D are (56739.19, 3743442.95) m and (57504.72, 4037407.47) m, respectively. Line EF is geographically oriented from northwest to southeast and is 264 km long. The end coordinates of points E and F are (-113503.96, 4147584.60) m and (71806.78, 3962273.86) m.

The potential field properties of the rocks are expressed by various polygons embedded in the cross-section, which are described as positive or negative contrasts relative to a magnetically and gravitationally homogeneous crust. All cross-sections extend 50 km below the surface. The contacts between polygons on the surface and a few kilometers below the surface are constrained by the location of geological units and structures in the geological map of Alabama. The physical properties of the geometrical bodies in the cross-sections are estimated based on the study of available geologic cross-sections, previous gravity and magnetic studies, and reflection seismic profiles from Georgia

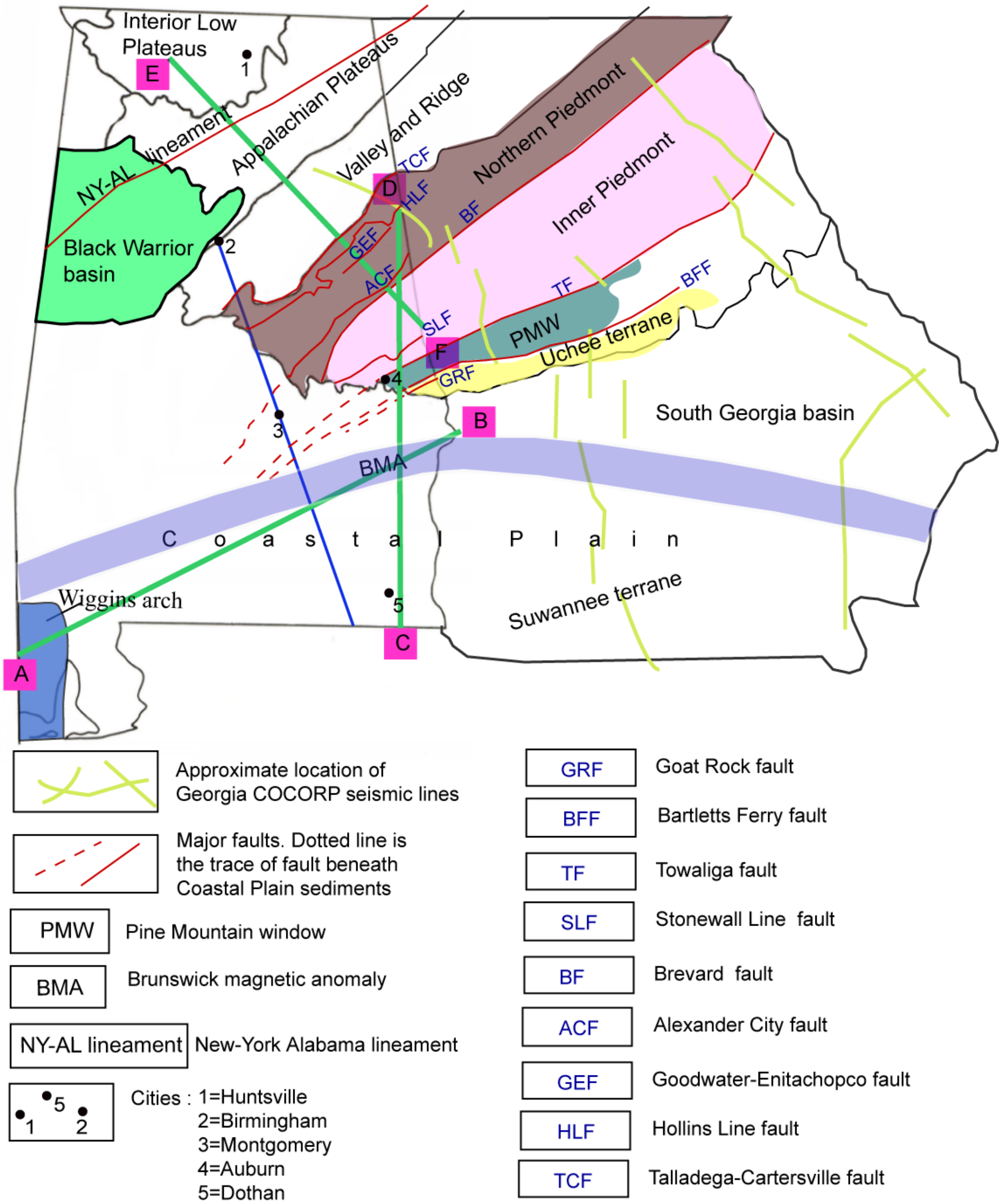

Figure 19. Location of modeled profile lines AB, CD and EF (green lines) on the map with major tectonic structures in Alabama and adjacent Georgia area (modified from Sapp and Emplaincourt, 1975; Hopson and Hatcher, 1988). Blue line from Gulf coast to Birmingham is the profile location of gravity modeling by Savrda (2008). Location of NY-AL lineament (after Steltenpohl et al., 2010).

COCORP lines (e.g., McBride and Nelson, 1988, 1991; Cook et al., 1979, 1981; McBride et al., 2005). Approximately 15 models were constructed along three transects to achieve the best fit of the calculated to the observed data. Results from each model are described below separately.

Gravity and Magnetic Model Along AB

Gravity and magnetic data for the southwest-northeast (profile AB) are modeled by three to four shallow bodies overlying a deeper body (Figure 20). Several wells along or near profile AB are listed in Table 3, along with their location, depth and description of lithology. The observed gravity is the lowest (~ -20 mGal) at the southwestern end of profile AB, and it is maximum (~ 30 mGal) towards northeastern end. Profile AB shows a negative gravity from 0 to 168 km, and the gravity is positive for rest of the profile. There is a long-wavelength gravity anomaly from 163 to 277 km. There is a noticeable magnetic high (~ 320 nT) at 200 km followed by smaller magnetic highs and lows. Sudden changes in magnetic values are observed at 330 km and 350 km, which create peaks of magnetic highs (Figure 20).

Table 3. Description of well logs along profile AB (Guthrie and Raymond, 1992).

Wells no	Distance from A, km	Depth below the surface(m)	Description of rock at the depth in column 3
98	0	-5996.0256	Chlorite-sericite-quartz phyllite
62	92	-5006.9496	Aphanitic basalt (diabase)
47	117.5	-4119.372	Altered diabase
38, 39, 42, 45	128	-4153.2048	Granophyre and granophyric diabase
34/36	140	-3695.0904	Mylonitic granophyre
33	162	-2505.456	Chlorite-epidote granophyre and aphanitc diabase
7	181.13	-2490.5208	Eagle Mills Fm., granophyric olivine diabase, and tuff

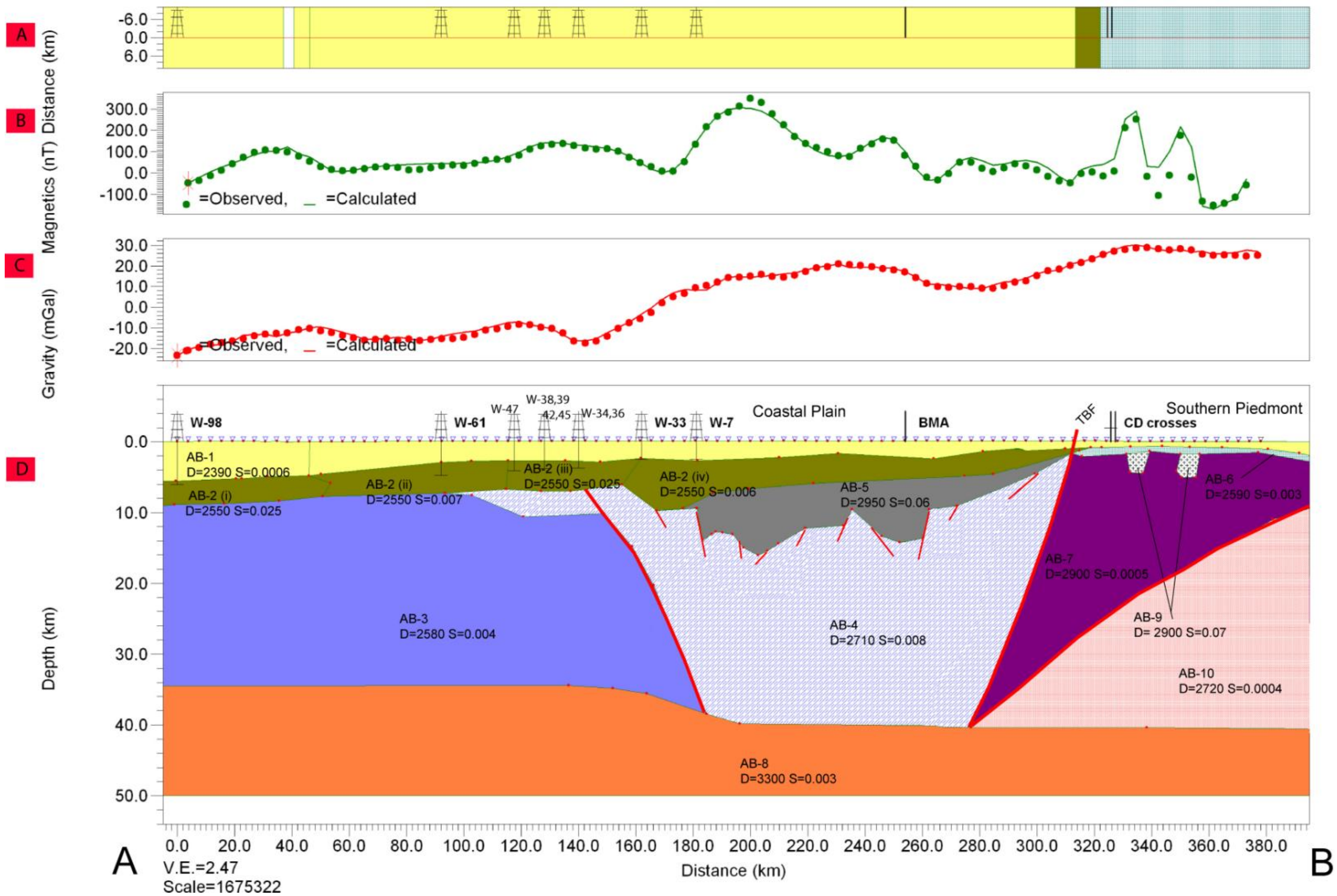

Figure 20. Gravity and magnetic model along AB. (A) Plan view of cross-section with well locations indicated. (B) Magnetic profile along AB. Green dots represent observed magnetic data; green line represents calculated data predicted from model. (C) Gravity profile along AB. Red dots represent observed gravity data; red line represents calculated gravity data predicted from model. (D) Cross section along AB. Density, D, and magnetic susceptibility, S, used is shown for each polygon in cross-section. Profile location is given in Figures 17, 18 and 19.

The deepest body in the model is a high-density body representing the upper mantle layer (D=3300, S=0.003) (Figure 20, Table 4). Based on varying observed gravity and magnetic values, six to eight other bodies are embedded in the cross-section to match the observed and calculated gravity and magnetic curves.

Table 4. Location of polygons and their depths for profile AB, with densities and susceptibilities for each profile.

Body number		Position from point A (km)	Depth to top (km)	Maximum thick ness of the body (km)	Expected		Rock description based on geological map, well description, literature and density/magnetic susceptibility
					Rock density (D)	Magnetic susceptibility of rock (S)	
AB-1		0-320	0	5	2390	0.0006	Sediments
AB-2	i	0-52	5	3.75	2550	0.025	phyllite with basalt and diabase
	ii	52-114	2 to 4	4.5	2550	0.007	granophyric olivine diabase, and tuff
	iii	114-160	2.38	4.5	2550	0.0025	Basalt and diabase
	iv	160-319	2	7.22	2550	0.006	Felsic igneous rock
AB-3		0-195	8.5	26.5	2580	0.004	Low grade metamorphic rock
AB-4		90-380	1.5-20	38.5	2710	0.008	Felsic volcanic rock and few mafic rock
AB-5					2950	0.06	Mafic rock
AB-6		314-380	0	3.25	2590	0.003	Mylonite
AB-7		314-380	3.25	18.25	2900	0.0005	Gneiss, schist, migmatite
AB-8		0-380	35-40	15	3300	0.003	Upper mantle
AB-9		330 and 352	3	5	2900	0.06	Mafic rock
AB-10		280-395	> 18	28	2720	0.0004	Crystalline metamorphic rocks

The data from wells along the profile are used to constrain the contact between the uppermost low-density and low-susceptibility layer AB-1 (D=2390, S=0.0006) and the underlying layer AB-2, with its relatively higher density (D=2550). Layer AB-2 varies in magnetic susceptibility from place to place (0.035, 0.06, 0.07, and 0.0025), thus forming four distinct bodies within AB-2.

The first section (0-180 km) has five different layers, including upper mantle layer AB-9. A 26.5-km-thick body, AB-3, with low-density (D=2580, S=0.004) is inserted to match a gravity low of ~ -20 mGal. From 103 to 180 km, layer AB-4 (D=2710, S=0.008) replaces the upper part of AB-3. This relatively high- density layer helps to match the increasing gravity at 160 km. Other bodies in this section of profile are AB-1 and AB-2. The second section (180 to 312 km) in model AB also has five different layers including AB-1, AB-2, AB-5. A layer AB-5 (D=2.95, S=0.06) with high density and high magnetic susceptibility is embedded underneath AB-2 in order to match the gravity-high (~ 20 mGal) and magnetic-high (~ 320 nT) anomaly at that place. The zigzag shape of the base of AB-5 is required to match the other smaller magnetic highs. The thickness of AB-5 is only 2 km between 277 km and 305 km to match the relatively low-gravity (~ 9 mGal) and low-magnetic (~ 10 nT) anomaly. The third section (315 to 376 km) contains four layers. A low-density layer AB-6(D=2590, S=0.003) is underlain by a layer with low density and low magnetic susceptibility AB-1 (D=2390, S=0.0006). A thick wedge shape layer AB-7 in this section compensates for the high-gravity anomaly. Two circular bodies with high magnetic susceptibility AB-9 (D=2900, S=0.07) are embedded within layer AB-7 to match the abrupt change in observed magnetic value. A thick basement layer AB-10 (D=2720, S=0.0004) is inserted in order to match the gravity high anomaly. Model AB intersects with model CD at 325 km and similar polygons are embedded in model CD at the region where two lines intersect with one another.

Gravity and Magnetic Model Along CD

Gravity and magnetic data for the south-north profile (profile CD) are well modeled by a series of shallow bodies overlying two to three deeper bodies of varying properties (Figure 21 and Table 5). The model is constrained by four wells (Table 6). These wells only penetrate few kilometers and help to constrain the boundary between a top layer with an underlying layer.

On the south end of profile CD, a long-wavelength gravity high and a magnetic high can be observed. The positive gravity anomaly reaches a maximum of ~ 23 mGal at the distance of 90 to 120 km, and then decreases to ~ -50 mGal at ~ 190 km. There is a gravity low (~ -50 mGal) from 190 to 230 km and then the gravity increases slightly toward north. The magnetic anomaly reaches its maximum (~440 nT) at 70 km, and then it decreases to ~ -200 nT from 105 to 130 km. There are two magnetic highs at 220 km and 280 km and a magnetic low of ~-34 nT from 260 km to 270 km. The magnetic anomaly is not smooth like the gravity anomaly, as magnetic anomalies are more influenced by shallow bodies, and the long-wavelength gravity anomalies are usually caused by deep-seated bodies. There are abrupt changes in magnetic values at ~155 km, ~187 km and ~245 km. These are places where the profile line crosses major faults according to geologic map of Alabama (Figure 19).

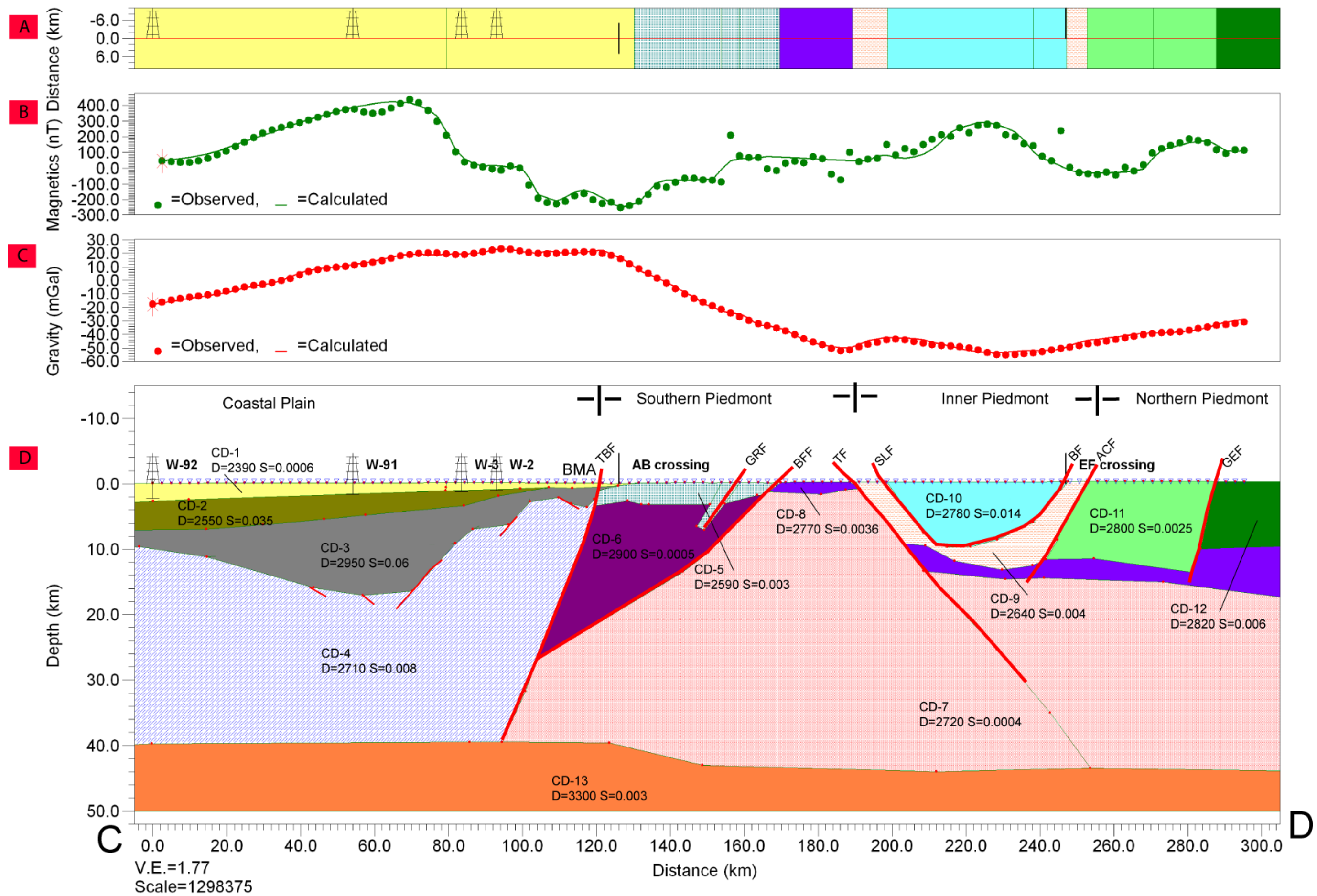

Figure 21. Gravity and magnetic model along CD. (A) Plan view of cross-section with well locations indicated. (B) Magnetic profile along CD. Green dots represent observed magnetic data; green line represents calculated data predicted from model. (C) Gravity profile along CD. Red dots represent observed gravity data; red line represents calculated gravity data predicted from model. (D) Cross section along CD. Density, D, and magnetic susceptibility, S, used is shown for each polygon in cross-section. Profile location is given in Figures 17, 18 and 19.

The lowermost body in the cross-section is a high-density (D=3300, S=0.003) rock and represents the upper mantle. Profile CD is divided into four sections described separately here. The first section (0 to 120 km) requires five different bodies with different densities and magnetic susceptibilities to match the gravity and magnetic curves.

Table 5. Location of bodies' and their depths, and the expected rocks' densities and susceptibility for profile CD.

Body number	Position from point C (km)	Depth to top (km)	Maximum thickness of the body (km)	Expected		Rock description based on geological map, well description, literature and density/magnetic susceptibility
				Rock density (D)	Magnetic susceptibility of rock (S)	
CD-1	0-125	0	2.5	2390	0.0006	Sediments
CD-2	0-125	0.7-2.5	4.75	2550	0.035	Shale, sandstone, phyllite, evaporite with few basalt, diabase and granophyre
CD-3	0-125	7.25-1.0	12	2950	0.06	Mafic rock
CD-4	0-120	3-16	33	2710	0.008	Felsic volcanic rock and few mafic rock
CD-5	120-167	0	3.25	2590	0.003	Mylonite
CD-6	120-167	3.25	18.25	2900	0.0005	Gneiss, schist, migmatite
CD-7	95-300	2-104	41	2720	0.0004	Crystalline metamorphic rocks
CD-8	167-192	0 and 10	2	2700	0.0036	Quartzite, marble, schist
CD-9	192-253	0-9	6.25	2640	0.004	Schist
CD-10	199-247	0	9.25	2780	0.014	Amphibolite, schist, gneiss
CD-11	252-287	0	13	2800	0.025	Gneiss, schist, quartzite
CD-12	287-300	0	13	2820	0.006	Garnet schist with quartzite and graphitic schist

CD-13	0-300	40-43	10	3300	0.003	Upper mantle

The low-density top layer CD-1 (D=2370, S=0.0006) is underlain by a polygon CD-2 (D=2550, S=0.035). Both of the layers are thick at the south end and the thickness decreases towards the north to match the increasing gravity. Below CD-1 and CD-2, a high-density and high-susceptibility body CD-3 (D=2950, S=0.0553) is necessary in order to match the gravity high (~ 22 mGal) and magnetic high (~420 nT). Another 33 km-thick body CD-4 (D=2710, S=0.008) is embedded at the basement and extends to the mantle.

Table 6. Description of wells along CD.

Wells no	Distance from C (km)	Depth below the surface(m)	Description of rock at the depth in column 3
92	0	-2260.7016	Light-gray quartz arenite, gray shale and sandstone
91	54	-1629.7656	Eagle Mills and diabase/basalt
3	83.5	-1179.8808	Eagle Mills Fm. and olivine diabase
2	93	-1033.272	Eagle Mills Fm. basalt (diabase)

The second section (120 to 170 km) has a low-density top layer CD-5, which overlies a polygon CD-6 (D=2900, S=0.0005). The high-density body CD-6 is forming a wedge and extends ~ 25 km to match the gravity high (~ 20 mGal). The prominent magnetic low at 100 to130 km is achieved by the magnetic susceptibility contrast of 0.0523 SI units and the oblique contact between bodies CD-1, CD-3 with CD-5, CD-6.

The third section (170-190 km) exposes a basement body, CD-7 (D=2720, S=0.0004), in the surface, covered by a thin polygon, CD-8 (D=2770, S=0.0036). Lastly, the fourth section (190 to 300 km) has three polygons (CD-10, CD-11, and CD-12) with similar densities, but different magnetic susceptibilities. Polygon CD-10 (D=2780, S=0.014), which extends ~9 km below the surface, is required from 200 to 245 km to match the magnetic high (~ 285 nT) there. CD-10 is surrounded by a gravity-low body CD-9 (D=2610, S=0.004), so that an asymmetrical convex gravity anomaly can be matched perfectly. Toward the end of the profile line, two different bodies (CD-11 and CD-12) with $\Delta D = 20$ kg/m^3 are embedded. The magnetic susceptibility of CD-11 is 0.025 and that of CD-12 is 0.006. This difference in magnetic susceptibility is required to create a magnetic anomaly there.

Gravity and Magnetic Model Along EF

The gravity and magnetic model along profile EF contains several shallow bodies (EF-1 to EF-11) underlain by a 35 to 40-km thick layer, EF-12 (D=2720, S=0.0004), which is underlain by EF-13(D=3300, S=0.003) (Figure 22). The bottom layer of the model comprises of upper mantle layer EF-13. Location of bodies and their depth, as well as their theoretical densities and magnetic susceptibilities are given in Table 7. The gravity and magnetic model along EF starts with a gravity high (~ 20 mGal) and magnetic-low (~ -400 nT), followed by a gravity low (~ -50 mGal) and magnetic high (~ 475 nT) at 100 km. The gravity gets relatively high (~ -20 mGal) from 100 to 160 km and again lowers to ~ -55 mGal towards the end of the profile. The magnetic data reaches its maximum (~ 500 nT) at 135 km and decreases to ~ -150 nT from 150 to 200 km. Towards the end of the profile, a positive magnetic anomaly is seen, which reaches up to ~ 200 nT.

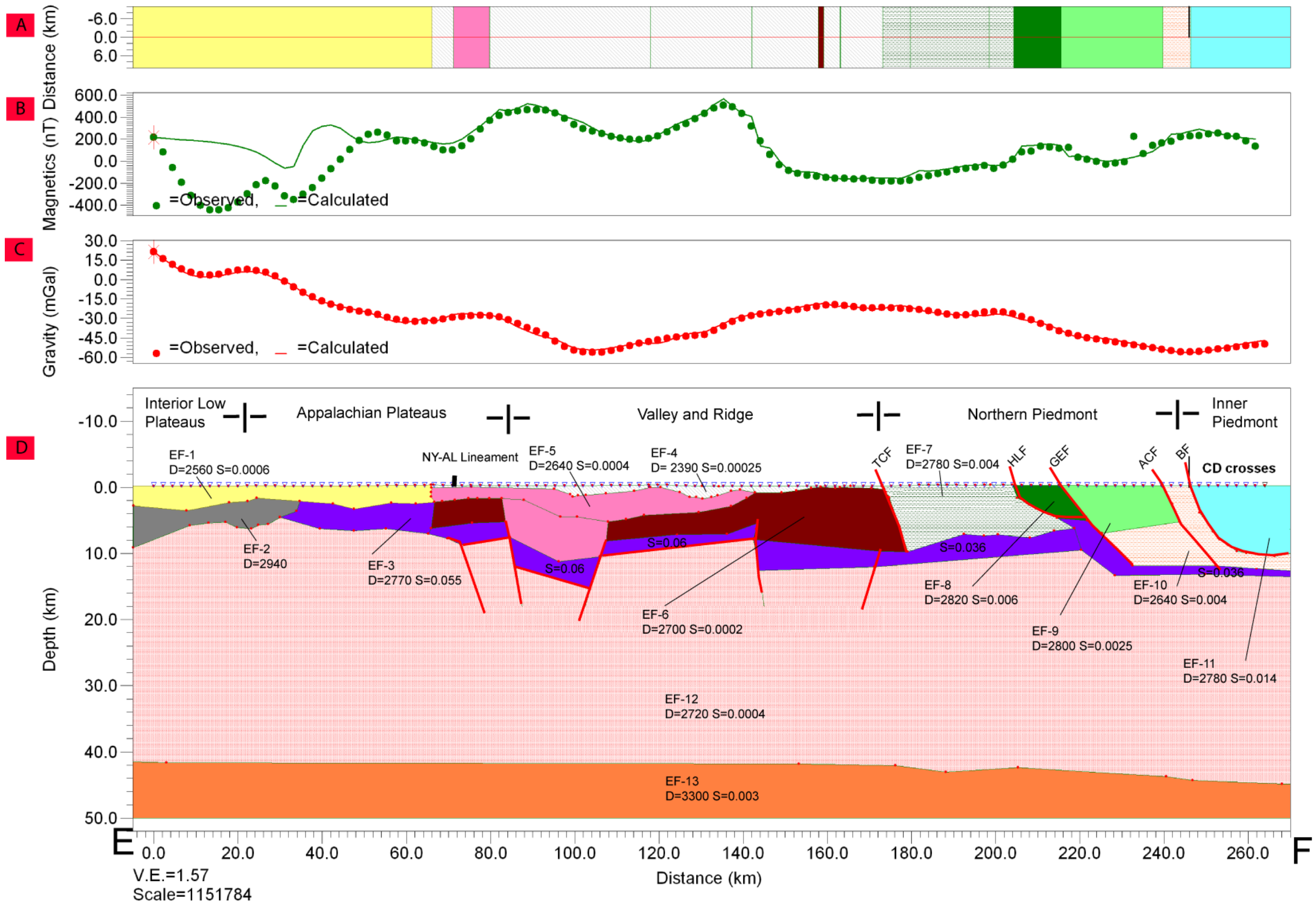

Figure 22. Gravity and magnetic model along EF. (A) Plan view of cross-section with well locations indicated. (B) Magnetic profile along EF. Green dots represent observed magnetic data; green line represents calculated data predicted from model. (C) Gravity profile along EF. Red dots represent observed gravity data; red line represents calculated gravity data predicted from model. (D) Cross section along EF. Density, D, and magnetic susceptibility, S, used is shown for each polygon in cross-section. Profile location is given in Figures 17, 18 and 19.

Table 7. Location of bodies, their depths, and expected densities and susceptibility for profile EF.

Body number	Position from point E (km)	Depth to top (km)	Maximum thickness of the body (km)	Expected		Rock description based on geological map, literature and density/magnetic susceptibility
				Rock density (D)	Magnetic susceptibility of rock (S)	
EF-1	0-65	0	3.3	2560	0.0006	Sedimentary rocks
EF-2	0-34.5	3.5-1.3	4.8	2940	0.06	Mafic rocks
EF-3	0-265	Various	2	2770	0.0036-0.06	Quartzite, marble, schist
EF-4	94-173	0	1	2390	0.00025	Sandstone, siltstone, clay, shale, and coal
EF-5	94-173	0.03-11	10.5	2640	0.0004	Phyllite, slate, greywacke, conglomerate, Silurian to Devonian limestone
EF-6	94-173	0.3-9.8	9.5	2700	0.0002	Cambrian rocks (Knox Group dolomitic limestone and Copper Ridge dolomite)
EF-7	173-204	0	9.75	2820	0.009	Phyllite, marble, slate
EF-8	204-215	0	4	2820	0.006	Garnet schist with quartzite and graphitic schist
EF-9	215-239	0	12	2800	0.025	Gneiss, schist, quartzite
EF-10	239-246	0-9	6.5	2640	0.004	Schist
EF-11	246-265	0	10.3	2780	0.014	Amphibolite, schist, gneiss
EF-12	0-265	5-44	39	2720	0.0004	Crystalline metamorphic rocks
EF-13	0-265	41-44	9	3300	0.003	Upper mantle

The bottom layer of the model comprises of upper mantle layer (D=3300, S=0.003). From 0 to 65 km of the profile, 3-km-thick low-density and low-susceptibility layer, EF-1 (D=2560, S=0.0006) is embedded. Below EF-1, a high-density body EF-2 (D=2940) is required to match the gravity high (~15 mGal) there. A thin, low-density layer, EF-4 (D=2390, S=0.00025), forms the top layer from 94 to 173 km. Below this top layer, EF-5 (D=2640, S=0.0004) and EF-6 (D=2700, S=0.0002), are embedded. These relatively low-density layers are required to match the low-gravity anomaly here. EF-5 is 10 km thick at 100 km to match a gravity low (~-55 mGal), and the layer becomes thinner towards northeast.

From 173 to 215 km three bodies (EF-7, EF-8, EF-9) with similar density are embedded. The contacts between these bodies are based on the geological map of Alabama (Geological Survey of Alabama, 2006). At 215 km, a low-density body, EF10 (D=2640, S=0.004), is required to match the low-gravity curve. At the end of the profile, body EF-11 (D=2780, S=0.014) is 10 km thick and underlain by EF-10 and EF-3 (D=2700, D=0.0036). Layer EF-3 is embedded throughout the profile as a Laurentian basement cover sequence indicated from COCORP Seismic studies (Cook et al., 1979).

Interpretation

Interpretation of Profile AB

The location of profile AB was selected in order to model the tectonic terranes buried beneath the Coastal Plain sediments (Figures 17, 18 and 19 for profile location). The geologic interpretation of the models, described below, is the result of comparing the densities and magnetic susceptibilities of bodies in the model with the values of rock types listed in Tables 2 and 4, and superimposing the locations of mapped geologic structures, such as contacts and faults, on the models (Figure 23).

AB-8 is the deepest body in the model and is interpreted as the upper mantle layer based on its high density (D=3300, S=0.003). Depth to this layer is similar to that found in various locations crossed with the Georgia COCORP lines (Nelson et al., 1985; McBride and Nelson, 1988, 1991). Layer AB-1 represents the Coastal Plain province, which consist of rock units and sediments younger than the Middle Jurassic Period. The sedimentary rocks and unconsolidated sediments of AB-1 have low density (D=2390). The susceptibility of this unit is low (S=0.0006) and thus not responsible for the magnetic anomaly. Layer AB-2 is interpreted as the Pre-middle Jurassic basin fill, which includes the siliciclastic red beds of Eagle Mills formation intercalated with mafic rocks (basalt, diabase and, granophyre) (Guthrie and Raymond, 1992). The expected density of this

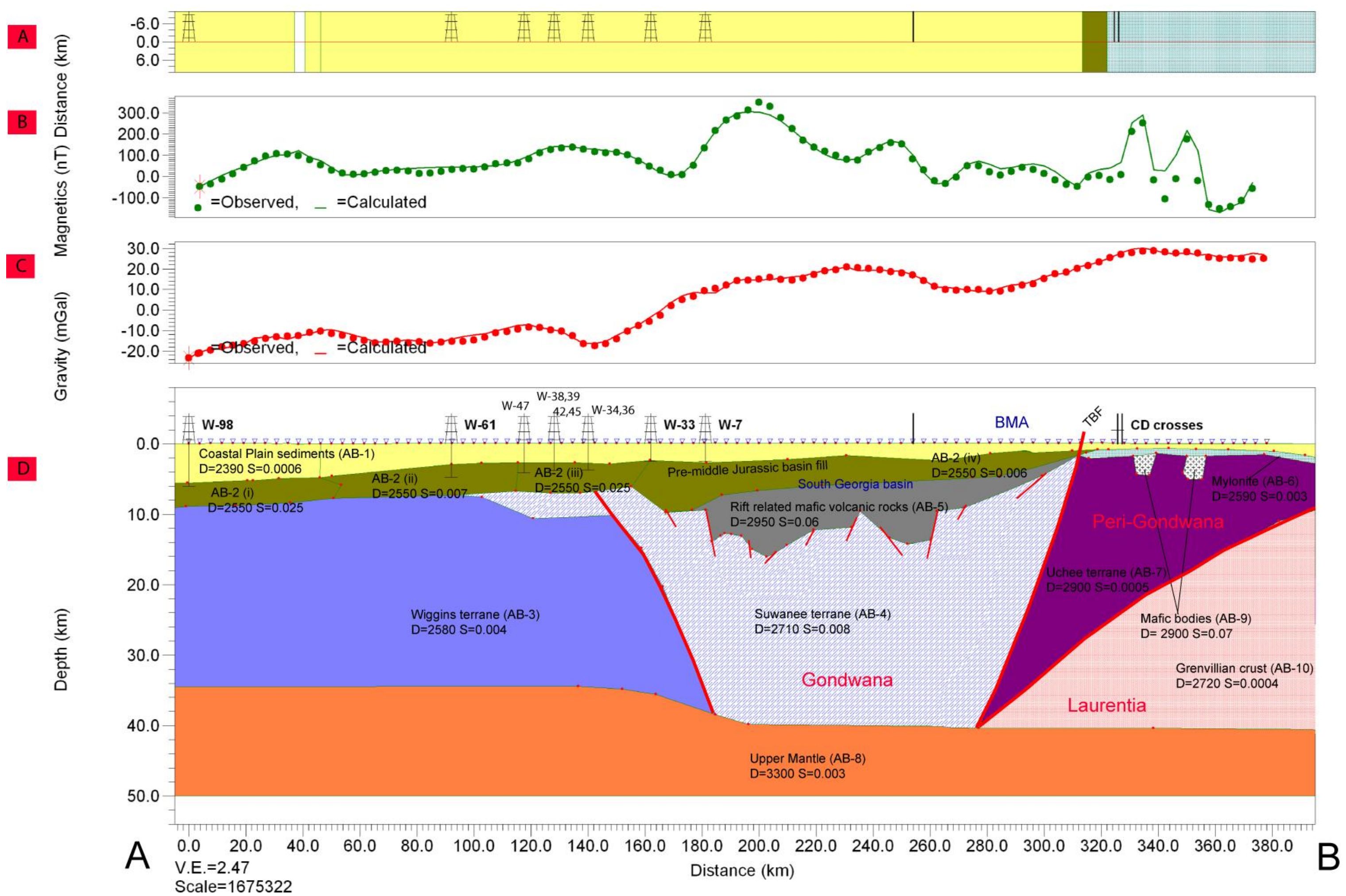

Figure 23. Geologic interpretation of profile AB based on gravity and magnetic modeling. (A) Plan view of the cross-section. (B) Magnetic profile along AB. Green dots represent observed magnetic data; green line represents data predicted from model. (C) Gravity profile along AB. Red dots represent observed gravity data; red line represents gravity data predicted from model. (D) Interpreted geologic cross-section along the modeled profile. Profile locations are given in Figures 17, 18 and 19.

layer is 2550 kg/m^3. Since layer AB-2 contains mafic minerals, high magnetic susceptibility is expected. The magnetic susceptibility of this layer could be variable based on the percentage of mafic minerals. Based on magnetic anomalies, the Pre-middle Jurassic basin fill layer is divided into four blocks with varying magnetic susceptibility, ranging from 0.0025 to 0.035.

About 27 km thick, low density (D=2580) layer AB-3 is interpreted as Wiggins terrane. This low density layer is required by the negative gravity anomaly from 0 to 140 km. The low grade metamorphic rock of Wiggins terrane might be the cause of this negative gravity anomaly.

Another thick body AB-4 with high density (D=2710) is interpreted to be felsic volcanic rocks and high grade metamorphic rocks of Suwannee terrane. Body AB-5 (D=2950, S=0.06) is interpreted as rift-related mafic volcanic rocks. This high density body is required for the gravity and magnetic high anomaly from 170 to 300 km. A series of magnetic high and low from 165 to 315 km is interpreted to be formed by the magnetic susceptibility difference of 0.054 SI units between mafic rocks (AB-5) with overlying Pre-middle Jurassic rocks (AB-2, iv) and underlying felsic volcanic rock (AB-4).

The zone of low magnetic and low gravity anomaly from 253 to 324 km is the area, where Brunswick Magnetic Anomaly (BMA) lies. From 314 to 380 km, there are four different rock units.

The top layer AB-6 is interpreted as the mylonite based on geological map and the low density (D=2590) value. This mylonite zone overlies a high density layer AB-7 (D=2900), which is interpreted as the Uchee terrane. High grade gneiss, schist and migmatite of Uchee terrane is creating the gravity high toward the end of profile AB. A couple of high susceptibility circular bodies AB-9 (S=0.07) are interpreted as localized mafic intrusive bodies. At 326 km, profile AB crosses profile CD. Similar bodies are embedded in model along profile CD to match both of the observed gravity and observed magnetic curves with the calculated curves.

Interpretation of Profile CD

The profile line CD, which runs from south to north, is an important profile line among three because CD intersects with both AB and EF. Also, profile CD crosses major faults and tectonic terranes in Alabama (Figures 17, 18 and 19). Using the gravity and magnetic model (Figure 21) and comparing the physical properties of embedded bodies with the properties of geological units (Table 2 and 5), an interpreted geological cross-section is proposed (Figure 24).

Polygon CD-1 in gravity and magnetic model along profile CD is interpreted as Coastal Plain sediments. The low gravity and low magnetic susceptibility of this layer is appropriate for the sedimentary rock and unconsolidated sediments found in this province (Table 5, Guthrie and Raymond, 1992). The depth to the contact of this layer with underlying relatively high density layer CD-2 is constrained by wells 91, 3 and 2(Table 6).

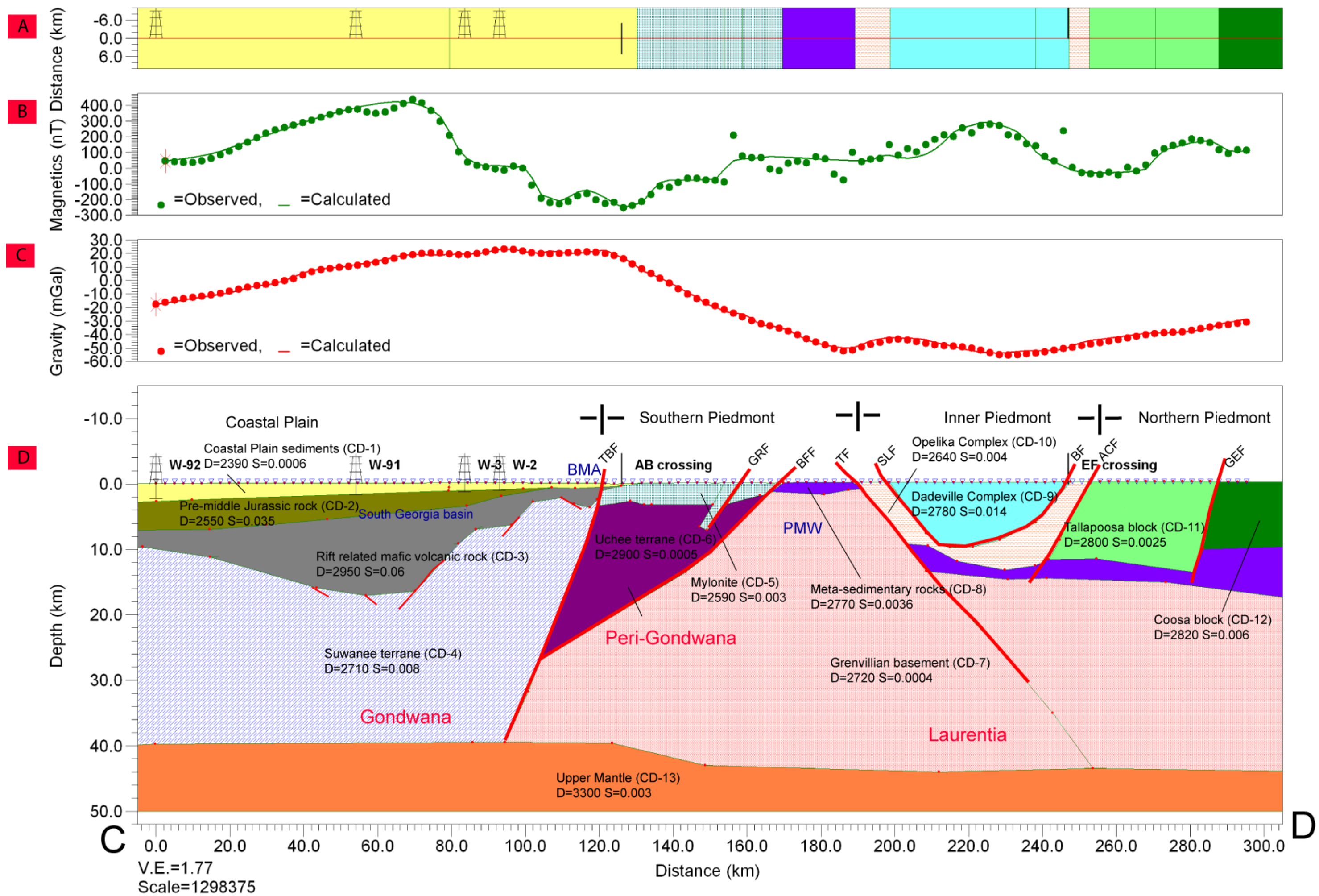

Figure 24. Geologic interpretation of profile CD based on gravity and magnetic modeling. (A) Plan view of the cross-section. (B) Magnetic profile along CD. Green dots represent observed magnetic data; green line represents data predicted from model. (C) Gravity profile along CD. Red dots represent observed gravity data; red line represents gravity data predicted from model. (D) Interpreted geologic cross-section along the modeled profile. Profile locations are given in Figures 17, 18 and 19.

CD-2 is interpreted as Pre-middle Jurassic basin fill, which are denser than the Coastal Plain sediments. Due to the presence of diabase and basalt in this layer, the susceptibility is high (S=0.035). Layer CD-2 overlies a high-density (D=2950) and high-susceptibility (S=0.06) layer CD-3, which is interpreted to be rift-related mafic volcanic rock. This high-density layer is compensating the 8-km thick low-density rocks of CD-1 and CD-2. A thick layer, CD-4, is interpreted as felsic volcanic rocks and high-grade metamorphic rocks of Suwannee terrane. The rocks of layer CD-3 are also of the Suwannee. The northern boundary of Suwannee terrane is marked by Terrane Boundary fault (TBF) of Guthrie and Raymond (1992). TBF also performs as a boundary between the Piedmont province and the Coastal Plain province.

To the north of the TBF, layer CD-5 is a low density (D=2590) body interpreted to be mylonite. This layer is underlain by a high-density (D=2900) block, AB-6. This block is considered as the schist, gneiss and migmatite of Uchee terrane. The felsic volcanic rock (S=0.008) of Suwannee terrane and Uchee terrane (S=0.003) has low magnetic susceptibility compared with the mafic rocks (S=0.06) associated with rifting. This difference in susceptibility creates a magnetic low from 103 to 128 km (Figure 24). This area of low magnetic values is interpreted as Brunswick magnetic anomaly (BMA). A sharp magnetic gradient at 155 km is produced by the Goat Rock fault (GRF). There is another magnetic gradient and a gravity low of -54 mGal at 186 km, which is the location of Bartletts Ferry fault (BFF), which is the northern boundary of Uchee terrane, thus marking the suture between Laurentia and the Peri-Gondwanan arc terranes (Steltenpohl et al., 2008, 2010)

A thin layer, CD-8, is interpreted as the metasedimentary cover rock of the Pine Mountain window. The high density (D=2770) and moderate susceptibility (S=0.0036) of this layer reflects the presence of marble, quartzite and schist. This layer overlies a thick layer, CD-7, which is interpreted as the crystalline metamorphic basement of Pine Mountain window. Layer CD-7 is also known as Grenvillian basement. This layer is faulted by Towaliga fault at 194 km. To the north of the Towaliga fault, the cover rocks of the Pine Mountain window (CD-8) are about 9 km below the surface, according to COCORP Seismic profile line 15 (Nelson et al., 1987: Figure 9). Towaliga fault serves as a boundary between the Southern Piedmont and Inner Piedmont and is expressed as the magnetic gradient and gravity low.

Layer CD-9 is a low density (D=2640) unit, which is interpreted as the meta-sedimentary rocks of Opelika Complex. To the north of Opelika Complex, a high-density (D=2780) block, CD-10, is present. This block is interpreted as the meta-igneous rocks of Dadeville Complex. The gneiss, quartzite and schist of Opelika Complex continue beneath the Dadeville Complex. The high magnetic susceptibility (S=0.014) of CD-10 is predicted here due to the presence of abundant of amphibolite in the Dadeville Complex rocks. Stonewall Line fault (SLF) separates the Opelika Complex with Dadeville Complex. This fault also has a small magnetic signature. Due to the presence of a series of faults along the profile line at 152, 168, 189, 199, 245 and 252 km, the observed magnetic curve shows several areas with sharp magnetic gradients (Figure 21). The Brevard fault is the northern boundary of Dadeville Complex. The Brevard fault (BF) and the Alexander City fault (ACF) are marked by a prominent magnetic high at 245 km. These faults separate the Inner Piedmont with Northern Piedmont.

Layer CD-11 (D=2800, S=0.0025) is interpreted as the Tallapoosa block of the Northern Piedmont. The high-grade meta-sedimentary rocks and felsic plutonic rocks of the Tallapoosa block are thought to have higher density than the Dadeville Complex rocks as suggested by the increasing gravity toward north. The lower susceptibility of Tallapoosa block is generating a low magnetic anomaly there. To the north of CD-11, relatively high density (D=2820) block, CD-12, is interpreted to be the Coosa block. The garnet mica schist and graphitic schist of Coosa block is separated from Tallapoosa block by Goodwater-Enitachopco fault (ECF). This fault is marked by a magnetic gradient at 287 km. The lowermost layer in the gravity and magnetic model CD-13 represents rocks of the upper mantle.

Profile CD crosses profile AB at 125 km and profile EF at 246 km. There is good tie between the anomalies and embedded bodies at the point of intersection. The density, magnetic susceptibility, and thickness of embedded bodies are identical at this tie point.

Interpretation of Profile EF

Profile line EF is located in northeastern part of Alabama (Figures 17, 18 and 19 for profile location). This line crosses the Interior Low Plateau, the Appalachian Plateau, the Valley and Ridge, and the Piedmont provinces. Using the modeled section (Figure 25) and comparing the result of modeling with the density and magnetic susceptibility of different geological units (Table 2 and 7), a geological interpretation of modeled body is proposed (Figure 25). This profile crosses profile CD at 245 km, and there is a good tie between the models CD and EF from northwest to southeast along profile EF, several

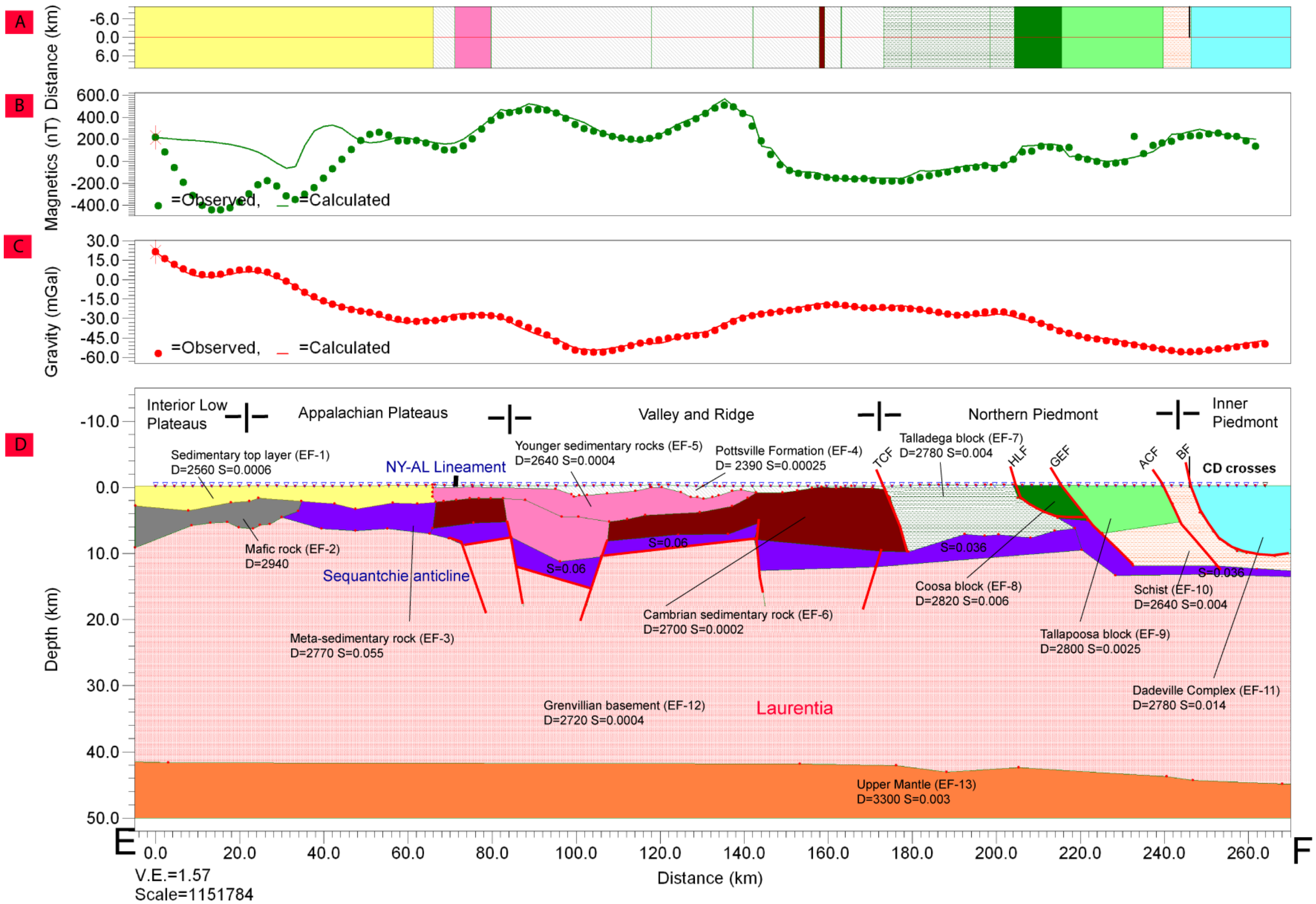

Figure 25. Geologic interpretation of profile EF based on gravity and magnetic modeling. (A) Plan view of the cross-section. (B) Magnetic profile along EF. Green dots represent observed magnetic data; green line represents data predicted from model. (C) Gravity profile along EF. Red dots represent observed gravity data; red line represents gravity data predicted from model. (D) Interpreted geologic cross-section along the modeled profile. Profile locations are given in Figures 17, 18 and 19.

bodies are interpreted. EF-1 (D=2560, S=0.0006) is interpreted as the sedimentary top layer.

This layer overlies layer EF-2 (D=2940), which is considered to be mafic igneous rock, but the observed magnetic low could not be matched with the high susceptibility of mafic rock. This block could be the high density rock (dolomite?) located east of magnetic-high body (see magnetic and gravity map, figures 17 and 18). This magnetic-high body is interpreted as mafic rocks by Steltenpohl and others (submitted). EF-3 (D=2770) is interpreted as a basement layer to the overlying bodies. These are platformal Laurentian continents and mylonites within the Southern Appalachian detachment (Cook et al., 1979). The high density of this body suggests that the body is composed of meta-sedimentary rock or high-grade metamorphic rock. Layer EF-3 is spread throughout the whole profile. This layer is faulted, generating horst and graben structures beneath the Valley and Ridge province (Figure 25). Layer EF-4 serves as the top layer from 94 to 173 km, which is interpreted as a combination of sandstone, siltstone, clay, shale and coal of Pottsville formation.

The Pottsville Formation overlies layers EF-5 and EF-6. EF-5 (D=2640, S=0.0004) constitutes Middle Paleozoic sedimentary rocks (sandstone, shale, conglomerate, limestone, combined) of the Valley and Ridge province. EF-6 (D=2700, S=0.0002) is interpreted as Cambrian sedimentary rocks, which include Knox Group dolomitic limestone and Copper Ridge Dolomite. Rocks in EF-5 and EF-6 are not associated with significant amounts of magnetic minerals. Thus, the magnetic highs in the area may be generated from the mafic minerals in underlying meta-sedimentary rock (EF-3). Layer EF-3 varies in magnetic susceptibility from place to place, likely reflecting variability in the percentage of magnetic minerals. The Valley and Ridge province is bounded in the south by Talladega-Cartersville fault (TCF) system. Even though this fault is not associated with a gravity or magnetic signature, the fault's location is shown on the profile.

To the south-east of the TCF, block EF-7(D=2780, S=0.004) is interpreted as the phyllite, marble and slate of the Talladega block, Northern Piedmont. The magnetic susceptibility of underlying meta-sedimentary rock (EF-3, S=0.036) is comparably low here. Block EF-8 (D=2820, S=0.006) is interpreted as the Coosa block of the Northern Piedmont. The estimated density of rocks in Coosa block is higher compared to Talladega block because of the presence of garnet schist with quartzite in. Layer EF-3 is thicker below the Coosa block, which caused a magnetic high at that point. The Hollins Line fault (HLF) and Goodwater-Enitachopco fault (GEF) shows small magnetic signatures. Another block EF-9 (D=2800, S=0.0025) is interpreted as gneiss, schist, and quartzite of Tallapoosa block. Lower density of layer EF-10 (D=2640, S=0.004) compensates the high density rocks EF-9 and EF-11 and helps to match the lower values of observed gravity. Layer EF-10 is interpreted as schist of the Brevard zone and Opelika Complex (i.e., Emuckfaw Formation, as in Steltenpohl, 2005) based on geological map of Alabama. The estimated density of EF-10 is low because the Brevard zone is the shear zone with low grade metamorphic rock. Another polygon EF-11 (D=2780, S=0.014) is interpreted as the Dadeville Complex of the Inner Piedmont. The high susceptibility of amphibolite and ultramafics of the Dadeville Complex generates magnetic high anomalies and a corresponding rise in gravity values

Similar to the interpretation of profile line CD, the Brevard fault and the Alexander City fault show prominent magnetic highs at 240 km. Profile EF has a thick body, EF-12 (D=2720, S=0.0004), below the meta-sedimentary rock (EF-3), which is interpreted as the crystalline metamorphic rock of the Grenvillian basement of Laurentia. This layer overlies the upper mantle layer (EF-13).

Discussion

Profile AB

The gravity and magnetic model along profile AB reflects three different tectonic terranes: the Wiggins terrane, the composite Suwannee terrane and the Uchee terrane on top of Laurentian basement (Figure 23). The origin of the Wiggins terrane is not clear. Thomas and others (1989) proposed the Wiggins terrane may be metamorphosed Suwannee terrane, or amalgamation of the Suwannee terrane with another terrane, or an entirely different terrane. Dallmeyer (1989b) suggested that the Wiggins terrane has a Laurentian origin based on mineral cooling ages of basement rocks on the Wiggins Arch. The low-density rocks of Wiggins terrane separates it from the high-density Suwannee terrane rocks in the gravity model. Although the origin of the Wiggins terrane is unclear, the boundary between Wiggins terrane and Suwannee terrane is clearly delineated in the model. The Wiggins terrane, therefore, can be described as a distinct tectonic element in the gravity and magnetic models (Figure 23). McBride and Nelson (1988, 1991) observed that the Moho depth decreases abruptly near the suture zone between ancient Laurentia and the Suwannee terrane. Similarly, the Moho depth along Profile AB is 35 km at 150 km distance and it increases to 40 km at 250 km (Figure 23). This change in depth to the upper mantle is interpreted to mark the tectonic suture zone in this location.

Fossils observed in the cores of lower to middle Paleozoic strata beneath the Coastal Plain indicate that the Suwannee terrane rocks have Gondwanan origin (Applin, 1951; Wilson, 1966; Barnett, 1975; Pojeta et al., 1976). A long-wave gravity high in the middle of the profile AB is interpreted to be caused by rocks of South Georgia basin and the Suwannee terrane. During the Mesozoic rifting of Pangaea, upwelling mantle created new, dense crust between the separating continents as extension created a graben system that was intruded by basaltic sills (Chowns and Williams, 1983; McBride and Nelson 1987; Thomas, 1988; McBride et al., 1989). The gravity high and magnetic high anomalies from the South Georgia basin and the Suwannee terrane are associated with these mafic intrusions (Figure 23).

To the northeast of the profile, the boundary between the Suwannee terrane and the Uchee terrane is marked by a gravity gradient. This gradient partly corresponds to the Terrane Boundary fault that marks the northern limit of the South Georgia basin (Guthrie and Raymond, 1992). To the north of Terrane Boundary fault, the Uchee terrane is interpreted as Peri-Gondwanan crust. This terrane is exotic to Laurentia and was thought to have been accreted to Laurentia before the beginning of Appalachian orogeny (Steltenpohl et al., 2008). High-grade gneiss, schist and migmatite of the Uchee terrane produce the gravity high observed there.

The magnetic and gravity low in Figure 23, between 260 and 300 km, corresponds to the BMA. Nelson and others (1985b) and McBride and Nelson (1988) interpreted BMA in Georgia as a Paleozoic suture between Laurentia and Gondwana. The hypothesis accepted by them about the origin of BMA is that it was formed as a consequence of Mesozoic rifting during the breakup of supercontinent Pangaea (McBride and Nelson, 1988). Since the rift-stage crust separated oceanic crust seaward from continental crust landward, the rifted area should include highly magnetic mafic intrusive rocks. The difference in susceptibility between these mafic rocks with the felsic volcanic rock and metamorphic rocks of Uchee terrane is ~0.054 SI unit causes the low-magnetic anomaly along BMA (Figure 23).

Profile CD

The gravity model along profile CD indicates a trend of decreasing gravity from A (south) to B (north). The modeling suggests the suture between relict Gondwanan crust and composite Peri-Gondwanan/Laurentian crust is located at 120 km in the model (Figure 24). To the south, the gravity and magnetic highs are interpreted to result from intruded mafic rocks within the South Georgia basin and the Suwannee terrane. North of the Terrane Boundary fault, a wedge of Peri-Gondwanan crust (Uchee terrane) compensates for the decreasing gravity values there. The Bartletts Ferry fault zone constitutes the Uchee terrane boundary in the north and separates the Uchee terrane rocks from the Pine Mountain window. The Pine Mountain window exposes Grenville basement and its metasedimentary cover (Clarke, 1952; Hooper et al., 1988; Steltenpohl et al., 2004). Using zircon dates of 2.2 Ga, Steltenpohl and others (2004) suggest that the Pine Mountain window may have derived from a non-Laurentia source (possibly Gondwana), but docked with Laurentia before the Alleghanian orogeny. A prominent gravity-low and a high magnetic gradient near the Bartletts Ferry fault zone delineate the boundary between Gondwanan crust/Peri-Gondwanan crust with Laurentian crust.

From the COCORP reflection data, Nelson and others (1987) estimated that Towaliga Fault offsets Grenville basement by ~ 9 km. The offset of the Grenville basement can be seen in the model CD (Figure 24). The Grenville basement rocks to the north of the Bartletts Ferry fault zone are thought to have Laurentian origin (Clarke, 1952; Wilson, 1966; Dallmeyer, 1989a). Different rock units in the Inner Piedmont and the Northern Piedmont province are introduced in the model based on the geological map of Alabama. Calculated gravity and magnetic curves derived from the density and magnetic susceptibility of these rock units matches well with the observed curves. Some major faults, such as the Goat Rock fault, Bartletts Ferry fault, Stonewall Line fault and Alexander City fault, are marked by strong magnetic gradients. Shallow fault usually show a steep gravity gradient and no magnetic anomaly if non-magnetic sedimentary rock is faulted. The sharp magnetic gradients and long-wave gravity gradients along the fault show that the high-susceptibility rocks of Piedmont are faulted in greater depth.

Continuation of magnetic expression related to the Piedmont and Valley and Ridge rocks to the south of fall line suggests that the Laurentian crust continues beneath the Coastal Plain sediments (Neathery et al., 1976; Horton et al., 1984; Steltenpohl et al., submitted). The southern boundary of BMA cuts northeast to southwest trending magnetic lineations within the Laurentian crust (Horton et al., 1984).

Strong magnetic gradients from Laurentian crust south of fall line, which are truncated by BMA suggest the tentative southern boundary of Laurentian crust. Based on the observed gravity change between Laurentian crust and Gondwanan crust from models AB and CD and the projection of Bartletts Ferry fault beneath the Coastal Plain observed in the magnetic map, the Uchee terrene is thought to extend from the southern boundary of BMA to the Bartletts Ferry fault (Figures 26 and 27). The Uchee terrane forms a wedge in Alabama and remains sandwiched between Laurentian and Gondwanan crust. In western Alabama, the Uchee terrane disappears and the suture line marks the boundary between Laurentian crust and Wiggins terrane (Figures 26 and 27).

Profile EF

The gravity and magnetic model along EF reflects the geologic units and structures exposed at the surface in northern Alabama (Figure 25). These include rocks of the Plateaus, the Valley and Ridge, the Northern Piedmont and the Inner Piedmont, which are all underlain by Grenvillian basement (Clarke, 1952; Wilson, 1966; Dallmeyer, 1989a). Thus, the entire profile EF is thought to lie on crust with a Laurentian origin (Figures 25, 26 and 27).

The Interior Low Plateau and the Appalachian/Cumberland Plateau consists of sedimentary rocks. The rocks of the Valley and Ridge province are sedimentary rocks of Cambrian to Pennsylvanian age, which were folded and faulted during late Paleozoic Appalachian suturing of Laurentia and Gondwana (Thomas, 1982; Neathery and Thomas, 1983; Raymond et al., 1988). Upper Neoproterozoic sedimentary rocks lie between the Paleozoic sedimentary rocks in Valley and Ridge and the Grenville basement (Cook et al., 1979; Hatcher et al., 1990; Steltenpohl, 2005).

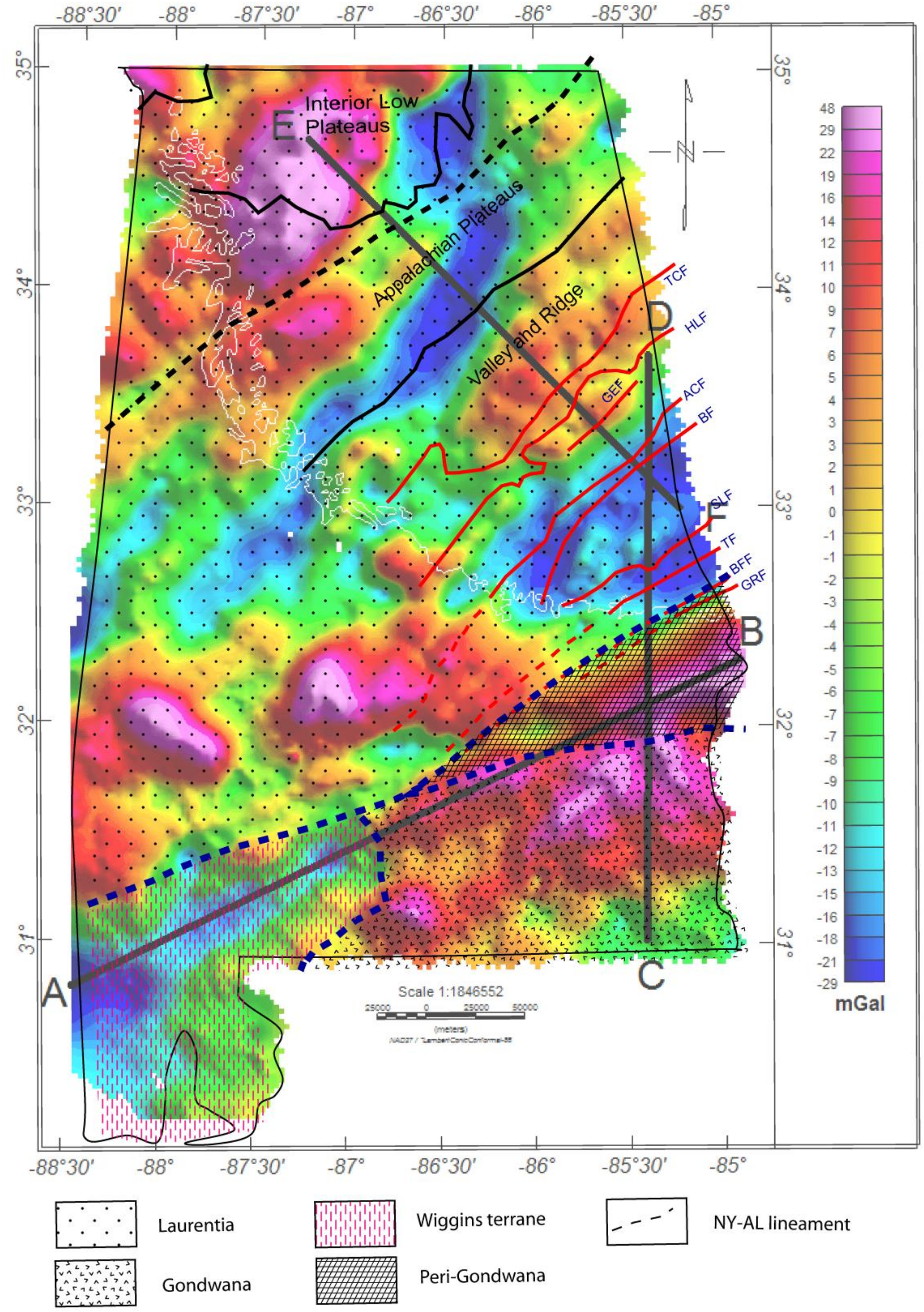

Figure 26. Map of gravity data showing the different types of crust beneath Alabama. Blue dotted lines show the suture between crusts of different origin such as Laurentia, Peri-Gondwana, Gondwana and Wiggins terrane. Lines AB, CD and EF are the profile locations along which gravity and magnetic modeling were performed. See figure 2 and 19 for the abbreviations of faults in blue letters. See text for more description.

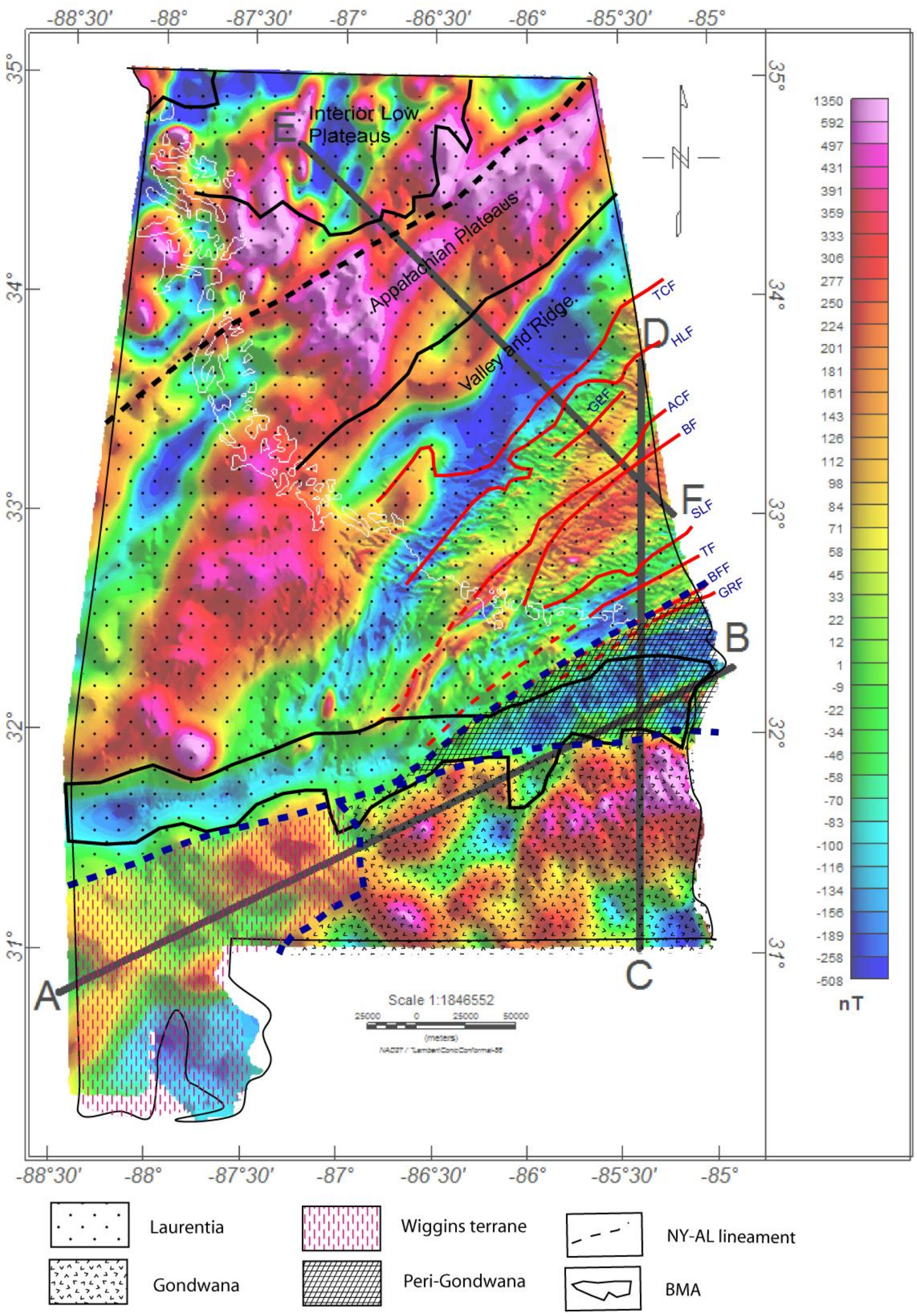

Figure 27. Map of magnetic data showing the different types of crust beneath Alabama. Blue dotted lines show the suture between crusts of different origin such as Laurentia, Peri-Gondwana, Gondwana and Wiggins terrane. Lines AB, CD and EF are the profile locations along which gravity and magnetic modeling were performed. See figure 2 and 19 for the abbreviations of faults in blue letters. See text for more description.

These sedimentary rocks and their associated basement faults are modeled well along profile EF (Figure 25). A gravity high at the northern end of the profile requires a high-density rock below the sedimentary rock of the Interior Plateau. A mafic rock body is embedded to match this gravity high, but the magnetic low there was not successfully matched by the model. This area represents the gravity high and magnetic low associated with the northwestern block of NY-AL lineament (Steltenpohl et al., 2010).

Steltenpohl and others (2010) suggest that the NY-AL lineament is a large strike-slip fault zone, which delineates a major crustal boundary beneath the Appalachian basin. The gravity and magnetic model along EF cannot identify the exact location of NY-AL lineament because the lineament is a large scale anomaly and only a part of it lies in northern Alabama. The lineament is positioned in the crustal model along profile EF based on its location in gravity and magnetic maps (Figures 5, 6 and 7). This lineament separates the magnetic-low and gravity-high rocks of the Plateaus from the magnetic high- and gravity-low rocks of the Valley and Ridge province.

The southeastern boundary of the Valley and Ridge province is marked by Talladega-Cartersville fault system. This fault separates the high-susceptibility Neoproterozoic through Early Paleozoic platform of eastern Laurentian preserved-the Valley and Ridge with relatively low-susceptibility platform rocks of Talladega slate belt. Therefore, the fault has a strong magnetic expression. The Hollins Line fault separates the Talladega block in northwest from the rest of the Northern Piedmont and the Inner Piedmont in southeast. Inner Piedmont rocks produce a gravity low that reflects the presence of low-density rocks in the Brevard fault zone. Bentley (1964) described the Brevard zone in Alabama as a lithologic break between the Inner and Northern Piedmonts.

This zone is marked by a zone of deformation and shearing that is rooted in the main sole thrust beneath the Northern Piedmont and Inner Piedmont (Hatcher, 1971; Cook et al., 1979). Gravity modeling in the Piedmont also supports the presence of low-density deformed rocks in Brevard fault zone.

Tectonic Significance

All three gravity and magnetic models combine together to provide information about the nature of different continental crusts in Alabama (Figure 26 and 27). The modeling shows that the thickness of crust in Alabama increases from south (40 km below Dothan, 35 km below Mobile) to north (44 km below Auburn). The modeling also explains that crust thickness increases from South-west Alabama (35 km below Mobile) to North-east Alabama (40 km below Dothan). In conclusion, the models reveal the thickening of crust toward north and North-east.

From the similar work in the Mississippi Coastal plain, Harry and Londono (2003) demonstrated that the Laurentian crust thins abruptly near the Paleozoic suture. This abrupt change of crustal thickness, and lack of synrift magmatism and normal faulting on the margin are interpreted as characteristics of a transform margin(Harry and Londono, 2003; Harry et al., 2004). In Alabama, the Gulf Coast in the southwest (below the Wiggins Arch) appears to be a continuation of a similar transform margin structure in Mississippi, Louisiana, and western Arkansas (Thomas, 1976, 1991). Abrupt change in crustal thickness and lack of mafic intrusive below in that area supports the interpretation of Thomas (1976, 1991) (Figure 28). From the result of the gravity and magnetic modeling across the Texas Gulf coast, Mickus (2009) interpreted the presence of volcanic rifted margin beneath the Texas Gulf coast, and they also suggested this volcanic margin gradually changes strike into a transform boundary beneath Louisiana Gulf coast. The Alabama Gulf coast does not show evidence for a volcanic rift margin, such as, a large deeply buried mafic igneous complex. However, basaltic flows and diabase sills of the South Georgia basin, document it as a Mesozoic rift (Chowns and Williams, 1983; McBride et al., 1989).

Disclaimer

Density and magnetic susceptibility contrast cross-sections are generated to test possible geometries causing gravity and magnetic anomalies. Interpretation of gravity and magnetic data is subject to some limitations. The first limitation is the uncertainty in the source of a given gravity and magnetic anomaly. Other limitations are the cases where there is no quantitative information of density and magnetic susceptibility contrasts between modeled polygons. Gravity and magnetics are sensitive to lateral density and magnetic susceptibility contrasts, and modeling will vary with size, shape, and depth. More intensive and careful research is required to gather all information of the subsurface so that models presented here present the most probable solutions, as many combinations of subsurface geometries can produce identical models.

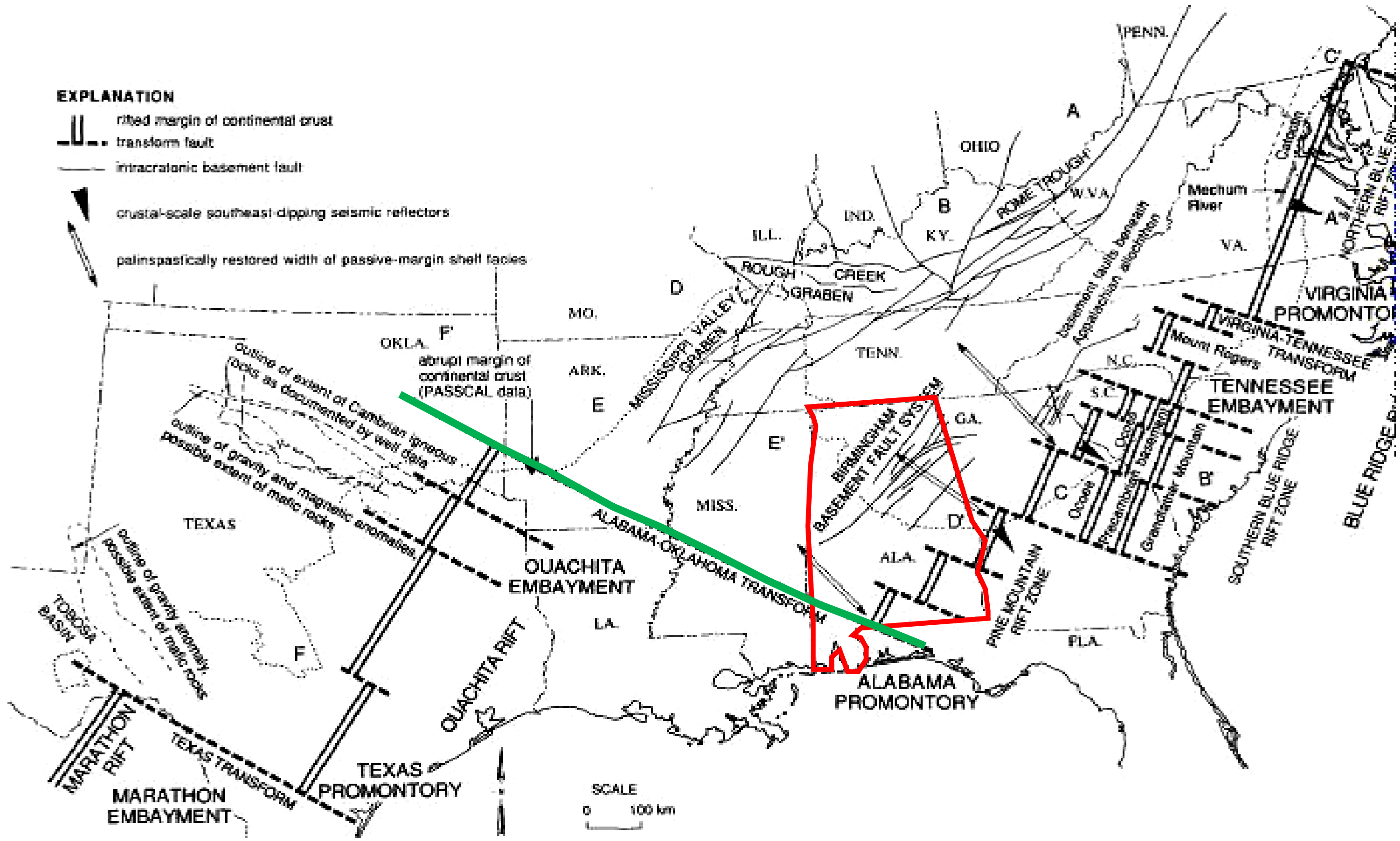

Figure 28. Generalized map of interpreted Late Precambrian to Early Paleozoic continental margin bounded by rift segments and transform faults (After Thomas, 1991). Red polygon is the State of Alabama. A transform margin structure (Alabama-Oklahoma transform, green line), which continues from Oklahoma to western Arkansas, Louisiana, Mississippi and southwestern Alabama, tectonically separates southwestern Alabama from the other part of the state.

Conclusions

Gravity and magnetic modeling along three transects in Alabama suggests that gravity and magnetic methods can be useful tools in tectonic and geological studies such as determining the location of sutures between different continental crusts covered by thick sediment packages, marking the continuation of faults hidden below the young sediments, determining the presence or absence of magmatism, and define the type of plate margin. This study compared the obtained results with the models derived from the COCORP transects in adjacent Georgia and other parts of Gulf Coastal plain (e. g., Mississippi, Louisiana, Texas).

Result of gravity and magnetic modeling indicates that the crust of the Wiggins terrane is a separate tectonic terrane. This study reveals the suture of the Wiggins terrane with Gondwanan crust and with Laurentian Crust. The Uchee terrane exhibits the different gravity and magnetic characteristics than the adjacent Laurentian crust. This study supports the hypothesis of Steltenpohl (2008) that the Uchee terrane is exotic to Laurentia and it has Peri-Gondwanan origin and was accreted to Laurentia during the later stages of the Appalachian orogeny. Based on the observation of change in crustal thickness along profile AB and CD, this study determines that change in Moho depth is usually common near the suture zone. This study interprets the southern boundary of BMA as the Late Paleozoic suture between Gondwana with Peri-Gondwana/Laurentia as proposed by Nelson and others (1985b) and McBride and Nelson (1988) in Georgia. In southwestern Alabama, the BMA serves as a suture zone between the Wiggins terrane and Laurentian crust. Due to the large scale of NY-AL lineament, gravity and magnetic modeling in this study cannot define the exact location of NY-AL lineament along the profile EF.

Some of the faults in Alabama and adjacent Georgia are fundamental to the tectonic evolution of the southeastern U. S. The Bartletts Ferry fault zone is the suture between the Peri-Gondwanan crust and Laurentian crust. The Towaliga fault separates the Pine Mountain window from Inner Piedmont, and it offsets Grenvillian basement by about 9 km. The Brevard fault zone is a major lithologic break between the Inner Piedmont with Northern Piedmont.

Gravity and magnetic modeling reveal the thickening of crust from south to north. The results of this study indicate continuation of the transform margin structure of Mississippi, Louisiana, and western Arkansas into southwest Alabama. Although there is no indication of a volcanic rifted margin like that along the Texas Gulf coast, Mesozoic rift basins are present in southeast Alabama.

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