Spring 2024

Geology of the Bishop 7.5-Minute Quadrangle and Geophysical Modeling to Constrain Syndepositional Deformation of the Black Warrior Foreland Basin

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GEOLOGY OF THE BISHOP 7.5-MINUTE QUADRANGLE AND GEOPHYSICAL MODELING TO CONSTRAIN SYNDEPOSITIONAL DEFORMATION OF THE BLACK WARRIOR FORELAND BASIN

A Master's Thesis

Presented to

The Graduate College of

Missouri State University

In Partial Fulfillment

of the Requirements for the Degree

Master of Science, Geography and Geology

By

Joseph Lane

May 2024

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Geography and Geology

Missouri State University, May 2024

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ABSTRACT

The Black Warrior basin is a foreland basin formed by the Ouachita–Appalachian orogeny during the late Paleozoic Era. Mississippian strata record the transition from the carbonate platform within the Tuscumbia Limestone to the clastic-rich units within the overlying Pride Mountain Formation and Hartselle Sandstone. The western and southern-most fringes of the Black Warrior basin are covered by thin, post orogenic Cretaceous sediments of the East Gulf Coastal Plain. Geological mapping of the Bishop 7.5-Minute quadrangle revealed subtle deformation in Mississippian strata in the form of gentle folds (<20°), possibly due to far-field stress from the incipient Ouachita orogeny, concentrated along structural boundaries. Late Mississippian clastic wedge deposition marks the onset of the Ouachita deformation within the Black Warrior basin. Low-magnitude gentle folding, possibly related to far-field stresses transmitted from the Ouachita orogeny, may have been concentrated along pre-existing structural boundaries. Erosion, following an early Cretaceous thermal uplift and subsidence of the Mississippi embayment, may have exploited structural trends which preserved deposition of the fluvial-deltaic, conglomeratic Cretaceous Tuscaloosa Formation. Modeling of the basal Cretaceous unconformity unveils a paleotopographic erosional surface that resembles the modern landscape, which parallels deformational orientation within the Mississippian strata. Regional gravity and magnetic data were used to investigate crustal structures within the subsurface. Geophysical mapping and modeling suggest regional trends in Late Mississippian sedimentation and deformation may be controlled by preexisting faults, variations in the density of crust, and crustal compositional heterogeneity, which may have been the result of Precambrian orogenic events or Iapetan rifting. This study implies tectonic inheritance is a defining feature not only at the margins of the continent, but within intracontinental basins and as it influences the sedimentary and topographical expression throughout time.

KEYWORDS: Black Warrior basin, Ouachita orogeny, foreland basin, geological mapping, geophysical modeling

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OVERVIEW

The Middle to Late Mississippian rocks of the Black Warrior foreland basin represent the transition from a carbonate dominated ramp to siliciclastic wedge, which was deposited during the initial stages of deformation induced by the Ouachita orogeny. The basin forms a homocline of Paleozoic strata that dips to the southwest and formed in the syntaxial bend of the Ouachita-Appalachian Orogeny as the orogenic fronts converged on the southern margin of the Laurentia continent. 1:24,000 scale geologic mapping was undertaken to better delineate the stratigraphic contacts in northwestern Alabama and northeastern Mississippi. Mapping revealed subtle deformation and paleotopography within the variable Paleozoic strata and paleo-relief in the basal Cretaceous unconformity. Erosion during the Late Cretaceous found preferential pathways along the deformational structures int the Paleozoic strata. The mapping and modeling of the Cretaceous unconformity suggests preferential erosion may be repeated in the modern fluvial landscape. Gravity and magnetic grids compiled from databases for the conterminous United States were used to construct geophysical maps and preliminary 2D forward models. Modeling of crustal structures revealed faults in the magnetic basement and necessitated anomalously dense crust to produce geologically feasible models. The Bouguer gravity anomaly and reduced-to-poles (RTP) magnetic anomaly maps were compared with regional isopach trends in the Mississippian clastic units of the Black Warrior basin, which reveal that antecedent basement faults and crustal structures characterized by maps and models may have influenced a structural control on Middle to Late Mississippian sedimentation. The flexure along the basin likely reactivated pre-existing faults and compositional differences between structures within crustal blocks in response to regional changes in the stress field. The migration of the orogenic front altered the rate of subsidence, which impacted the topography of the region and thus the sedimentation.

PART 1: GEOLOGY OF THE BISHOP 7.5-MINUTE QUADRANGLE, COLBERT COUNTY, ALABAMA AND A PORTION OF TISHOMINGO COUNTY, MISSISSIPPI

ABSTRACT

The Bishop 7.5-minute quadrangle in western Colbert County, Alabama and eastern Tishomingo County, Mississippi is located within the Highland Rim province of the Interior Low Plateau and the Gulf Coastal Plain. Paleozoic bedrock in the quadrangle is composed of sedimentary rocks of the Black Warrior basin, a Carboniferous foreland basin formed by the onset of the Ouachita Orogeny within the syntaxial bend of the respective Ouachita and Appalachian Orogenic front. These rocks have undergone both syndepositional and post depositional deformation due to subsidence and migration of the flexural bulge caused by the encroaching orogenic fronts. Within the quadrangle exposed Paleozoic rocks are Mississippian in age. Mesozoic age sediments of the Gulf Coastal Plain directly overlie rocks of the Mississippian age; the unconformable surface represents paleotopography of the region prior to the Cretaceous Coastal Plain sediments.

Formal stratigraphic units mapped within the quadrangle include the Tuscumbia Limestone, the Pride Mountain Formation, the Hartselle Sandstone, the Tuscaloosa Group, and undifferentiated Quaternary aged alluvium and low terrace deposits.

Mississippian strata on the Bishop quadrangle preserve a sequence of prograding clastic facies and onlapping carbonate facies as the basin experienced flexural subsidence and clastic deposition at the onset of Ouachita tectonism. Mississippian strata dip 3-14° and generally strike EW or NW-SE. The Black Warrior basin is a homocline that dips <1° to the southeast. Rocks are subtly folded into gentle, low angle (<20° axial hinge angle) anticline-syncline pairs. Minor folds within the Pride Mountain Formation are located on the eastern portion of the quadrangle, which continue onto the adjacent Barton quadrangle. A fold striking predominantly EW is found near the northwestern portion of the map and is located within the Tuscumbia Limestone and Pride

Mountain Formation. A fold that deforms the Pride Mountain Formation and Hartselle Sandstone is present in the southeastern portion of the quadrangle striking NW-SW. The NW-SE and E-W trending folds are parallel to the inferred local basement structures and likely formed as the Ouachita orogenic front induced compression from the southwest.

The Cretaceous Tuscaloosa Group is the only member of the Gulf Coastal Plain present on the Bishop quadrangle. It unconformably overlies Mississippian strata and records fluvial-deltaic clastic deposition along the southern margin of North America and the emerging Mississippi embayment. This unconformity caused, by a large drop in base level, created a high degree of differential topography including a major incised valley system prior to the Cenomanian. The erosional surface represented by the basal Cretaceous unconformity was approximated using outcrop data from the quadrangle and the surrounding area to create a paleo-topographic reconstruction at the time of deposition of Cretaceous age sediments.

INTRODUCTION

The Bishop quadrangle is located within southwestern Colbert County, Alabama with a small portion within southeastern Tishomingo County, Mississippi (see Appendix). The quadrangle is located in a predominantly rural area with the unincorporated communities of Allsboro and Maud, Alabama. Approximately one mile to the north is U.S. Route 72 that connects the nearest population centers of Iuka, Mississippi and Cherokee, Alabama to higher density population regions of Corinth, Mississippi to the west and the Florence-Muscle Shoals Metropolitan Area to the east. Bear Creek flows from the southwest portion of the map to the north to enter Pickwick Bay on the Margerum quadrangle and is managed by the Tennessee Valley Authority. Pickwick Bay is an extension of the Tennessee River which is located to the north on the Waterloo quadrangle and to the northwest on the Cherokee quadrangle. To the north on the Margerum quadrangle is an active railroad running parallel to U.S. Route 72 operated by Norfolk Southern. The quadrangle contains both active and inactive quarries with the primary products being crushed limestone, gravel aggregate, and historically rock asphalt. The Natchez Trace Parkway bisects the Bishop quadrangle northeast to southwest as part of the National Park System. To provide accurate and current geologic data for the construction and maintenance of the Natchez Trace Parkway by the National Park Service, geologic mapping of the Bishop quadrangle was conducted at the 1:24,000 scale. The objective of this study was to determine the distribution and thickness of stratigraphic units and describe the lithologies exposed in the Bishop quadrangle. This chapter is a summary of the efforts and investigations of this region to produce new geologic maps where basic geologic data are needed to support updates to infrastructure.

LOCATION

The Bishop 7.5-quadrangle (lats. 88°00’ 00” and 88°07’30”; longs. 34°37’30”and 34°45’ 00”) is located in southwestern Colbert County, Alabama and Tishomingo County, Mississippi. The elevation of the Bishop quadrangle ranges from approximately 400 feet to 800 feet above sea level. The location is shown on the physiographic map in figure 1. The quadrangle is located within the Highland Rim of the Interior Low Plateau and the eastern Gulf Coastal Plain physiographic regions (Sapp and Emplaincourt, 1975). The quadrangle is characterized by valleys with wide, low floodplains and with broad ridges. A north-south oriented valley located on the western portion of the map is underlain by limestone containing alluvium and low terraces. This valley bifurcates near the southern portion to the quadrangle where Bear Creek enters from the Tishomingo quadrangle to the west and is joined by Rock Creek from the Pleasant Site quadrangle to the south.

Bear Creek drains into Pickwick Bay on the Margerum quadrangle to the north where the Tennessee Valley Authority operates a small corridor along Bear Creek. The western and central ridges form subtle cuestas composed of interbedded sandstone, limestone, and shale of the Pride Mountain Formation and Hartselle Sandstone. These ridges are capped by a resistant conglomerate with numerous smaller valleys and creeks incising along the eastern and western slopes. The southern portion of the Bishop Quadrangle has a ridge separated from the central ridge by a smaller valley that contains Chandelower Creek which flows west to meet Rock Creek as it enters from the Pleasant Site quadrangle. The western, central, and southern ridges are forested and are the main location for the Freedom Hills Wildlife Management Area, which is provided by the Wildlife Section of the Division of Wildlife and Freshwater Fisheries, Alabama Department of Conservation and Natural Resources, in cooperation with the Forever Wild Land Trust. The Natchez Trace Parkway enters from the northeast portion of the quadrangle where it then crosses over the central ridge to join the Bear Creek Valley to run parallel with Bear Creek where it exits to the Tishomingo quadrangle to the southwest. The Natchez Trace Parkway is managed and controlled by the National Park Service. The valleys and low-lying regions are primarily agricultural land containing farmland and pastures.

GEOLOGIC SETTING

The Paleozoic rocks of the Black Warrior basin are unconformably overlain by Cretaceous sediments of the Gulf Coastal Plain (Thomas, 1972b). Paleozoic rocks are Mississippian in age and part of the Black Warrior basin, a foreland basin formed in the syntaxial bend of the Ouachita—Appalachian Orogeny (Thomas, 1989). It is bound to the southwest in the subsurface by the faulted thrusted crust of the Ouachita front, to the southeast by the folded thrust belt of the

Appalachian Mountains (Mellen, 1947; Thomas, 1976, 1988), and to the north by the Nashville Dome (Welch, 1958; Thomas 1972a). The basin is a homocline that dips <1° to the southeast (Thomas 1972a). The input of clastic sediments recorded in the Mississippian rocks of the Bishop quadrangle represents a shift on the stable Alabama promontory from a passive margin to an active foreland basin as tectonic flexure induced subsidence caused by the obduction of an accretionary wedge to the southwest as the Sabine terrane collided with Larentia (Thomas 1976; Keller et al., 1989; Mickus and Keller, 1992). Paleozoic bedrock was buried beneath the Gulf Coastal Plain sediments of the Mississippi Embayment during the late Cretaceous, as the region subsided following thermal uplift due to the Bermuda Hotspot as it passed under the Mississippi Valley graben and the Reelfoot rift (Cox and Van Arsdale, 1997, 2002).

PREVIOUS INVESTIGATIONS

Hilgard (1860) was among the first published investigations of the area, commenting on the Paleozoic rocks. Smith et al. (1894a) produced the first geologic map of Alabama at the state level to include the Bishop quadrangle. A report of the Tennessee Valley region included Colbert County, Alabama describes the Paleozoic strata and included outcrops located on the Bishop quadrangle (McCalley, 1896). Adam et al. (1926) continued the work of prior geological mapping efforts and further defined stratigraphic relationships, including a summary of Paleozoic (Butts, 1926) and Mesozoic (Stephenson, 1926) stratigraphy and paleontological observations. Morse (1928) published a report of the Paleozoic rocks of Tishomingo County, Mississippi, which was expanded to the surrounding areas in Mississippi and Alabama, providing detailed stratigraphic and paleontological information for the outcrops of Mississippian age rocks located on the Bishop quadrangle (Morse, 1930). Welch (1958) produced stratigraphic charts of the Mississippian outcrops in Colbert County, Alabama and Tishomingo County, Mississippi. Welch (1959) expanded this work to include corresponding outcrops and subsurface stratigraphy of the Black Warrior basin to the south. A detailed map of the geology of Colbert County, Alabama was published in 1962 at the 1:63,360 scale (Moore and Harris, 1962). A report on the geology and ground-water resources of Colbert County, Alabama provided the regional distribution of rocks by the use of well logs to assess the thickness and lithology of the geology (Harris et al., 1963). The most recent summaries of stratigraphy and geologic mapping to include the Bishop Quadrangle were based on the synthesis of data by Szabo et al. (1988), Osborne et al. (1989), and Raymond et al. (1988). A survey of the geology and mineral resources of Tishomingo County, Mississippi was completed by Merrill et al. (1988). The most recent geologic mapping efforts at the 1:24,000 scale are the Pride (Szabo, 1975) and the Tishomingo quadrangle, including a portion of the Bishop quadrangle (Merrill et al., 1988). The adjacent Margerum and Cherokee quadrangles are also in the process of being mapped.

STRATIGRAPHY

The surface geology of the Bishop-7.5 minute quadrangle is composed of Paleozoic rocks of Mississippian age that are unconformably overlain by Mesozoic sedimentary rocks of Cretaceous age (Thomas, 1972a), with quaternary sediments deposited by the modern streams. Formal stratigraphic units present on the Bishop quadrangle (Fig. 2) include the Tuscumbia Limestone; Pride Mountain Formation; Hartselle Sandstone; Tuscaloosa Formation. Cretaceous Sediments of the Tuscaloosa Group are primarily fluvial deltaic in origin and the basal Cretaceous unconformity preserves paleotopography within the Paleozoic strata suggesting it was deposited in the streams and valleys of karst topography and low ridges, akin to modern valley fill deposits of the western Highland Rim (Marcher and Stearns, 1962).

MISSISSIPPIAN

The Mississippian age clastic and carbonate sediments of the Bishop quadrangle comprise a predominantly clastic wedge that overlies primarily carbonate rocks of the lower Mississippian to Cambrian, which suggests that this region on the Alabama promontory has been a persistent passive margin shelf (Thomas 1972b, 1976, 1988). The clastic wedge of Middle to Upper Mississippian rocks are cyclical shales, sandstone, and limestone and marks the initial input of clastic sediment into the Black Warrior basin (Thomas, 1972a; 1995). These sediments thicken to the southwest, though the Bishop quadrangle is positioned along the northern erosional limits of these units. The clastic wedge pinches out to the northeast between the Bangor Limestone and the Monteagle Limestone (Thomas 1972a, 1988; Thomas and Mack 1982) and to the southwest between the Bangor Limestone and the Neal Shale (Thomas, 1972a; Kidd 2008). There have been multiple interpretations of the source and origin of these rocks though typically thought to comprise deltaic to shallow marine environments controlled by sea level, tectonic, and climatic factors (Pashin, 1993; Pashin and Rindsberg, 1993).

TUSCUMBIA LIMESTONE

The Mississippian Tuscumbia Limestone is the oldest stratigraphic unit exposed on the Bishop quadrangle. It overlies the Fort Payne Chert in a gradational contact between the cherty limestone beds of the Tuscumbia Limestone and the predominantly chert beds of the Fort Payne Chert (Thomas, 1972a). Smith et al. (1894b) used the name in relation to upper limestone of the Fort Payne Chert (Smith, 1890). The Tuscumbia Limestone is contemporaneous to the St. Louis Limestone to the north (Smith, 1892). The type locality of the Tuscumbia Limestone is Tuscumbia, Colbert County, Alabama (Smith et al., 1894b). Smith et al. (1894b) estimated the thickness of the Tuscumbia Limestone to be 20-150 feet and described it as having a gentle undulating topography, dividing the Fort Payne Chert into the Tuscumbia, synonymous with the St. Louis limestone, and the Lauderdale limestone. It was named the Tuscumbia Limestone in Smith et al. (1894b) for exposures at the town of Tuscumbia, Alabama, its stratigraphic limits were restricted to include all strata above the Lauderdale cherty limestone and below the Oxmoor Shale and Sandstone and Bangor Limestone. McCalley (1896) referred to the Limestone as either the Tuscumbia or St Louis Limestone from limestone in the Mississippi Valley and identified the thickness of the limestone in this region to be 125 to 175 feet. Butts (1926) related the St Louis Limestone of the Mississippi Valley and Tuscumbia Limestone and Warsaw Limestone of Alabama. Later the St. Genevieve Limestone, a thin, oolitic, and discontinuous limestone, was added that overlies the Tuscumbia Limestone. The Tuscumbia Limestone contains bryozoans, crinoids, brachiopods, echinoderms, and corals (Butts, 1962; Thomas, 1972a; Rodriguez and Kopaska-Merkel, 2014). An outcrop demonstrating the massive bedding and chert nodules is shown at an exposure along U.S. Route 72, on the Margerum quadrangle (Fig. 3).

The Tuscumbia Limestone is exposed along the streams and valleys of the Bishop quadrangle at low elevations and underlies much of the quadrangle. A drillers log from a well on the Bishop quadrangle provided a detailed stratigraphic report that included the description and depth of subsurface units from Mississippian to the Ordovician. The thickness of the Tuscumbia Limestone at this location, NW ¼ sec. 10, T. 4 S., R. 15 W., is approximately 250 feet. A small basal limestone bed of the Pride Mountain formation is also present near the top of the Tuscumbia Limestone in some areas (McGlamery, 1955). The Fort Payne Chert is approximately 160 feet in this location, however, due to the gradational nature of the contact between the Fort Payne Chet and the Tuscumbia Limestone, the thickness may be closer to 150 to 200 feet, as approximated by Thomas (1972a). The Tuscumbia Limestone has a consistent description: a light gray micrite to light gray coarse bioclastic limestone with massive beds, generally more than one foot thick. The bioclastic limestone may contain oolites and often abundant crinoids on the Bishop quadrangle. Thin beds of fine grain calcarenite may also be present in the upper portion of the Tuscumbia Limestone on the adjacent Margerum quadrangle. Light gray chert nodules may be several inches in thickness and several feet in length. The Tuscumbia Limestone is exposed on a large cliff along the western bank of Bear Creek (NE¼SE¼ sec. 10, T. 4 S., R. 15 W). Caliche found within the Tuscumbia Limestone suggests there may have been subaerial exposure following the deposition and prior to the Pride Mountain Formation (Fisher, 1987). Driese et al. (1994) found evidence of regional paleotopography on the Monteagle Limestone to the west, which may be contemporaneous with the exposure of Tuscumbia Limestone and deposition of the Pride Mountain Formation, further arguing for erosion and development of topography within the Tuscumbia Limestone. Exposures of stratigraphic contacts at the Allsboro Quarry display the erosional surface within the Tuscumbia and unconformity signifies deposition of the Pride Mountain within the topographic lows. Pashin (1993) found that the Tuscumbia Limestone of the upper ramp contains interfingering of calcarenite and oolitic limestone, which suggests it was a carbonate sand bank with isolated tidal bars or islands. The lower ramp may have been fault controlled, which was hypothesized to have formed during a period pre to early Cambrian extension and reactivated during the Mississippian.

PRIDE MOUNTAIN FORMATION

The Pride Mountain Formation overlies the Tuscumbia Limestone and underlies the Hartselle Sandstone (Thomas, 1972a). It was defined as a sequence of thick shale with alternating siltstone, limestone, and sandstones (Welch, 1958). Thickness may range from 150 to 280 ft in counties to the south of the Bishop quadrangle (Thomas, 1972a). Hilgard (1860) was the first to recognize the alternating and laterally discontinuous sandstones and limestones of the Pride Mountain Formation along the northern eastern edge of the Bishop quadrangle, however it was referred to simply as the Carboniferous Formation. Clark (1925) listed asphaltic portions of the limestone to measure approximately 10 feet in thickness, with overlying fossiliferous limestone, clay, shale, or mineralized limestone directly in contact with the Tuscumbia Limestone. The crinoidal asphaltic limestone of the Pride Mountain Formation were mined in the early 1920's in Colbert County and the northern portion of Bishop quadrangle (Sec. 1, Sec. 12, T. 4 S., R. 15 W.) (Clark, 1925). Butts (1926) recognized units of sandstones and shales within northwestern Alabama as the Bethel Sandstone, the Gasper formation, Cypress sandstone, and the Golconda formation, named after locations in Kentucky and Illinois, though was attributed stratigraphically to the Hartselle series.

Based on a survey of Tishomingo County, Morse (1928) further separated and redefined the rocks of the Mississippian, doing away with the prior nomenclature and developing new formation members previously unrecognized formally. The Hartselle Sandstone was abandoned in favor of the unconformity separating the Chester series of this age with the underlying strata containing a basal limestone due to biostratigraphic relationships. New stratigraphic names included the Alsobrook formation, Allsboro sandstone, Southward Pond formation, Southward Spring formation, Southward Bridge formation, and the Forest Grove Formation. Welch (1958) was the first to formally recognize the Pride Mountain Formation. The type locality was named for exposures near Pride Mountain in central Colbert County, approximately 10 miles east of the Bishop quadrangle and half a mile southwest of the town of Pride along U.S. Route 72. This new naming scheme recognized all strata between the underlying Tuscumbia Limestone and overlying Hartselle Sandstone to include the Pride Mountain Formation based on the correlation of outcrop data. The stratigraphic divisions proposed by Butts (1926) and Morse (1928) were revised. The St. Genevieve Limestone of Butts (1926) was included into the Alsobrook member of Morse (1928). The newly proposed Tanyard Branch replaced the Bethel Sandstone and Gasper Formation of Butts (1926) was divided into to include the Wagnon member, and the Southward Spring member, the Sandfall member. Additionally, the Mynot sandstone member and the Green Hill member were introduced to replace the Cypress and Golconado formations described by Butts (1926). The previous stratigraphic scheme of Welch (1958) was applied to oil and gas wells in northwestern Alabama and northeastern Mississippi and correlated to the Lewis and Evans Sandstones used by the oil industry (Welch 1959).

The Pride Mountain represents the preliminary input of clastic sedimentation into the Black Warrior basin (Thomas, 1972a, 1972b, 1974; Pashin and Rindsberg, 1993). It marks the transition from a carbonate ramp system of the Tuscumbia Limestone to a sediment starved basin (Pashin, 1993). The depositional environments in the Pride Mountain interval are interpreted as deltaic (Cleaves and Bat, 1988; Stapor and Cleaves, 1992), shallow marine (Thomas, 1972a; Kidd, 2008), shelf (Higginbotham, 1986), and shore zone to beach environments (Pashin and Kugler, 1992; Pashin and Rindsberg, 1993). The thickness of the Pride Mountain Formation is estimated to be between 150 to 300 feet (Welch, 1958, 1959; Thomas 1972a) based on well log and outcrop correlations, though it varies substantially and abruptly. The units are not laterally extensive (Moser and Thomas, 1967). On the Bishop quadrangle the sandstones, shale, and limestone outcrops demonstrate a variability in the lithology and thickness, and lateral continuity. Pride Mountain is composed of interbedded shale, limestones, and sandstones, but due to the variable nature of these intraformational members and lack of subsurface data it will be undifferentiated for the purposes of mapping.

Thomas (1972a) describes the boundary between the Tuscumbia Limestone and the overlying Pride Mountain Formation in Colbert as a gray shale with a distinct thin limestone that is absent of chert, but may contain interbedded oolitic, shaly, sandy layers (Thomas, 1972a) due to subaerial erosion of the Tuscumbia Limestone produced in shallow depositional environment. It has been identified by the brachiopod *inflatia inflata* (Butts, 1926; McGlamery, 1955; Welch, 1958; Thomas, 1972a; Pashin and Rindsberg, 1993). The thickness ranges between 4 to 50 feet (Thomas 1972a; Pashin and Rindsberg, 1993), and found abundant in outcrops, not laterally continuous in core from the surrounding area (Pashin, 1993). Oolitic beds may also represent the basal Pride Mountain Formation interpreted as deposition during a transgression. These shallow deposits may have formed during storms with lenses of asphaltic quartzarentite bodies encased in clay shale as shelf sand bodies that were reworked within the mesotidal shore zone (Pashin and Kulger, 1992).

The Lewis Sandstone is very fine to medium grain quartzarenite, may locally contain echinoderms, brachiopods, bryozoans, and asphalt. Sedimentary structures preserved include hummocky cross bedding, wave ripples, and both high and low angle crossbeds, with scours, fill structures, horizontal laminae, and current ripples (Pashin and Rindsberg, 1993). May be interbedded with limestone, or sandstone that may be calcareous or argillaceous in parts (Thomas, 1972a). The sandstone is exposed at N½ sec. 11, T. 4 S., R. 15 W. (Fig. 4).

The Lewis Limestone replaces the middle sand unit of the Pride Mountain Formation on the Bishop quadrangle (Fig. 5) (Thomas, 1972a). It represents transgressive carbonate sedimentation in an intertidal to marine depositional setting. With this intraformational sandstone topography again seems to be a defining feature of the sedimentation as there is a great variability among outcrops (NE½ sec. 13, T. 5 S., R. 15 W. (Fig. 4).). (Pashin and Kugler, 1992; Pashin and Rindsberg, 1993). Locally, it may transition between a calcarenite to micrite packstone and may show signs of being intensely bioturbated and bioclastic with abundant brachiopods, bryozoans, ooids (Pashin and Rindsberg, 1993).

The Evans Sandstone is the upper sandstone of the Pride Mountains Formation. It is a fine grained quartzarentite (Higginbotham, 1986) and may be argillaceous in portions. A basal bed of channel filling limestone with a bioclastic pebble conglomerate may be present. It obtains a thickness of approximately 25 feet in Colbert County (Thomas, 1972a).

Shale is present between these named sandstone and limestone units (Fig. 6). It contains brachiopods and bryozoan fossils typical of shallow marine environments, as well as minor sandstone units that may contain wavy, flaser, and lenticular bedding and be locally calcareous. Preserved sedimentary structures include current ripples, wave ripples, mudcracks, and siderite nodules (Thomas, 1979). It may also contain fossil traces such as burrows and trails, bivalves, root traces, and plant fossils. Shale units have been interpreted as a swampy wetland or intertidal setting (Pashin and Rindsberg, 1993) or lagoonal (Thomas, 1972a). Shale units are exposed at SE¼ sec. 10, T. 4 S., R. 14 W., SE¼ sec. 1, T. 4 S., R. 15 W. (Fig. 6).

Multiple complex environments are represented by the sandstones with rippling, crossbedding, bioturbation, localized channel-fill conglomerates, evidence of marine reworking, in addition to being bound by sandy, oolitic, and/or bioclastic limestone and thin calcareous shale thin suggests an offshore shallow marine bars or intermittent low barrier islands (Thomas, 1972a, 1979). The onset of sedimentation was initiated by the subsidence due to the Ouachita orogenic thrusting to the southwest (Whiting and Thomas, 1994). Thrusting induced the formation and migration of a flexural bulge craton ward (Ettensohn and Pashin, 1993; Hines, 1988; Beaumont et al., 1987).

The origin of the clastic sediments that comprise the Pride Mountain Formation is uncertain. Deltaic depositional originating to the north systems (Cleaves and Broussard, 1980; Cleaves, 1983) was corroborated by petrographic analysis (Cleaves and Bats, 1988) and subsurface mapping of incised river channels suggest a cratonic source from southward flowing rivers draining the craton (Stapor and Cleaves, 1992). The erosional surface of the Tuscumbia Limestone formed by shoaling deposition of the Pride Mountain Formation sequentially filling the paleotopography in the leeward side based on paleoclimatic factors may also suggest a northern cratonic source (Pashin and Rindsberg, 1993). An alternative origin of the Mississippi clastic wedge is the transport of sediments from a southwestern source onto the cratonic shelf from the evolving Ouachita orogenic uplift created a distributary network that prograded to the northwest within the nascent Black Warrior (Thomas, 1974; Higginbotham, 1986; Thomas, 1988).

HARTSELLE SANDSTONE

The Hartselle Sandstone is a predominantly massive bedded quartzose sandstone (Fig. 7). It is the thickest and most laterally persistent of the Mississippian clastic series of the Black Warrior basin. It overlies the Pride Mountain Formation, underlies the Bangor Limestone, and to the southwest it pinches out into the Floyd Shale (Thomas, 1972a). The Hartselle Sandstone was initially known as the Lagrange Sandstone (Smith, 1879), but was later referred to as the Oxmoor Shale and Sand (Smith, 1890). Smith et al. (1894b) describe the Hartselle Sandstone as capping mountainous spurs east of Huntsville, Alabama. It was formally recognized as the Hartselle Sandstone by Smith et al. (1894b), initially grouped with the Bangor Limestone as part of the Mountain Formation. The Hartselle of McCalley (1896) did not differentiate the interbedded sandstone, limestone, and shale of the Pride Mountain Formation and the massive, bedded Hartselle Sandstone. Butts (1926) recognized the Hartselle Sandstone was marked by a thick-bedded, medium-grained sandstone that transitions to thinly bedded with interlayered shale in the upper portion, which separated it from the Bangor Limestone, which locally contains brachiopods, bryozoans, blastoids, and fossilized stumps.

The Hartselle Sandstone is defined as predominantly clastic sediments above the Pride Mountain Formation and below the Bangor Limestone (Thomas, 1972a). It is thickest along the outer edge of the Tuscumbia ramp, forming a series of elongated lenses in a narrow discontinuous southeast trending isolith maximum that extends to the southeast and pinches out to the southwest (Thomas 1972a, 1979). The thickness of the Hartselle Sandstone ranges from 100 feet to 150 feet (Welch, 1958; Harris et al., 1963; Thomas, 1972a, 1979), though on the Bishop quadrangle only the base of the unit is present.

The Hartselle Sandstone is a light colored, fine to medium grained quartzose sandstone that is typically well sorted (Thomas, 1972a, 1979; Mack et al., 1981). There is some variability in lithology with calcareous regions and the presence of pebble beds. It contains six internal facies that can be distinguish by criteria of two distinct subsets: a massive, ledge forming, fine grained sandstones, and thin, fine-grained, clay and mud dominated facies (Thomas and Mack, 1982; Kidd, 2008). These facies occur in large lenses where local channel fillings may be present (Thomas 1972a, 1979). Sedimentary structures include horizontal laminae, cross beds, and ripples. Sections of thin ripple bedded sandstone, interlaminated claystone, and thick calcareous shale increase near the contact with the overlying Bangor Limestone, as can be seen on the Barton quadrangle to the east. A massive cross bedded exposure can be found at NE ¼ sec. 32, T. 5 S., R. 14 W., with thinner bedded interlaminated claystone facies found at SE ¼ sec. 18, T. 5 S., R. 14 W (Fig. 7).

The Hartselle Sandstone pinches out to the northeast in Jackson County, Alabama, where it overlies the Monteagle Limestone (Thomas, 1972a) and thins into lenses of alternating quartzose sandstone and bioclastic limestone. It overlies the Pride Mountain Formation, with a gradational contact that is undulatory, separated by a calcareous shale. The Hartselle Sandstone pinches out to the southwest further into Mississippi and transitions to the Pearce Siltstone (Thomas, 1972a; Kidd, 2008).

The Hartselle Sandstone is interpreted to have formed in a massive, northeast facing barrier island and shelf bar complex (Thomas, 1972a; Thomas and Mack, 1982; Higginbotham, 1986) The orientation of sedimentary structures suggests the shoreline was oriented northwest to southeast and local channel fills indicate tidal channels bisected the barrier islands. Marine invertebrate fossils accumulated during storms that prograded the barrier island complex northwest across the carbonate ramp of the Monteagle Limestone (Thomas, 1972a). It grades upwards into a carbonaceous mud and the Bangor Limestone signaling a marine transgression occurred (Thomas and Mack, 1982).

Cleaves and Broussard (1980) suggested an elongated delta distribution system supplying sands from the northwestern Hartselle distribution from the north through the Illinois basin or from craton to northwest. Stapor and Cleaves (1992) presented paleocurrent data from exposed facies to provide that the Hartselle was deposited during a period of transgression that shifted the shoreline northward from deltaic systems originating from northwest, which would explain the presence of it overlying the Monteagle Limestone to the east (Driese et al., 1994). This differs from the extensive and narrow trend of thick sand present from isopach maps (Thomas, 1972a) that suggests it prograde northeastward from a southwestern Ouachita derived source (Thomas, 1976; Mars and Thomas, 1999; Whiting and Thomas, 1994). It also conflicts with stratigraphy and depositional history further to the east (Thomas and Mack, 1995).

A petrographic comparison of the Hartselle Sandstone with the Pennsylvanian Parkwood Sandstone and sandstones from the Illinois basin found similarities in composition of the Hartselle and Parkwood sandstones, but differ in the relative grain type abundance, and both differ in the populations from of the Illinois basin. This difference was attributed to the depositional environments, as the barrier island system of the Hartselle Sandstone would experience intensive reworking that would cause the more mechanically strong grains to be preserved. The sandstone of the Illinois basin suggests a differing source due to the presence of volcanic grains and polycrystalline quartz in the Hartselle, requiring a source of low-grade metamorphic origin which may be currently buried beneath the Gulf Coastal Plains. The Hartselle Sandstone and Late Mississippian to Early Pennsylvanian Parkwood Sandstones were derived from similar source terrane, however the Hartselle differs petrographically from sandstones in the Illinois basin (Mack et al., 1981), which was interpreted to indicate sediments from the eastern Granite Rhyolite province was an unlikely source for the sediments that would comprise the sandstones of the Black Warrior basin.

Recent detrital zircon U-Pb age analysis aided by petrographic and geochemical analysis continue to debate the origin of Hartselle. Sandstone analysis and U-Pb detrital zircon geochronology invoked input of a northern Appalachian source through a major drainage coincident with the Mississippi Valley graben (Xie et al., 2016). In an integrated analysis of petrography, XRF geochemistry, and U-Pb detrital zircon geochronology interpreted a large input of sediments from northeast of the Black Warrior basin from recycled grains derived from the Appalachian front to the northeast (Gifford et al., 2020). An intercontinental drainage system might have been present by the Pennsylvanian that could have routed sediment into the Ouachita Basin from the northeast as suggested by the analysis of detrital zircon U-Pb ages and Hf analysis (Allred and Blum, 2021). However, this would require the Ouachita Front did not extend across the Mississippi Valley graben. Detrital zircon analysis by Thomas et al. (2021) suggests that by the late Mississippian the Illinois basin of the inner craton was being routed to the west, making it an unlikely source for the Hartselle Sandstone. Combined U-Pb detrital zircon geochronology and zircon (U-Th)/He thermochronology suggest local uplift of the southern Appalachians and sediments supplied by southern terranes could supply the necessary detrital zircon ages and does not require input from northeastern Appalachian or interior cratonic sources (McKay et al., 2021). Mixing models using U-Pb ages from detrital zircon samples collected from the Hartselle Sandstone and sandstone members of the Pride Mountains Formation suggest source terranes from the southern Appalachian source and contribution from similar source that supplied to the Arkoma basin and accretionary wedge of Ouachita orogenic belt (Konopinksi et al., 2022).

CRETACEOUS

The Cretaceous sediments that unconformably overlie the strata of Black Warrior basin are the conglomeratic Gordo Formation of the Tuscaloosa Group, which is the basal unit of the Gulf Coastal Plain (Drennen, 1953a). The Gordo Formation present on the Bishop quadrangle is primarily composed of interbedded clay, sand, and subangular chert gravel what varies between clasts and matrix supported (Fig. 8). Hilgard (1860) was first to remark on the Cretaceous sediments of the Coastal Plain. Initially the lowest Cretaceous unit was known as the Eutaw, however a substantial portion of this group were found to be covered by the Quaternary Orange Sand, that are often difficult to separate from the sandy upper portions of Eutaw. Its name originated from exposures near Tuscaloosa, Alabama and includes clays, sands, and gravels between the rocks of Paleozoic age and Upper Cretaceous Eutaw formation (Smith and Johnson, 1887). Smith et al. (1894b) included the Tuscaloosa Group in the Cretaceous stratigraphy of Alabama. Smith et al. (1894a) commented on the extensive gravels of the Tuscaloosa and the variation between the height of the Tuscaloosa near the waterways and the capping of hills of Colbert County, Alabama. Smith et al. (1894b) considered all Cretaceous strata about the Paleozoic and below Eutaw to be mapped as the Tuscaloosa Group. Stephenson (1914) correlated Tuscaloosa Group in Mississippi and Alabama and distinguished it from the Eutaw by the characterization of interlayered sands, clays, and gravels that indicated a probable estuarine or shallow marine origin. Stephenson (1926) initially provided identification of fossil plant fragments and noticed the variation in the basal gravels of the Tuscaloosa Group. The composition of gravel shifted the ratio of quartz to chert decreases westward across Alabama and Tennessee. Subdivisions of the Tuscaloosa Group were assigned in ascending order: Cottondale, Eoline, Coker, and Gordo Formations. This was based on geologic mapping of the Upper Cretaceous in Tuscaloosa and Cottondale quadrangles, Alabama (Conant et al., 1945) and expanded to other western Alabama counties (Monroe et al., 1946).

The subdivisions devised by Conant et al. (1945) were altered based on study of the Tuscaloosa age sediments in other regions, the Cottondale and Eoline as previously defined could not be differentiated (Drennen, 1953a). The name Cottondale was abandoned, and the Eoline formation was considered the lower member of the Coker formation. An upper member was recognized within the Tuscaloosa group, dividing the unit into two members, the lower Coker Formation, and the upper Gordo Formation (Drennen, 1953a).

The Gordo Formation is composed of heterogeneous bedded clay, sand, and gravel forming extensive lenses with internal lenses of subangular chert gravel. The base of the formation is marked by an irregular layer of limonite cemented chert-pebble conglomerate. The coloration ranges from brown, maroon, red, orange, yellow, and white with medium to coarse grain friable sand, including an upper bed of primarily thinly bedded sand and massive clay (Drennen, 1953b). The Gordo Formation is the unit only observed member of the Tuscaloosa Group within the mapping area, and thus will be referenced as the Tuscaloosa Group.

The division between the eastern and western facies was further delineated in Tennessee due to the inclusion of quartz pebbles and chert that must have originated in the east (Marcher and Stearns, 1962). The eastern facies of the Tuscaloosa is found in a small belt near the eastern margin of the Western Highland Rim in Tennessee which are distinguishable from the western facies in containing quartz pebbles, with sizable portions of sand and the only fossils found within the Tuscaloosa. The difference in the Tuscaloosa lithologies of the western and eastern facies was concluded to be a result of source area and transport modes (Marcher and Stearns, 1962). The western facies of predominantly chert composition are present on the Bishop quadrangle. Russell (1987) recognized two major late Cretaceous stream systems that transported the gravels that would become the Gordo Formation of the Tuscaloosa, a southeastern flowing stream along the Pascola arch (Stearns and Marcher, 1962) into northeastern Mississippi, and a southwest flowing stream into northern Alabama to contribute metamorphic quartzite grains.

All Cretaceous strata above the Paleozoic rocks, not including Quaternary deposits, are of the Tuscaloosa Group. The Tuscaloosa is of variable thickness and ranges from 50 feet to over 200 feet (Merrill, 1998), though the elevation of unconformity changes over the course of the Bishop quadrangle from 800 feet to 400 feet. The overall trend is less than 2 degrees to the west. Fossils are generally absent but may contain plant fragments or lignified logs (Stephenson, 1926). Drennen (1953b) found borings that may be linked to *Halymenites* and other sources of bioturbation.

Exposure of this unit displaying western chert dominated facies of the Gordo Formation is located S½ sec. 6, T. 4 S., R. 14 W. (Fig. 8). The Tuscaloosa Group is a thick conglomerate unit that represents the formation of the Mississippi embayment (Cox and Van Arsdale, 1997; 2002). Detrital zircon U-Pb ages of samples taken from Mississippi and Alabama suggests the Tuscaloosa fluvial system did not connect to the western U.S. or the continental interior and its drainage was sourced from the Appalachian-Ouachita cordillera (Blum and Pecha, 2014; Blum et al., 2017).

QUATERNARY

All Quaternary age sediments are represented by alluvium and low terrace deposits. The thickness of quaternary deposits varies over the quadrangle, with thin deposits in the valleys and creeks. It is primarily composed of unconsolidated sediments with clay, sand, and gravel lenses. (Harris et al., 1963). Chert pebbles and fragments are common, likely to have originated from the breakdown of the Tuscaloosa and Tuscumbia, though recent quarrying and use in infrastructure often obscures natural deposits.

STRUCTURAL GEOLOGY

Structural features mapped within the Bishop Quadrangle are displayed on the geologic map (Plate 1). The accompanying cross section is a generalized schematic of the underlying stratigraphy and structures. Only a singular well has been drilled on the quadrangle and was referenced for thickness of the Tuscumbia and Fort Payne. For a more detailed subsurface stratigraphy (Welch, 1958; Welch, 1959; Merrill et al., 1988). The Paleozoic rocks dip primarily to the southeast and are nearly horizontal on the quadrangle; however, variation in the local dip has been attributed to a variety of factors, such as folding or slumping. It is difficult to distinguish structures due to variability in the lithology and thickness of the units and the erosional unconformities between Mississippian age strata. The Tuscaloosa group also blankets a substantial portion of the quadrangle and unconformably overlies the Paleozoic strata, further obscuring and complicating determining the extent and location of structures.

McCalley (1896) noticed changes in the Tuscumbia Limestone and clastic units of the Pride Mountain formation, describing waves within the Tuscumbia Limestone creating crests and the clastic units of Pride Mountain formation filling the troughs. These were recorded to be approximately 500 to 600 feet long and 20 to 25 feet deep. Harris et al. (1963) noted in their report on the water resources of Colbert County an anticline in the Tuscumbia Limestone on the Bishop quadrangle, from sec. 11, T. 4 S. R. 15 W. that continues west to sec. 9, T. 4 S., R. 15 W. The axis of the fold is primarily oriented east to west. The strata on either side dip between 4-14°. This deformation is consistent in both Tuscumbia Limestone and the Pride Mountain formation.

Other folds occur on the western portion of the quadrangle with the axis crossing through the centers of sec. 21, 28, 33, T 4 S., R 14 W. A series of small anticline-syncline pairs are arranged northeast to southwest and can be traced onto the adjoining Barton quadrangle. They are gently folded with hinges less than 20 degrees and mapped only within the Pride Mountain formation as due to the lack of outcrops it was not possible to confirm if these structures extend into the underlying Tuscumbia limestone. A larger anticline occurred to the south of these folds that begins on sec. 36., T 4 S., R. 14 W. and extends to sec. 9, T. 5 S., R. 14 W. The southern portion of the fold parallels the plane view hinges of the folds to the north. There is a slight change in the axis of the fold as it continues north. It displaces the Pride Mountain and Hartselle units.

The direction of these eastern and western fold groups parallels the orientation of the Tennessee River further to the north. Mapping on the Cherokee quadrangle to northwest has revealed folding that is parallel to the primary orientation of the Tennessee River Valley (McKay, personal communication). It is possible that the structures were created during syndepositional deformation within the Paleozoic and were later exploited in ancestral drainage systems during the Cretaceous and continue to exert control onto the modern river systems.

CRETACEOUS PALEOTOPOGRAPHY

The basal Cretaceous Unconformity is a defining feature of the Bishop quadrangle but also much of the Gulf and Atlantic Coastal Plain as it represents not only a significant hiatus and passage of time between the two units, but also this contact between Cretaceous sediments and Paleozoic strata preserves the paleo-relief of the Cretaceous (Barineau and Ortega-Ariza, 2021). This contact also demonstrates a tectonic uplift, subsidence, and changes in sea level, but also differing mechanisms of erosion and deposition closer resembling those of today. Mellen (1937) linked the “red prairie soils” as a product of the erosion and weathering of the Chesterian series. Modern soils laterally ramp into the residuum that is in direct contact with Tuscumbia Limestone and Fort Payne Chert in Tishomingo County, Mississippi, west of the Bishop quadrangle. It is possible the landscape was formed by fluvial systems removing clastic material, carving into carbonate bedrock to create a karst topography that was filled by the fluvial deltaic deposition of the Tuscaloosa Group. Recent landscape development was influenced by the preferential erosion of Coastal Plain sediments, the basal Cretaceous unconformity, which preserved within it the regional highs of the Chesterian series, functioned as a topographic intaglio to re-entrain modern river systems to prior drainages etched into the Paleozoic bedrock. Exposures of mottled red and white clays with a basal gravel containing large wood fragments and quartz pebbles (Fig. 9) to the north on the Margerum quadrangle suggests the exposure of Little Bear Residuum was present on the surface at the Time of deposition of the Tuscaloosa. Gravel lenses and plant fragments both overlying and underlying the clay suggest reworking of this layer into the Tuscaloosa (Fig. 10). These clays are largely absent from the Bishop quadrangle, suggesting this may have been a regional high that did not preserve the clays. Alternatively, there is a small exposure of quartz gravel lens underlying the chert conglomerate typical of the Tuscaloosa found on the Bishop Quadrangle. The inclusion of quartz clasts has been associated with the eastern facies of the Tuscaloosa Formation (Stearns and Marcher, 1962). Other areas further west in Tishomingo County, Mississippi have found the eastern facies Tuscaloosa underlying the western chert dominated facies. This is due to the structural changes that occurred throughout the Cretaceous, when the presence of the Pascola arch, a regional extension of the Ozark dome, continued further to the southeast (Marcher and Stearns, 1962). This unconformity is regional dipping to the west. Marcher and Stearns (1962) reported that it increased in slope of the unconformity in Tennessee as it neared the Tennessee River, which may suggest this was a regional low prior to the Cretaceous, acting as a drainage during this time. The orientation of the modern Tennessee River aligns with the structural orientation of deformation in Paleozoic strata, suggesting the current and ancestral drainage pattern may be influenced by structures that formed due to the deformation during the Ouachita Orogeny. Subsurface paleo-valleys within the basal sequence of the Lower Tuscaloosa of Mississippi indicate there may have been a major component coming from the north (Woolf, 2012). This suggests a proto-Mississippi River may have already begun to form, or a now defunct ancestral drainage system was present.

The basal Cretaceous unconformity was mapped previously in Mississippi in the adjacent Tishomingo County (Merrill et al., 1988) and for western Colbert County (Szabo, 1992). Szabo (1992) produced a contour map of the base Cretaceous sediments in western Colbert County, Alabama and portions of Tishomingo County, Mississippi. A change in the elevation of this contact was inferred to be caused by the displacement of Pride Mountain Formation due to dissolution of underlying limestone within both the Tuscumbia and Pride Mountain Formation. This would have produced collapsed basin-like subsidence structures containing disturbed shales and sandstones of Mississippian strata, as well as displaced Cretaceous sediments with undeformed deposits during the Quaternary. No evidence for this style of chemical dissolution and subsidence was found within the Bishop quadrangle. There may have been minor dissolution of carbonate rocks, but unlikely the hundred or more feet of removal required to cause overlying strata to slump and increase slope of this contact near the areas of subsidence. Merrill et al. (1988) interpreted a portion of the structure hypothesized on the Margerum quadrangle, near the northwestern edge of the Bishop quadrangle, to contain a paleovalley to account for the thickness of Tuscaloosa Group in this location.

Reconstructing the paleorelief represented by the erosional surface required delineating the unconformity for the Bishop quadrangle and the surrounding region. The elevation of outcrop data was added for Bishop Quadrangle, as well as from the Margerum, Cherokee, and Barton Quadrangles and approximated beyond based on available maps and data. These outcrops were plotted in ArcMap 10.8.2 and elevation data was extracted from a composite regional digital elevation model for points along the contact. To create a three-dimensional surface of the unconformity the points with extracted elevation data were used in an ordinary Kriging algorithm and inverse weighted distance technique from ERSI ArcMap and found delivered comparable results. The first order regional trend was removed creating a raster by global polynomial interpolation based on the regional contacts, which was then subtracted from the Kriging model. This removed any regional tilting and approximated what the surface would have been prior to the deposition of the sediments of the Tuscaloosa Formation. This reconstruction of the basal

Cretaceous unconformity serves as an approximation of the paleotopographical surface that may have existed during the deposition of the Tuscaloosa Group (Fig. 11). The main channel of the Tennessee River remains coming from the west, but the drainage to the north is now higher in elevation and would have restricted deposition. This may indicate the difference in distribution of the eastern and western facies were drainage systems that were active at different times as the area experienced subsidence or regional deformation to separate or redirect the drainages in reaction to the formation of the Mississippi embayment. Other areas of low relief are similar to the current trend of the Tennessee River, which further suggests that the bedrock structures and orientation may exert a control on both past and present river systems.

SUMMARY

Recent efforts to map the geology of the Bishop 7.5-quadrangle (lats. 88°00’ 00” and 88°07’30”; longs. 34°37’30”and 34°45’ 00”) in southwestern Colbert County, Alabama and eastern Tishomingo County, Mississippi at the 1:24,000 scale has provided a higher resolution for the locations of stratigraphic contacts, and thus a more complete understanding of the stratigraphic relationships and geologic structures. Paleozoic strata located within the Bishop quadrangle is part of the Black Warrior basin, a Late Mississippian to Pennsylvanian foreland basin. Formal stratigraphic units of the Bishop quadrangle include the Tuscumbia Limestone, the Pride Mountain Formation, the Hartselle Sandstone, the Tuscaloosa formation, and undifferentiated Quaternary aged alluvium and low terrace deposits. The stratigraphy of the Bishop quadrangle is predominantly defined by the clastic deposition into a basin experiencing synorogenic differential deformation and eustatic adjustments. This is expressed in shifts from the carbonate ramp of the Tuscumbia Limestone to the lithologic and lateral variability of the Pride Mountain Formation and

Hartselle Sandstone. Folding within these units is hypothesized to be defined by basement faults reactivated during the flexural deformation induced by the Ouachita deformation. The unconformity between Mississippian and Cretaceous rocks varies in elevation which preserves valleys and ridges formed at the time of deposition. This erosional surface was approximated using outcrop data and was used to map the elevation of the basal Cretaceous unconformity. This map serves as a paleogeographic reconstruction that demonstrates the similarity to the modern landscape, with lower elevations aligned with the course of the Tennessee River and its drainages. Folds within the Paleozoic strata on the Bishop quadrangle and adjacent quadrangles are oriented with the river, suggesting the modern Tennessee River system may have been defined by the orientation of bedrock deformation and exploiting ancestral drainage patterns.

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PART 2: GEOPHYSICAL MODELING TO CHARACTERIZE CRUSTAL STRUCTURES AND THEIR RELATION TO DEFORMATION AND CLASTIC SEDIMENTATION OF THE MISSISSIPPIAN BLACK WARRIOR BASIN

ABSTRACT

Middle to Late Mississippian strata of the Black Warrior basin records the initial deposition of siliciclastic sediments as the foreland basin experienced subsidence. Changes in sea level, paleoclimate, and tectonic factors combined with an evolving topographical surface produced a complex depositional history. Coeval influx of detrital sediments and differential subsidence occurred as the basin underwent deformation due to the obduction of an accretionary wedge as the Sabine terrane converged to the southwest of the Laurentian margin. As the flexural bulge migrated north it induced variable rates of subsidence and uplift throughout the basin. Deformation may have exploited pre-existing faults to accommodate the evolving tectonic load induced by the evolving Ouachita orogenic front. Gravity and magnetic data were analyzed using maps and 2D forward models of upper crustal structures. The geophysical analysis suggests the delineation of the depth and offset of Precambrian basement and denser bodies within the subsurface that were instrumental in accommodating stress from the tectonic forces. These models helped to further resolve a connection between basement faults and crustal structures, which may have impacted sedimentary accumulation within the region in response tectonic loading by converging of the Laurentian margin with outboard terranes during the Ouachita orogeny. Sedimentary thickness only records only a portion of this complex past, however the boundaries of sedimentary isopach maps demonstrate preservation of sediments along faults as they were reactivated, the positions of which may have been defined by anomalous density highs within the crust that may have been emplaced by intrusive suites or inversely may have followed structural weaknesses from prior tectonic events.

INTRODUCTION

The Alabama promontory is considered a stable passive margin, which was the site of a carbonate ramp from the Cambrian to the Late Mississippian. The region lies north of the Ouachita-Appalachian juncture, the syntaxial bend where the Ouachita and Appalachia orogenic fronts converged (Fig. 12). The Black Warrior basin is a foreland basin that experienced differential subsidence with the initial deposition of a clastic wedge during the Late Mississippian (Higginbotham, 1986). The discontinuous synorogenic clastic wedge of the Late Mississippian Pride Mountain Formation and Hartselle sandstone has been extensively studied through both outcrop and subcrop analyses (Thomas, 1972a, 1972b; Welch, 1958; 1959; Pashin, 1993). The sediments that constitute the clastic wedge coalesced within the basin at the onset of the Ouachita orogeny, induced by the flexural subsidence and uplift within the basin. There has been debate about the origin of clastic input into the Black Warrior basin and its depositional history. A northern cratonic source has been advocated based on petrographic analyzes, subsurface stratigraphy, paleotopographic and paleoclimatic evidence (Cleaves and Broussard, 1980; Cleaves, 1983; Cleaves and Bat, 1988; Pashin, 1993). Based on an alternative petrographic analysis, a southern source from the uplifted region produced by the Ouachita orogeny resulted from reworked grains in the subsiding margin that prograded to the northeast (Thomas, 1972a; Mack et al., 1981; Thomas and Mack, 1982). Tectonic activity of the Ouachita orogeny was suspected to play a role as the flexural bulge induced by thrust loading (Fig. 13) (Ettensohn and Pashin, 1993). The initial clastic sedimentation and accommodation induced by the onset of the Ouachita Orogeny represents the base of the Chesterian series in Alabama. Log patterns were interpreted and used to make a structural contour map of the top of the Tuscumbia Limestone (Pashin, 1993). Progressive topographical evolution occurred as a result of a complex erosional and depositional history determined using isopach maps and geophysical well log patterns indicate a possible tectonic control on sedimentation (Pashin and Rindsberg, 1993). The complex history represented by the depositional settings of the Late Mississippi clastic wedge indicate climatic, topographic, and eustatic factors played significant roles in the formation and distribution of clastic units, in addition to a tectonic influence (Pashin, 1993; Thomas, 1988).

The Late Mississippian siliciclastic wedge thins rapidly along two bands, forming an orthogonal trend that intersects in southern Marlow County and northern Lamar County, Alabama (Fig. 14). The northern edge of Lowndes-Pickens block was hypothesized to have acted as a structural hinge, which the vertical displacement allowed for the accommodation of tectonic loading and extensional setting of the foreland basin (Higginbotham, 1986; Kidd, 2008). The Lowndes Pickens block created a structurally high area north of block. Synsedimentary differential uplift was thought to have been primarily controlled along this trend with the thickest areas aligning with the region of rapid subsidence. Shallow water sandstones require that the sedimentation rate to be equal to the rate of subsidence to form and accumulate. Net sedimentation on the uplifted Lowndes Pickens block was negligible as sediments prograded northeast, following the deposition of the initially clastic deposition within the basin (Higginbotham, 1986; Kidd, 2008).

In the southern and southwestern portion of the basin, there are numerous northwest-southeast trending normal faults in the Late Paleozoic strata that were products of extension (Figs. 15 and 16) (Groshong et al., 2010; Thomas, 2010). They range from thin-skinned detachments that penetrated the basement as shown in seismic reflection that profiles demonstrate down-to-the-southwest normal faults that penetrate the basement (Thomas, 1988; Groshong et al., 2009; Groshong et al. 2010; Thomas, 2010). Farther to the south, on the western portion of the syntaxial bend of the junction between the Ouachita and Appalachian orogenic fronts a transpressional zone has been described from seismic reflection data (Hale-Erlich and Coleman, 1993). There is evidence of faulting within the Mississippian Black Warrior basin to the south based on the Tuscumbia Limestone (Stapor and Cleaves, 1992).

Extensive work has been done in the study area to record the surficial and subsurface stratigraphy (Welch, 1959; Thomas, 1972a, 1972b; Pashin, 1993; Pashin and Rindsberg, 1993; Kidd, 2008; Canton, 2011). However, this work has not been able to resolve the input source or how tectonism affected the sedimentation due to the complex history of the region. Graben systems formed during the Precambrian that paralleled rifted margins were reactivated in the late Paleozoic (Thomas 1991; 2010). The possibility of internal subsidiary grabens within the larger context of continental basins was introduced based on a geophysical data analysis within the northern Mississippi Embayment (Johnson et al., 1994). The underlying Precambrian basement composition is unknown (Bickford et al., 2015; Chen et al., 2018). Large amplitude gravity and magnetic maxima in the region have been interpreted to be buried mafic bodies (Steltenpohl et al., 2013; Johnson et al., 1994). They may fall along crustal boundaries from pre-existing crustal structures that formed during Proterozoic accretion or Precambrian rifting.

Gravity and magnetic anomaly maps produced by this study show improved resolution over previous maps and were used for resolving the depth to magnetic basement, the location of igneous intrusions, and to better resolve interpreted fault and grabens formed by Precambrian rifting. 2-D forward gravity and magnetic models were created to represent the possible lithological compositions and geometries within the research area. These models will serve to assist in defining the subsurface geology of the Black Warrior basin by creating working models of the crustal structures based on geophysical data in the absence of physical data. Maps of the complete Bouguer gravity anomalies, isostatic residual anomalies, reduced-to-poles magnetic anomalies, and vertical derivatives of the magnetic anomalies were produced and used to compare the sedimentation isopach maps with geophysical trends to infer the relationship of subsurface structures to Mississippian clastic deposition.

GEOLOGIC SETTING

The Middle to Late Mississippian System of the Black Warrior foreland basin represents the transition from a carbonate dominated ramp to a siliciclastic wedge that occurred during the initial stages of Ouachita orogeny (Thomas, 1974; 1985). The basin is a homocline of Paleozoic strata that dips to the southwest. There are numerous northwest-striking normal faults along this homocline, increasing in frequency and displacement toward the Ouachita orogenic belt (Thomas, 1988). This region lies between the seismically active zones of the Mississippi Valley graben to the west, the Rough Creek graben to the north, and the Eastern Tennessee seismic zone to the east (Thomas and Powell, 2017). Wells in the region have not penetrated the Precambrian crystalline bedrock beneath the Phanerozoic sedimentary sequences.

Exposed Paleozoic rocks in northwestern Alabama are predominantly Carboniferous in age, with exposures of Devonian, Silurian, and Ordovician outcrops near the Alabama and Tennessee border. The Alabama promontory was a stable carbonate dominated platform from the Cambrian to the Late Mississippian with the formation of the Black Warrior basin (Thomas 1972b). The Black Warrior basin is a foreland basin formed by the flexural subsidence and by the viscoelastic deformation of the margin (Quinland and Beaumont, 1984; Hines, 1988). Down warping produced a flexural basin adjacent to the orogenic front and a flexural bulge along the craton ward side that migrated inland with the advancing thrust load (Quinlan and Beaumont, 1984; Beaumont et al. 1987; Hines, 1988). The thrust load was induced by the obduction of the accretionary prism from the collision of the Sabine terrane with the southern margin of Laurentia (Mickus and Keller, 1992).

The Grenville orogeny occurred from 1.3-1.0 Ga (Bartholomew and Hatcher, 2010). Following the formation of Rodinia during the Grenville orogeny, rifting occurred. This late Precambrian to early Cambrian rifting shaped the continental margin of southern Laurentia forming the Alabama promontory and Ouachita embayment (Thomas and Astini, 1999). The margin continued to impact the region as tectonic inheritance and rifting from the breakup of Rodinia produced both synrift sedimentary and igneous rocks, in addition to the embayments and recesses the formed the Laurentian margin which defined much of the assemble of Pangea in the late Paleozoic (Fig. 12) (Thomas, 2006; 2010). Internal cratonic graben systems also were also activated during both the assemble of both Rodinia and Pangea (Thomas, 1991).

The resurgent Nashville dome was activated during the Mississippian, though it remained a topographic low during the Upper Mississippian (Stearns and Reesman, 1986; Quinlan et al., 1987). The region of focus for this study is to the south of the Nd-Line, an isotopic boundary that separates older granites and rhyolites to the northwest and more juvenile intrusions to the southeast (Bickford et al., 2015). The rocks that comprise much of the midcontinent south of this isotopic boundary were formed during a Proterozoic accretionary event at approximately 1.50–1.44 Ga and formed the Eastern Granite Rhyolite province and subsequent magmatism during 1.40–1.34 Ga formed the southern Granite Rhyolite province (Van Schmus et al., 1996; Bickford et al., 2015). It is unknown if the terrane on the eastern side of the Mississippi Valley graben is compositionally similar to the study region and continues to the southwest. If the Eastern Granite Rhyolite province is not a continuous then it may represent a more localized feature, with Precambrian basins not underlain by an extensive Eastern Granite Rhyolite province, but rather discrete volcanic regions, such as those of the St. Francois Mountains, with the surrounding region composed of predominantly sedimentary or metamorphic terranes with a thin layer of overlying volcanoclastic material (Pratt et al., 1992). The nearest borehole to penetrate igneous bedrock is located along the southern portion of the Nashville Dome, in Giles County, Tennessee. Granite was found at the base of the borehole at a depth of 5620-5750 ft and was dated to 1120 Ma. It was described as a massive granophyric porphyry containing microcline, glaucophane, plagioclase, and quartz (Wasserberg et al., 1962). To the northwest along the intracontinental grabens igneous intrusions have occurred at different times, accompanied by cycles of uplift, subsidence, and erosion (Hildenbrand, 1985; Thomas et al., 2012).

PREVIOUS INVESTIGATIONS

Previous studies on the sedimentary history of the region can be found in chapter 1 of this report. Gravity and magnetic surveys that directly studied northwestern Alabama and northeastern Mississippi are scarce. The study area lies at the periphery of two actively studied regions, the New Madrid seismic zone and the East Tennessee seismic zone. Gravity and magnetic data were used to model the northern Mississippi embayment and found a denser region which may be possibly igneous or volcanic underlying the region (Hildenbrand, 1985). Seismic reflection surveys of the Consortium for Continental Reflectional Profiling (COCORP) found a reflective zone that dips west that was suggested to be due to Grenville deformation penetrating further west in Alabama and southern Tennessee (Culotta et al., 1990). Gravity and magnetic anomalies were analyzed to interpret that these anomalies coincide with grabens in the magnetic basement consistent with depths to the Precambrian crystalline rocks. However, this assumption was complicated with compositional variability and regional changes in the depth to Precambrian crystalline rocks (Johnson et al., 1994).

The New York-Alabama (NY-AL) lineament delineates a crustal boundary beneath the Appalachian basin from Vermont to Alabama and Mississippi. It separates the magnetic anomalies of the eastern Granitic Rhyolite province and the Proterozoic bodies to the northwest from the Appalachian orogenic front to the southwest (Zietz and King, 1978; Steltenpohl et al., 2010). Extensional faulting mapped to the southwest of the study area was found to be both thin-skin deformation that penetrated only into Paleozoic rocks and thick-skinned where the faults connected to crystalline basement (Groshong et al., 2010). The gravity and magnetic anomalies in northwestern Alabama were correlated and interpreted to be buried basement and above basement mafic bodies, which agree with the interpretation of previous geophysical surveys (Steltenpohl et al., 2013) and highlighting the eastern Continental Gravity High. There have been multiple interpretations of the gravity anomaly associated with the eastern Continental Gravity High, including a possible be a southern extension of the eastern Mid-Continental rift system (Stein et al., 2017; Stein et al., 2018). 3D P-wave velocity models of the region shift from a high velocity zone to the lower velocity zone in the upper mantle based on P-wave tomography, which may be remnants from the passage over the Bermuda hotspot or a continental plate beneath the New Madrid seismic zone (Chen et al., 2014).

Isopach maps of crustal thickness from crystalline bedrock to the Moho boundary reveal a thickening of crust to the northeast beneath the Nashville Dome and thinning of crust to the southwest, with an isolated region that may be representative of shift in the intracratonic basin setting between the Reelfoot rift and the Ouachita-Appalachian orogenic front (Yang et al., 2017). Northwestern Alabama, northeastern Mississippi, and south-central Tennessee resides at the nexus of these various crustal thicknesses. Thicker crust is present to the northwest, in the region between the field area and the Nashville Dome. Thinner crust is present in the southwest and southeast (Chen et al., 2014). A high velocity zone beneath the Reelfoot rift may be a rift pillow, which may have formed during the emplacement of Eastern Granite Rhyolite province or Proterozoic rifting, could have influenced crustal strength and controlled the formation of the Mississippi Valley graben during the breakup of Rodinia (Yang et al., 2022).

The Eastern Embayment graben is a structure that lies to the northeast of the region (Marlow, 2021). Due to a lack of data to determine the depth of the basement limits the ability to delineate this structure, though it is assumed to follow the general northeast trend of structures determined from earlier studies (Johnson et al., 1994), or it may be composed of smaller basins lying along regional deformational trends. There is a possibility that this region contains intrusive suites that formed during the extension that produced the Reelfoot rift during Neoproterozoic or early Paleozoic rifting (Johnson et al., 1994). The depth to the magnetic basement may change by as much as 6 km of offset if it is approximated to coincide with the depth to the Great Unconformity (Marshak et al. 2017; Marlow, 2021).

Seismic ambient noise tomography of the northern Mississippian embayment revealed high velocity zones, which may indicate intrusive bodies that coincides with the eastern Continental Gravity High. The velocities suggest the anomalies are more related to Eastern Granite Rhyolite province than the Appalachian front. Localized high velocity areas form a distinct pattern at 5 km and again at 15-20 km, which overlap with gravity anomalies and suggest the presence of anomalous dense bodies within the subsurface (Yang et al., 2022).

The thickness of crust, based on a broadband seismic analysis, suggests an intracratonic basin formed in the early Cambrian, during Iapetan rifting or an early tectonic event. Mafic intrusions may exist in the lower crust beneath the Illinois basin, producing a denser lower crust and an overall thickening the crust, which may induce basin subsidence. Regions with a thinner crust may have been produced via delamination during the formation of the East Granitic Rhyolite Province, Grenville Orogeny, or following the failure of the eastern branch of the Mid-Continental rift system leading to underplating delamination, which induced isostatic adjustments (Yang et al. 2017) or re-lamination of lower crust (Chen et al. 2016). However, these structures are not consistent with patterns in crustal thickness, suggesting the dominant mechanism of basin and arch formation of the continent is not related to crustal thickness (Xiao et al., 2022).

METHODS

GRAVITY ANOMALY MAPS

Gravity data used in this study were obtained from the National Geospatial Imaging Agency. Free-air, Bouguer gravity and terrain corrections were applied using sea level as a datum, a density of 2.67 g/cm ^{3}, and 1 km DEM to produce a complete Bouguer anomaly map (Fig. 17). Then additional gravity and elevation data in a 3° rind surrounding the study area were used to create an isostatic residual gravity anomaly map. A crustal thickness of 40 km, crustal density of 2.72 gm/cc and mantle density of 3.2 gm/cc were used in creating the isostatic residual gravity anomaly map (Fig. 18) (Simpson et al.van , 1986). Using these maps with a reduced-to-pole magnetic anomaly map has previously been used in delineating orogenic boundaries and blocks that are not exposed at the surface (Mickus, 2007). The isostatic residual gravity anomaly map has the same gravity anomaly patterns seen in complete Bouguer gravity anomaly map. The isostatic residual gravity anomaly map has individual anomalies that have been slightly modified to reflect crustal upper crustal sources, removing the regional gravity effect present in the complete Bouguer gravity anomaly map.

MAGNETIC ANOMALY MAPS

Crustal magnetic anomalies are produced by varying magnetic susceptibility of lithologic materials. Large scale magnetic anomalies may be due to changes in the depth of magnetic basement or due to crustal heterogeneities, as sedimentary lithologies usually are not magnetic. The data for magnetic maps was obtained from the U.S. Geological Survey from a compilation of over 1000 magnetic surveys over 50 years to create a cohesive map of North America (Bankey et al., 2002). The total field magnetic data were gridded at a spacing of approximately 1km. The Earth's magnetic field dipolar effect was removed and reduced to the north magnetic pole. The grid was contoured at 200 gamma intervals producing a map of reduced-to-poles magnetic anomaly map (Fig. 19). To better delineate the edges of crustal structures, a vertical derivative filter was applied to the reduced-to-pole magnetic anomaly map to create a vertical derivative magnetic anomaly map, which has been used to identify the boundaries of sources of magnetic susceptibility (Fig. 20) (Verduzco et al., 2004; Matende et al., 2023).

GEOPHYSICAL MODELS

Detailed knowledge of subsurface lithologies and structural makeup is essential for the construction and interpretation of geophysical models. Constraints such as geological mapping, seismic reflection/refraction models and borehole data are needed to obtain resolved models. Additionally, various lithologies can produce similar magnetic anomalies thus past studies and analogues from exposed and better studied regions can help to resolve some of these uncertainties. Due to the unknown nature of the Precambrian basement composition in this region analogues were chosen from other geophysical studies for the possible geometries and lithologies. The St.

Francois Mountains are some of the few outcrops of the Eastern Granite Rhyolite province that were used to provide representative values of subsurface geophysical properties (Ives and Mickus, 2019; McCafferty et al., 2019). Other intracontinental basins such as the Mississippi Valley graben and the Rough Creek Graben were also considered in the relationship to this region and taken into consideration for representing conditions that would produce feasible working models (Hildenbrand, 1985; Johnson et al., 1994; Hildenbrand et al., 1996; McBride et al., 2016, McBride et al. 2018; Marlow, 2021).

Magnetic anomaly maps reflect the changes in the magnetic susceptibilities in the underlying magnetic basement. Since the Paleozoic sandstones and carbonates overlying the Alabama promontory usually have low magnetically susceptibilities, the values for sedimentary layers that ranged in density from 2.6-265 g/cm^3 and 0.00 emu for the magnetic susceptibility. The densities and magnetic susceptibilities for the basement lithologies, assumed to be the Precambrian crystalline basement with values between 2.7-2.75 g/cm^3 and magnetic susceptibilities between 0.003-0.01 emu (Ives and Mickus, 2019; McCafferty et al., 2019). The depths to crystalline basement were approximated based on gravity and magnetic models from beneath the Mississippi embayment that extended to the southeast (Hildenbrand, 1985; Johnson et al., 1994; Yang et al., 2022). The densities for the crustal bodies were set to 2.7-2.85 g/cm^3, with a magnetic susceptibility between 0.01-0.015 emu.

The depth and geometries of subsurface structures and bodies were based on seismic broadband analyzes surveys within the New Madrid seismic zone region (Marlow, 2019; Yang et al., 2022), Rough Creek graben (Hildenbrand et al., 1996), and Illinois basin (McBride et al., 2016; McBride et al., 2018). The subsurface density and magnetic susceptibility of bodies and their geometries were modeled in the Oasis Montaj GM-SYS program by creating discrete shapes with density related to the composition most likely to exist based on values from other studies (Telford et al., 1990; Hildenbrand, 1985; Johnson et al., 1994; Ives and Mickus, 2019; McCafferty et al., 2019). An iterative process was used to change the physical property values, depth, and geometries of the subsurface bodies' parameters until the error between the observed and calculated anomalies were below an acceptable limit. These results were then combined to create a cohesive model to provide an integrated interpretation of crustal structures.

The profiles were chosen based on the proximity of region to existing structures, availability of sedimentary isopach data, and proximity to geologic mapping (Chapter 1 of this report). Profile A-A' was selected based on the large magnetic and gravity anomaly that coincide at this location. It also is the furthest extend of the Late Mississippian clastic wedge sediments and there are Devonian sediments near the Tennessee River. Profile B-B' was chosen due to the circular structure and the proximity to the mapping region (Chapter 1 of this report), which revealed folds and soft sediment deformation that suggests syndepositional deformation of the region during the Late Mississippian.

An upper sedimentary layer was added to each model with a density of 2.6g/cm^3, in addition to a lower sedimentary representing Cambrian and older sedimentary layers with density of 2.65 g/cm^3. The upper and lower crust used depths from Chen et al. (2016), and densities of 2.7-2.75 g/cm^3, lower 2.8-2.85g/cm^3, respectively (Hildenbrand, 1985; Johnson et al., 1994). These depths and physical property values were used as initial conditions, for the depth to the magnetic basement was assumed to be the same as the depth to the Precambrian basement. This was approximated to be 2 km in the region (Marshak et al., 2017). The depth may be offset by up to 6 km in areas based on a more recent seismic survey of the Mississippi embayment (Marlow, 2021).

2D forward models were created to provide a better representative image of the subsurface. One was placed to cross a large gravity and magnetic anomaly directly to the northwest of the eastern Continental Gravity High, and one centered over the Bishop quadrangle and a circular anomaly that is visible on the magnetic anomaly maps. The models were derived using a 2D forward modeling algorithm (GM-SYS) in Geosoft Oasis Montaj. The Moho depth was approximated using models from previous seismic surveys (Chen et al., 2014; Yang et al., 2017). The interpretations are theoretical and non-unique as multiple density and structural geometries can produce the same geophysical values at the surface (Lowe, 1999).

REGIONAL ISOPACH MAPS

Sedimentary depocenters were assumed to align with isopach maps of clastic members of the late Mississippian units. The timing and the thickness of the sedimentary units represent the sequential deposition of clastic units within the basin. The gravity and magnetic anomaly maps were compared to these isopach maps correlate the possible relationships between increased sedimentary deposition to geophysical and structural trends. Isopach maps of the Late Mississippian sedimentary thickness for clastic members of the Pride Mountain Formation and Hartselle Sandstone were acquired from Kidd (2008) and Canton (2011) and georeferenced in ArcMap 10.8.2. The georeferenced isopach maps were overlain on the vertical derivate magnetic anomaly map.

RESULTS

GRAVITY ANOMALY MAPS

The isostatic and residual gravity anomalies are similar in amplitude, size, and geometry. The anomaly values range from ~10 to ~-90mGal on the complete Bouguer anomaly map (Fig. 1

) and 30 to -40 mGal for the isostatic residual anomaly map (Fig. 18). Gravity maxima are found in northwestern Alabama that continue to the northeast are related to the eastern Continental Gravity High, what has been interpreted as southeastern extension of the Mid-Continental rift or possible Grenville front. Existing highs occur along the Mississippi, Tennessee, and Alabama border are oriented in a northwest-southeast direction. The examination of this gravity anomaly that is arranged northwest, approximately in the same direction as the Ouachita front to the southwest, with orthogonal high and low trends to the southwest in both the complete Bouguer and isostatic residual anomaly maps (Figs. 17 and 18).

MAGNETIC ANOMALY MAPS

Magnetic anomaly maps usually have smaller in area and more elongated wavelengths anomalies when compared to the complete Bouguer gravity and isostatic residual anomaly maps. The RTP magnetic anomaly map has anomalies that range from -895 to 415 nT (Fig. 19). The NY-AL lineament (King and Zietz, 1978; Steltenpohl et al., 2010) is evident from the large amplitude linear magnetic anomaly that strikes northeast from Alabama to Tennessee. The eastern Continent Gravity High is also visible, though less pronounced than in the gravity anomaly maps. The highs in the southwestern portion near the Mississippi, Tennessee, and Alabama borders are more prominent than in the gravity anomaly maps. The vertical derivative magnetic anomaly map displays smaller wavelengths than the reduced-to-poles magnetic anomaly map (Fig. 20). This expected as the vertical derivate magnetic anomaly maps, is a gradient map based on the rate of change in the vertical component of the magnetic field induced by the source of the magnetic susceptibility.

GEOPHYSICAL MODELS

CROSS SECTION A-A'

Cross section A-A' shows the center of the center of the maximum in both the gravity and magnetic anomalies that requires denser and more magnetically susceptible material in the subsurface. Model 1 requires several dense bodies near the subsurface, but not intruding into Paleozoic sedimentary units. There is a significant offset in the magnetic basement, approximately 2 km, which may signify a fault boundary that contains denser rocks to fill the space that was created by the displaced boundary in the magnetic basement. The change in the density of the values used across the displaced boundary in the magnetic basement may indicate this is a normal fault that was filled with denser sediments grabens formed during the late Precambrian or Early Paleozoic (Fig. 21). Model 2 shows a large anomalously magnetic body approximately 10km beneath the surface that extends down to approximately 40km. If the same can be inferred from model 1 in relation to the offset in the boundary represented by the magnetic basement, then Model 2 would feature a large basin with an offset of 6km that formed in this region over the anomalous density body, before being covered by a thin (<2km) veneer of Paleozoic sediments (Fig. 22).

CROSS SECTION B-B'

The models of cross section B-B' share more consistent geometries than the possibilities represented by those of cross section A-A'. The final models for cross section B-B' indicate the magnetic basement is projected to reach a depth of <2 km below the surface elevation in the locations that coincides with the magnetic and gravity maxima. The offset in the magnetic basement may reach a magnitude of 4-8 km where there is a gravity and magnetic minimum, which is dependent on the physical property values and geometry used to create the model. There is the possibility of a denser region that would occur near the anomaly maxima as shown in model 2 (Fig. 23), which reduces offset in the magnetic basement. The geometry remains similar with either an uplifted portion with denser bodies within an uplifted portion as exhibited in model 1 (Fig. 23), or a dense, more magnetically susceptible core that could represent a structure with a denser body with faulted around it due to similar to volcanic structures of the St. Francois Mountains indicated by model 2 (Fig. 24).

REGIONAL ISOPACH MAPS

A comparison of the regional isopach maps reveals changes in thickness of clastic members of the Pride Mountain Formation and the Hartselle Sandstone follow general trends of the magnetic and gravity anomaly maps. The changes in extent and distribution of the isopach maps can best seen by overlaying the isopach maps with the magnetic tilt derivative map (Fig. 20). The results from modeling of gravity and magnetic data previously discussed indicate anomalous bodies within the near surface, with an offset in the magnetic basement occurs on either side of the vertical components of magnet anomaly maximums. If the same relationship of the offset in the magnetic basement and presence of anomalous bodies occurs in the subsurface extends to the southwest in the Black Warrior basin, then it may be inferred that trends of the maxima and minimum of the gravity and magnetic anomaly maps aligns with structural offset in the magnetic basements and anomalous crustal bodies. If the changes in the isopach of the clastic units are in part a function of increased or decreased accommodation space, this could be caused in part by the movement along structural blocks induced by far-field stress during the formation of the Black Warrior basin (Higginbotham, 1986).

The initial introduction of sediments to the Black Warrior foreland basin is contained are represented by the isopach maps of for the Lewis cycle (Canton, 2011), and the Lewis Sandstone (Kidd, 2008), which are both clastic members of Pride Mountain Formation. The isopach maps follow relatively the same shape and trend. Figure 25 represents progradation from the southwest. Figure 26 presents if the sediment prograded from the northwest. Despite the conflicting interpretation of the origin and orientation of the Lewis Sandstone it follows a magnetic minimum oriented northeast to southwest and another minimum oriented northwest to southeast. These magnetic minima may represent an offset in the magnetic basement preserved that may have served to increase the accommodation of the sandstone (Figs. 25 and 26). The Evans Sandstone is the clastic sequence between the Lewis Sandstone member of the and Hartselle Sandstone. The isopach map of the Evans Sandstone (Fig. 27) resembles a similar pattern to the Lewis Sandstone isopach map (Fig. 25), however region of greatest thickness is further to the east. The isopach map of the Evans cycle (Fig. 28) contains a much thicker region to the west that was interpreted to represent a deeper portion of the basin as the foreland basin deepened to the southwest. The Pearce Siltstone is an often-unrecognized member of the Hartselle Sandstone that formed that is hypothesized to have formed in a restricted basin (Kidd, 2008). The isopach follows a magnetic maximum oriented northwest to southeast adjacent to the edge the Hartselle Sandstone (Fig. 29). The isopach of the Hartselle Sandstone is oriented in a similar fashion to the Pearce Siltstone, predominantly in a northwest to southeast orientation, however it is located further to the east, which may represent a faulted region that is able to accommodate increased sediment (Figs. 29 and 30) (Kidd, 2008).

DISCUSSION

A primary purpose of this study was to use gravity and magnetic data to map anomalies within the northern Black Warrior basin. Maps of gravity and magnetic anomalies were used to model subsurface structures and anomalous crustal bodies. Isopach maps of sedimentary thicknesses of the clastic units of the Late Mississippian were correlated with gravity and magnetic anomaly maps. By integrating the results of geophysical models with maps of geophysical anomalies it may be possible to approximate the distribution of structural elements within the subsurface across a region. By incorporating isopach thicknesses of clastic units the goal of the study was to relate effect of deformation through time along structural trends on sedimentation related to the flexural subsidence or uplift induced by far field stress of thrust loading by the Ouachita orogeny.

Geophysical models suggest positive anomalies may be associated with anomalous crustal bodies or offset in the Precambrian basement that formed grabens from reactivation of fault bound structural blocks within the Precambrian basement, within the larger Black Warrior foreland basin. The transition from magnetic and gravity maxima to minima are assumed to align with the boundaries of graben within the magnetic basement (Johnson et al., 1994), and similar to perpendicular arranged faults in Precambrian terrane found in the St. Francois Mountains (Kisarsanyi and Kisvarsanyi, 1976). Even with the large offsets of Precambrian basement of up to 6km (Marlow, 2021) it was determined this would lack to account for the magnitude of magnetic and gravity anomalies within the subsurface in northern Black Warrior basin. There were few constraints for the depth and physical properties of the subsurface bodies in the study as no boreholes penetrated Precambrian bedrock. Multiple iterations continued to migrate back to similar interpretations, the most notable being the proximity of the magnetic basement, assumed to coincide with the Precambrian basement, to the surface (<2 km), with large offsets that likely represent basement faults.

A denser body, or bodies, may exist within the upper 5km of crust in the northern Black Warrior basin, as well as the existence of a larger dense body that extends down to 15-20km (Yang et al., 2022). The crustal thickness of the northern Mississippi embayment and the gravity anomalies shows a gravity high at the edge of transect across the Mississippi embayment, near the study area (Liu et al., 2017). The presence of anomalous crustal bodies indicated from 2D forward modeling in cross section A-A' (Figs. 21 and 22) may account for the positive gravity anomaly near the southeastern cross section of the embayment of Liu et al. (2017). The results of prior modeling of the Mississippi embayment were used as an assumption to produce two models for cross section A-A', one focusing on crustal structures within the upper crust, and another with anomalous body of contrasting density and magnetic susceptibility within the lower to middle crust. These parameters were explored through multiple models that resulted in two working models in profile A-A', one with near surface, relatively smaller vertical and horizontally oriented heterogenous crust with higher magnetic susceptibility than the surrounding crust (Fig. 21), as well as a deeper and larger body (Fig. 22). Profiles for cross section B-B' (Figs. 23 and 24) consistently resulted in a similar geometry resembling the ring dike complex or caldera collapse of diatreme structures of the St. Francois Mountains (Kisvarsanyi, 1980, 1981; Sides et al., 1981). This could suggest that these structures could be spatially and temporally related to the Eastern Granite Rhyolite province to the northwest owing to its relative stability as a cratonic platform through time. The location of the Black Warrior basin on the Alabama promontory is separated from the St. Francois Mountains of the Ozark Plateau by the Mississippi Valley graben, which does not have exposed outcrops of Precambrian rocks, though it is assumed the underlying strata of the Mississippi Valley graben is the same composition as the igneous terrane present in the St. Francois Mountains on the Ozark Plateau (Braile et al., 1986). Assuming the Eastern Granite Rhyolite province extends to the subsurface of the Black Warrior basin would be valid interpretation and consistent with the assembly of Laurentia by exotic terranes during the Proterozoic (Whitmeyer and Karlstrom, 2007). Due to the non-unique nature of geophysical models these are only interpretations of possible scenarios. There may be other interpretations that exist and with further study of the area it may be possible to refine the gravity and magnetic models to better understand the subsurface.

Interpretations from the above models (Figs. 21 and 24) suggest there exists subsurface structures like those found in the Eastern Granite Rhyolite province that is exposed in the St. Francois Mountains (Kisvarsanyi, 1980; 1981). An alternative interpretation could be a combination of these models to produce a structure like that found in the central Illinois basin (McBride et al., 2016; 2017), or along the Rough Creek graben that may have exploited preexisting crustal weakness during the Precambrian Iapetan rifting or Proterozoic tectonic events. (Hildenbrand et al., 1996). With no major deformation within the Mississippian or Cretaceous strata, any intrusive events likely predated their deposition or were too deep to have a considerable impact. Intrusive events of the mid-continent include the Devonian age Avon Volcanic district in southeastern Missouri and intrusions and the Permian age Hicks Dome (Zartman et al., 1967). Additionally, Cretaceous related intrusions in Arkansas and Mississippi may be related to the passage of the Bermuda hotspot over the Mississippi Valley graben or Reelfoot rift (Cox and Van Arsdale, 2002). The anomalous crustal bodies indicated in cross sections of gravity and magnetic anomalies of the study area (Figs. 21-24) may be related to the other intrusive events of the mid-continent and likely followed preexisting basement faulting. Anomalous crustal bodies could have been emplaced during Precambrian rifting, Proterozoic terrane accretion, or the Grenville orogeny.

The orientation of the gravity and magnetic maxima resemble orthogonal lineations of the St. Francois Mountains. Stress may have been concentrated along these crustal boundaries produced by basement faults and crustal heterogeneity to created faults, which developed Precambrian topography into a series of basins and highlands. As multiple periods of compression and extension reactivated these faults it may have created differences in the depth of the basement in polygonal sections. The faults likely propagated through Precambrian and early Paleozoic strata, as can be seen in the Illinois basin (Nelson and Lumm, 1992) and the southern portion of the Black Warrior basin (Thomas, 1985; Thomas, 1988; Thomas, 2010). Grabens in the Precambrian basement have been recorded in the northern Black Warrior basin (Johnson et al., 1994). There are similar faults in the basement found in an extensional setting of the southern Black Warrior basin (Groshong et al., 2010), as well as transpressional structures within Carboniferous strata that displace the Precambrian basement near the Ouachita-Appalachian junction (Hale-Erlich and Coleman, 1993). However, the faulted basement in the Black Warrior basin to the south, and along the southwestern Laurentian margin, would have likely defined sedimentary deposition of the Cambrian and continued to define sedimentation throughout the Paleozoic as extensional and compressional events altered the stress regime along pre-existing crustal structures (Thomas, 1972b; 2010). The regional scale of the response to flexure induced by far-field stress due to thrust front loading within the Black Warrior foreland basin formed due to viscoelastic deformation (Quinlan and Beaumont, 1984; Hines, 1988) and local crustal weaknesses would likely further exacerbate brittle to elastic deformation. The variability in the thickness and differing crustal composition of the crust may have further concentrated stress along faults may have caused a local elastic response. Precambrian igneous bedrock topography may have also influenced the depositional settings of sedimentary units in the St. Francios mountains (Kisvarsanyi, 1977). Pre-existing topography may have produced a similar effect on the deposition of Precambrian and Cambrian sediments of a pre-Black Warrior basin on the Alabama promontory, as Cambrian and Ordovician isopach maps follow similar trends found in Paleozoic strata (Thomas, 1972b).

Faulting is evident to the north and south of the study area that is related to structures within the crust which were active in the Precambrian and Paleozoic. Displacement on Rough Creek graben shows listric faulting that penetrates Precambrian crystalline basement and varying isopach of pre-Knox strata, with pre-Knox strata 8000m thicker on down faulted side equating to a 2-3 km of offset (Nelson and Lumm, 1992). A 2-3 km offset is within the realm of 2D forward gravity and magnetic models produced in this study (Figs. 21 and 24). Mafic igneous bodies intruded during the Grenville orogeny or Proterozoic terrane emplacement that formed the Eastern Granite Rhyolite province, may have caused a thickening of the crust beneath the Illinois basin, which may have contributed to subsidence of the basin. These intrusions may have produced similar effects further to the south, creating contrasting crust lithologies (Chen et al., 2016; Marlow, 2021). Detachment faults within the Pennsylvanian units of the Black Warrior basin may be due to pre-existing faulting and crustal structures as the Ouachita orogeny imprinted over the Alabama-Oklahoma transform margin of the southern edge of Laurentia. The crust in this region may have been less attenuated than the crust beneath the extended margin of the Appalachian front, thus would possess less structural rigidity and would be more prone to flexural bending (Groshong et al., 2010). Geologic mapping has revealed subtle folding in northwestern Alabama and northeastern Mississippi (Merrill et al., 1988, Chapter 1 of this report) that parallels the Ouachita margin. It is unknown if these folds were caused by basement faults, or if these folds represent the thin-skin style deformation similar to that found southwest of the study area in Pennsylvanian strata of the Black Warrior basin. Thin-skinned faults are generally attributed to extension caused by flexural bending of the crust under thrust loading (Harry and Mickus, 1998). Deformation of the region to the southwest has exhibited extensional deformation through the Paleozoic (Thomas, 1972b) likely due to lower crustal attenuation as it is assumed to be a relatively undeformed craton, unlike the presumed rifted margin that underlies the Appalachian from to the east (Harry and Londono, 2004). With no boreholes deep enough in the surrounding region to penetrate the

Precambrian crystalline bedrock it is impossible to provide a feasible constraint based on known lithologic composition to the model.

The use of the term Lowndes-Pickens block may be erroneous as it acted less like a discrete block and more as a structurally defined system of grabens that were defined by preexisting basement faults that reactivated by tectonic forces in an evolving stress regime as the Ouachita orogeny continued to converge with the southern Laurentian margin. The northwestern region of the Black Warrior basin was impacted by both the far-field forces and subsurface crustal heterogeneities that may behaved differently in response to the stress than the region to the southwest, closer to the orogenic front. Evolving stress regimes throughout the Mississippian likely caused the movement of these faults, localizing the stress along boundaries as other regions acted more rigidly in response to the flexural bulge induced by the obduction of the Sabine terrane. The oblique angle of convergence may have resulted in transpressional or transtensional forces acting on these grabens, leading to diachronous movement along Precambrian faults. These changes could manifest similar to those along the southern edge of the Illinois basin with contractional and extension duplexes that formed during reactivation through time based on the pre-existing structures to accommodate movement in response to far-field stress (Duchek et al., 2004).

If the offset in the Precambrian basement on the geophysical models for the cross sections A-A' (Figs. 21 and 22) and B-B' (Figs. 23 and 24) are assumed to correlate to other regions in the basin it may inferred that the changes in the anomalies in the other parts of the Black Warrior basin are due to similar structural features. The geophysical trends record possible basement faults and anomalous bodies where their extent seems to coincide with the boundaries of the isopach maps of the sedimentary cycles and sandstones within the Pride Mountain Formation (Figs. 25-28), as well as the extent of the Pierce Siltstone (Fig. 29) and Hartselle Sandstone (Fig. 30). These isopach maps may represent the preservation of sedimentary units as the far field stress induced reactivation of pre-existing faults with the basements, further accentuated by the contract in composition of the basement with anomalous crustal bodies within the crust. The crustal structures may have been the location of increased accommodation space as the subsidence along either the prograding deltaic system from the southwest or northwest due to the transition between extension and compressional forces within the basin it experienced thrust loading induced by the encroaching Ouachita orogeny.

Geologic mapping of the modern landscape has revealed the basal Cretaceous erosional surface that likely represent paleotopgraphy within the Cretaceous (Fig. 11), which follows the pre-existing deformation within the Mississippian strata. Subsidence structures found along the Tennessee River (Szabo, 1992) coincide with the transitions from gravity and magnetic maximum and minimum, which 2D forward modeling (Figs. 23 and 24) has interpreted to be displacement along the Precambrian basement. If paleochannels in the Cretaceous have the same orientation as Mississippian folds, this could suggest paleotopography was influenced by prior bedrock deformation associated with the Ouachita orogeny. The main trunk of the modern Tennessee River seems to be following similar paths of fluvial systems that shaped the landscape of the Cretaceous, which suggests the persistent effect of ancestral drainage networks on the modern landscape.

Analysis of modern drainage basins in western Tennessee and northeastern Mississippi has revealed asymmetry in drainage patterns, which may suggest neotectonic tilting (Cox et al., 2001; Garrote et al., 2006). There may be relationships between recent tectonic activity to the gravity and magnetic anomalies that correspond to basement faults and regional crustal structures. The asymmetry of basins may also be induced by regional changes due to dynamic topography (Liu, 2015). A study of basin asymmetry analysis could be expanded further to the east to include northwestern Alabama and southcentral Tennessee and compare gravity and magnetic anomalies to the location of active basin asymmetry. This may help to resolve the impact of current deformation to existing structures and allow to better infer the relationship of modern drainage systems to the recent deformation related regional structures and elucidate the possibility of reactivated basement faults. However, expanding basin analysis to the east would incorporate basins located in Paleozoic sedimentary units rather than unconsolidated Coastal Plain sediments, which may not be as affected by recent tectonic activity.

The study area has been relatively understudied as it is located on the periphery of other structures such as the Mississippi Embayment (Yang et al., 2022), the East Tennessee Seismic Zone (Powell and Thomas, 2016), and the Ouachita orogenic belt. There have been more focused studies on the upper crustal structure beneath the Black Warrior basin further to the south as it contains a greater capacity for hydrocarbon resources (Pashin and Gastaldo, 2009; Groshong et al., 2009). Due to the relative lack of geophysical studies in the study area there exists a substantial gap in the knowledge about the composition and structure of crust and upper mantle. By improving the resolution and refining models for the geometry and lithologies of crustal structures it may be possible to map how pre-existing basement structures are distributed in the subsurface. It may then be possible to also infer how they interact with evolving tectonic stress regimes of both the past and present, as exploitation of preexisting structural features may continue to influence the landscape to this day. This is a complex problem that incorporates a multitude of factors such as topographic, tectonic, eustatic conditions that existed at the time of the deposition of the Mississippian clastic wedge. This study demonstrates it requires a multidisciplinary approach to create a cohesive understanding of regional history to unravel the interplay of subsidence, sedimentation, and structure in the formation of clastic wedge within a foreland basin. Sedimentation records only a portion of this complex past. With higher resolution gravity, magnetic, and seismic data, in addition to added geophysical analysis, such as band pass filter and deep imaging, it may be possible to distinguish the location of crustal structures more accurately, as well as interrogate their geometries, composition, and depth, as well as relationship to deformation, both past and present.

CONCLUSION

The Late Mississippian represents a shift to clastic sedimentation within the Black Warrior basin. Due to the unknown nature of the crustal composition and structures underlying much of the Black Warrior basin maps of gravity and magnetic anomalies were assembled. 2D geophysical models were constructed to assess the feasibility of various crustal compositions and structures. Geophysical boundaries were found to align with the boundaries of isopach maps of Late Mississippian sedimentary deposition within the northwestern Black Warrior basin. The relationship between geophysical features interpreted to reflect crustal structure (i.e. faults) and clastic unit thickness suggests that faults and crustal heterogeneity at a subregional scale may have been responsible for the evolution of the foreland basin and preservation potential of sedimentary units within the basin may be influenced by pre-existing faulting and crustal structures. Crustal heterogeneity is a location prone to fault reactivation and fault development in later, overlying strata. Previously studies suggested the region sits within an area of relative cratonic stability, however far-field forces likely influenced the development of structures and stratigraphic sedimentation beginning in the Mississippian and continuing through the Cretaceous. In response to these forces the region experienced diachronous concentration of stress along preexisting structural weaknesses and contrasting lithologies, which may have induced topographical changes that influenced sediment deposition and accumulation. The geological and geophysical maps produced by this study provide a higher resolution view of the geophysical characteristics of the study area, and the possible influence of pre-existing structures on stratigraphic architecture within a foreland basin, as well as the persistent effect deformation may play in the role of topographic development.

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SUMMARY

Geologic mapping of the Bishop 7.5-minute quadrangle in Colbert County, Alabama and a portion of Tishomingo County, Mississippi was completed to provide higher resolution of surficial geologic data, with particular interest in the basal Cretaceous unconformity. This contact was modeled using ArcMap and restored to a pre-subsidence orientation. The alignment of folds indicates a possible structural control by bedrock deformation within Mississippian strata, which may have influenced the formation of the paleotopographic surface during the Cretaceous as well as modern surface fluvial geomorphology to a greater degree than previously known. In addition to field mapping efforts, gravity and magnetic data were used to create geophysical maps and crustal models of the region. These maps were the principal step in geophysical modeling allowing for the delineation of subsurface features. These crustal structures aligned with trends in Mississippian clastic deposition, which suggests deformation associated with the Ouachita orogeny may have exploited pre-existing structural features in the subsurface. These crustal structures may have influenced clastic depositional trends during the Late Mississippian as flexure within the foreland basin due to thrust loading of the encroaching Ouachita front as stress accumulated along lithospheric heterogeneities. Additional geophysical surveys and analysis could further constrain the depth and boundaries within the subsurface, which would assist in formulating a better understanding of the relationship between deformation and deposition within the Black Warrior foreland basin.

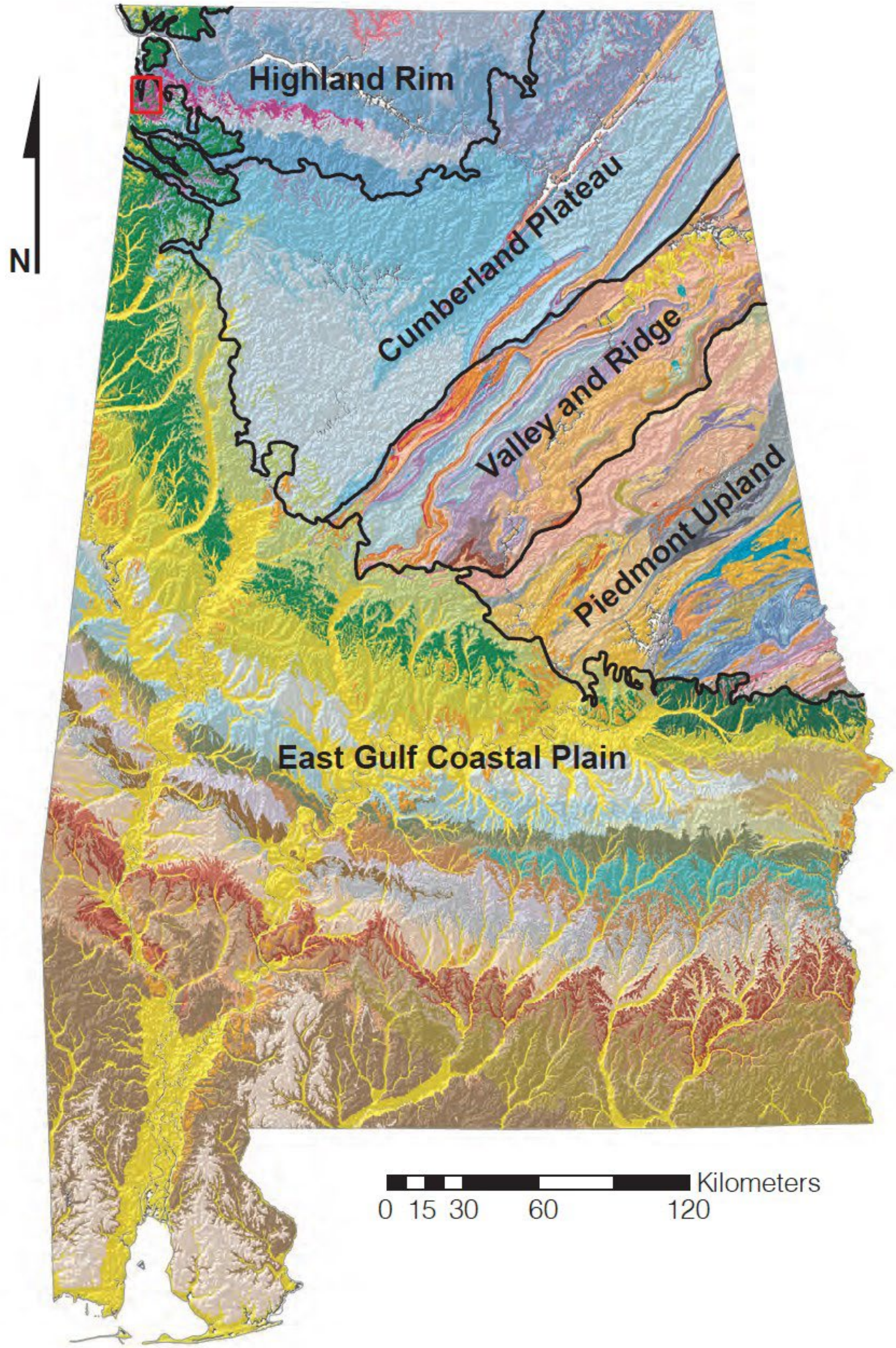


Figure 1. Physiographic map of Alabama with location of the Bishop quadrangle outlined in red, modified from Sapp and Emplaincourt (1975) and Szabo et al. (1988).

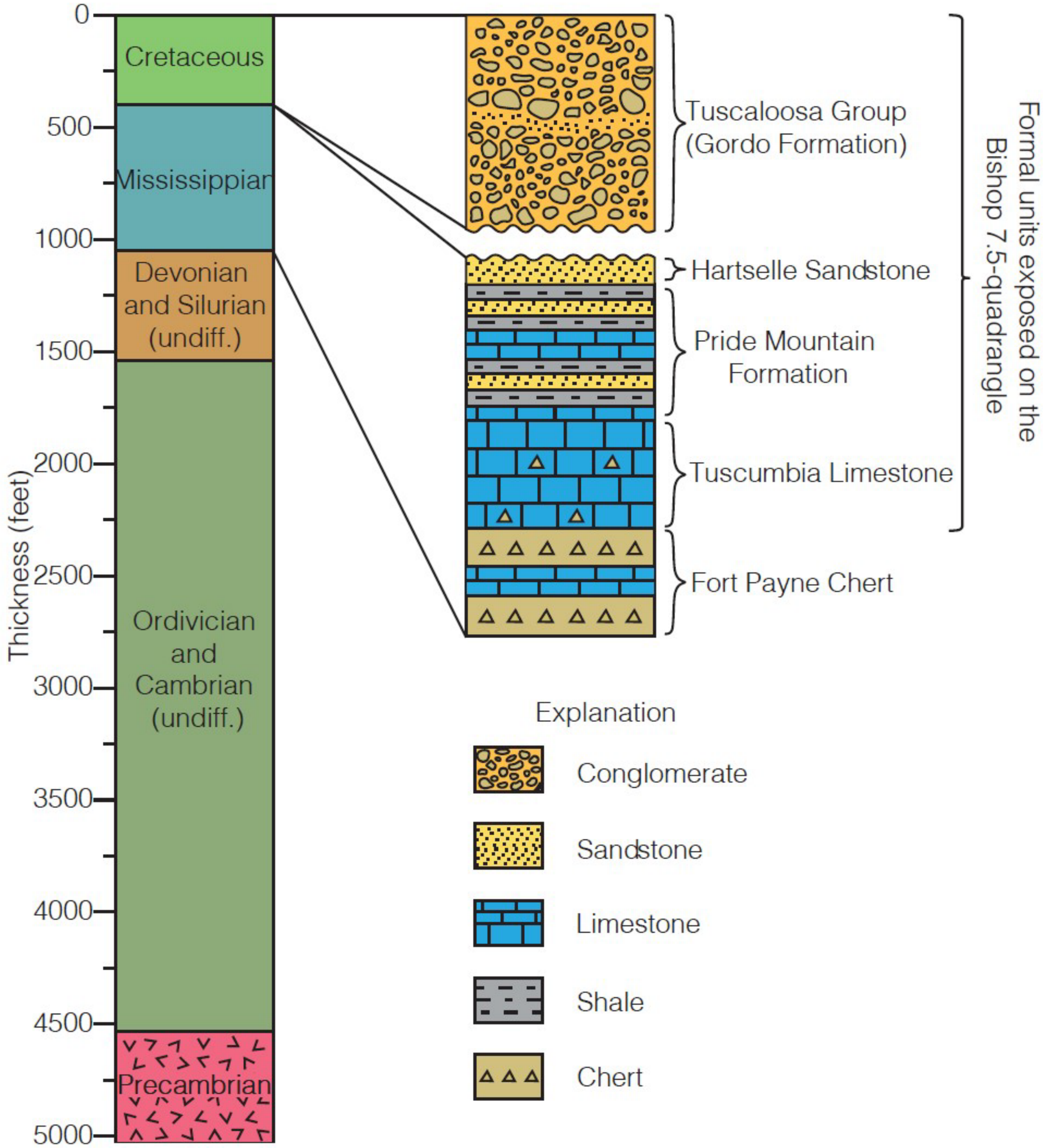


Figure 2. Schematic for the stratigraphy of the Bishop Quadrangle.

Figure 3. Outcrops of Tuscumbia Limestone are pictured above. A) The massively bedded Tuscumbia Limestone with chert inclusions on the Margerum quadrangle, north of the Bishop quadrangle. B) Tuscumbia Limestone, interbedded calcarenite has been preferentially eroded.

Figure 4. Outcrops of sandstone within the Pride Mountain Formation. A) The thick to thinly bedded sandstones of the Pride Mountain from the Barton quadrangle, near the eastern boundary of the Bishop quadrangle. B) The quartzose finely grained Lewis Sandstone from the Bishop quadrangle at NW ¼ sec. 11, T. 4 S., R. 15 W.

Figure 5. Outcrops of limestone within the Pride Mountain Formation limestone are pictured above. A) Limestone in a creek bed at NE ¼ sec. 7, T. 5 S., R. 14 W. B) Limestone located at NE ¼ of SE ¼ sec. 32, T. 4 S., R. 14 W.

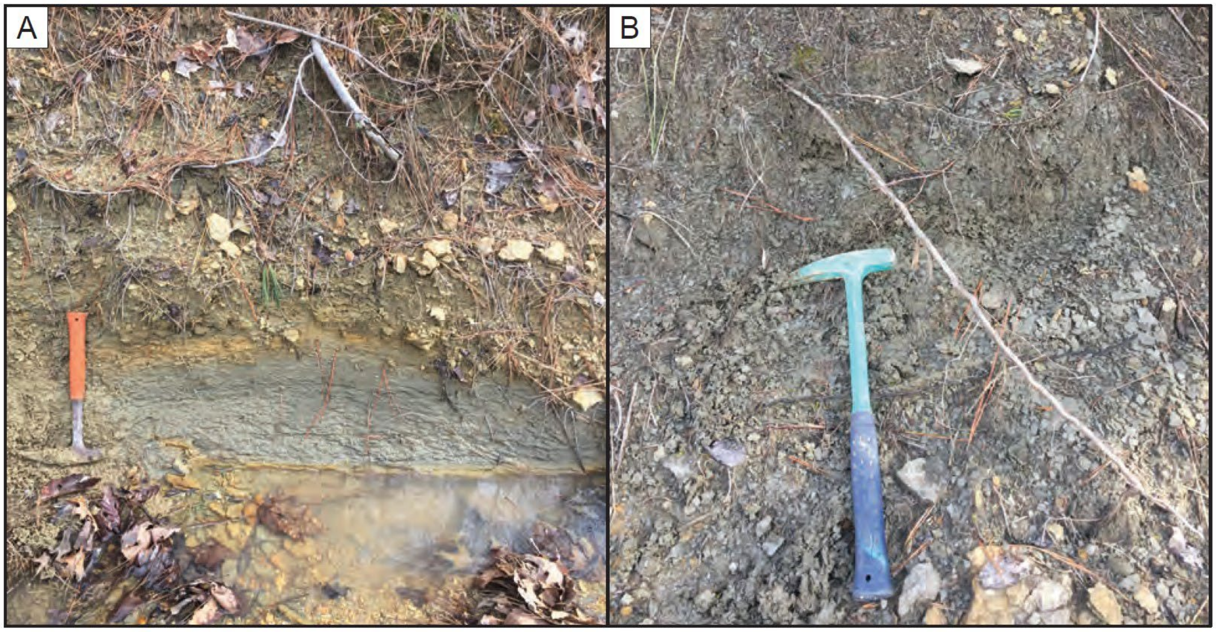


Figure 6. Outcrops of shale within the Pride Mountain Formation. A) Thinly bedded shale at SE ¼ sec. 6, T. 4 S., R. 14 W. B) Shale overlying the oolitic, asphaltic limestone determined by Clark (1925) at SE ¼ sec. 1, T. 4 S., R. 15 W.

Figure 7. Outcrops of the Hartselle Sandstone. A) Thinly bedded with interbedded shales at SE¼ sec. 18, T. 5 S., R. 14 W. B) Thick bedded quartz arenite sandstone with planar cross bedding at location NE ¼ sec. 32T. 5 S., R. 14 W.

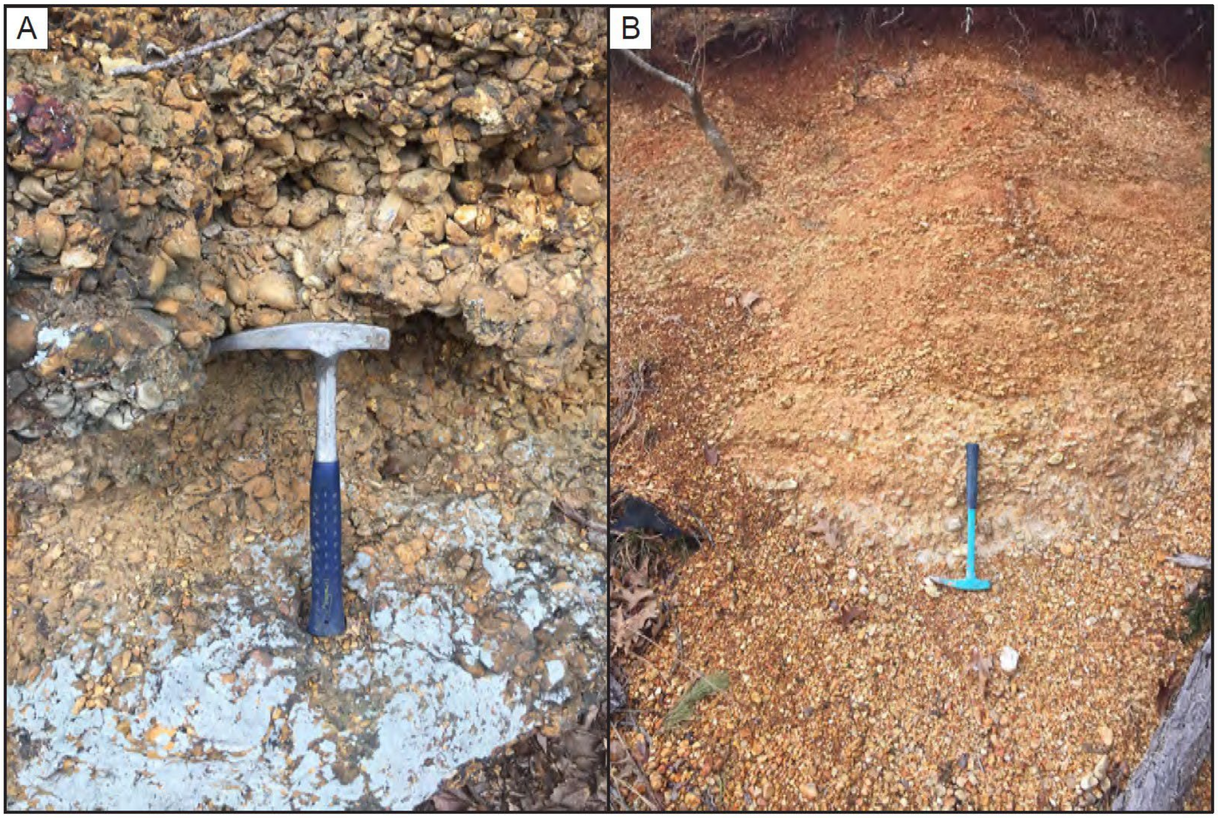


Figure 8. Outcrops of the chert conglomerate of the Tuscaloosa Group, typical of the western facies. A) Massively bedded clast supported of subangular to rounded chert conglomerate of the Tuscaloosa located at NW ¼ sec. 18, T. 4 S., R. 14 W. B) Chert conglomerate that contains higher portions of fine-grained sand, matrix supported in sections at SE ¼ sec. 6, T. 4 S., R. 14 W.

Figure 9. Outcrops of the "Little Bear Residuum" described by Mellen (1937). A) Mottled gray, red, and white clay, clay, silt, and tripolitic chert overlying conglomerate, contains plant fragments and carbonized wood. B) A large channel fill of quartz pebble conglomerate.

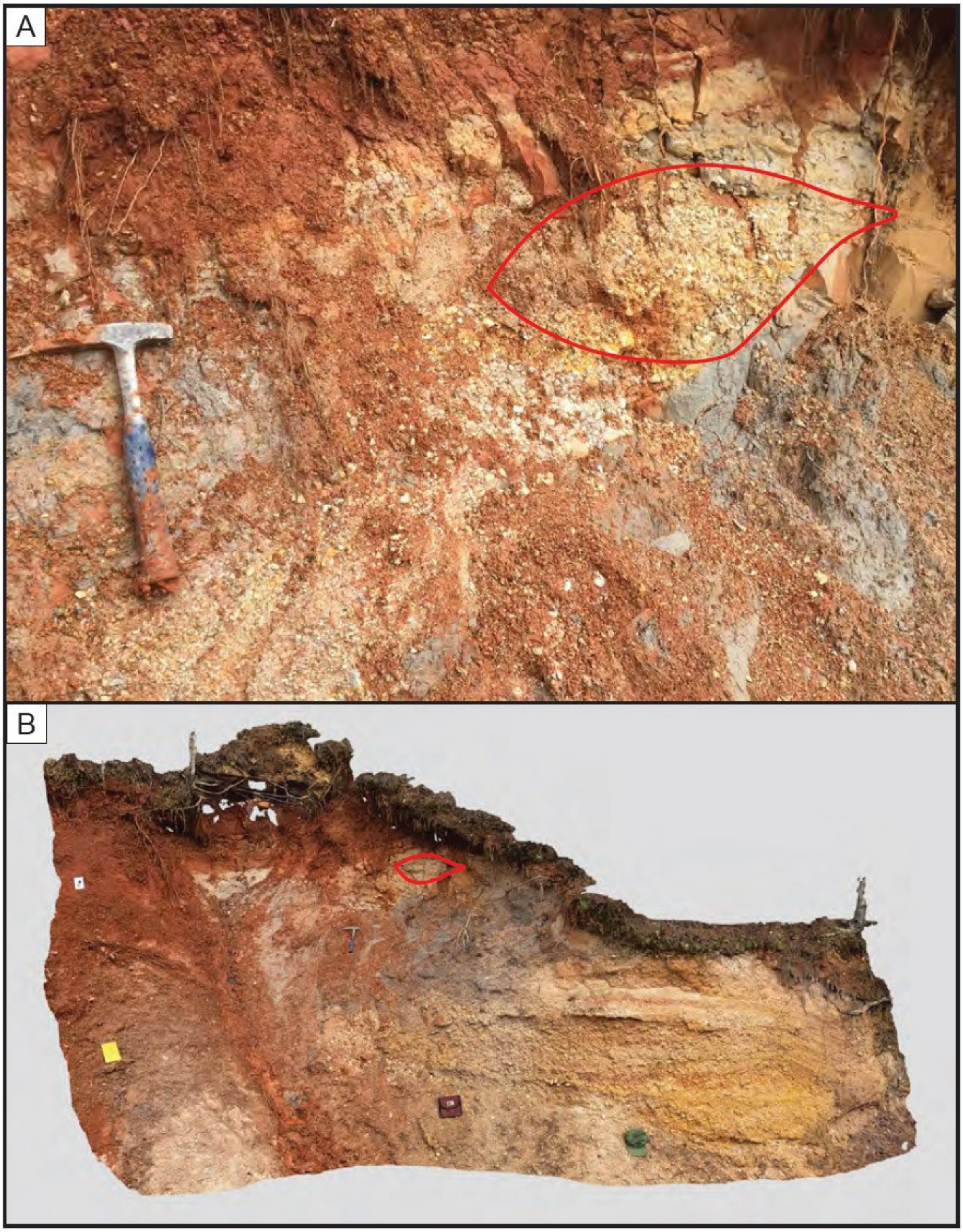


Figure 10. Basal Cretaceous unconformity (red outline shows chert lens). Mottled red, white, and gray clays suggest reworking of sediments later with chert deposition. B) 3D model of outcrop, available online at:
https://sketchfab.com/3d-models/tuscaloosa-w-texture1b52a3ee1f1c4e1d871d950e0cd70a71

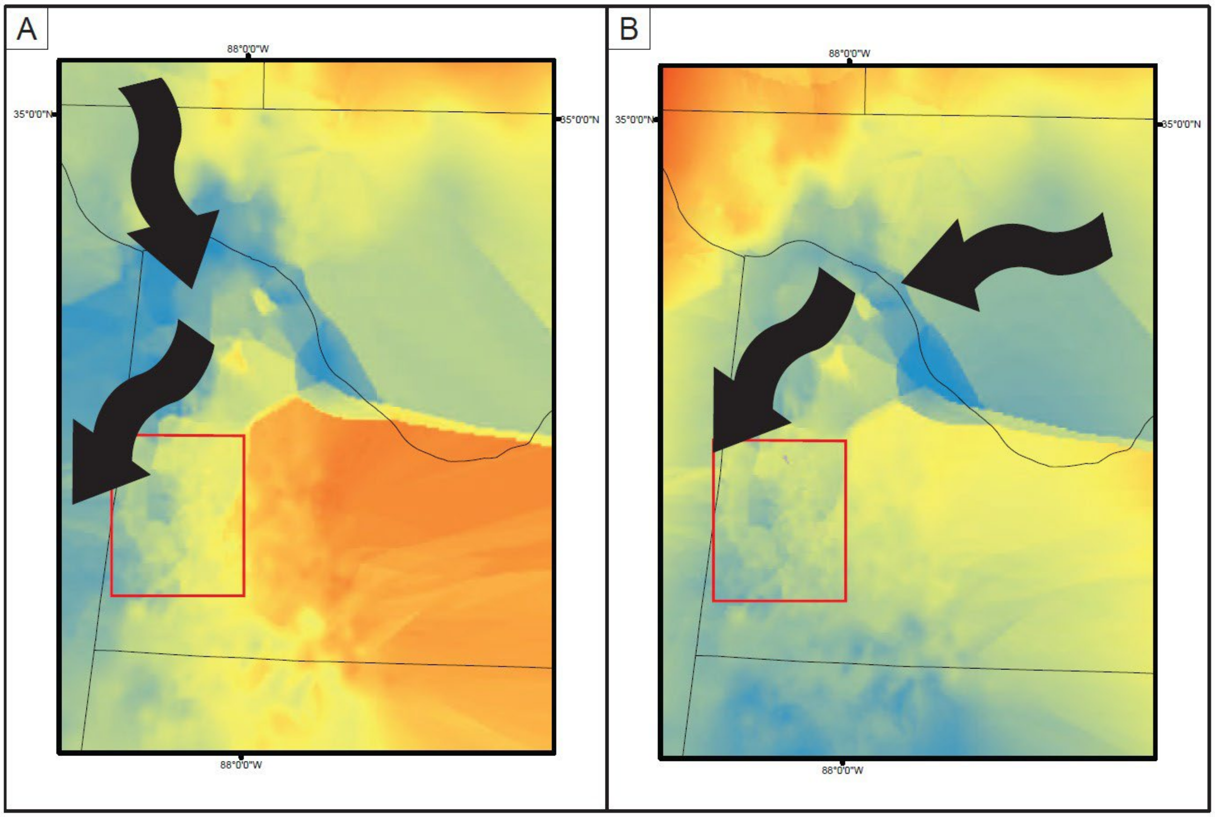


Figure 11. Reconstructions of paleotopographical surface based on mapping of the basal Cretaceous unconformity. Warmer colors are mapped as higher elevations. The Bishop quadrangle is outlined in red. A) Represents the unadjusted surface to the regional trend (arrows show possible distribution path of western Tuscaloosa facies). B) Represents the removal of the regional trend from the basal Cretaceous unconformity (arrows show possible distribution path of the eastern Tuscaloosa facies).

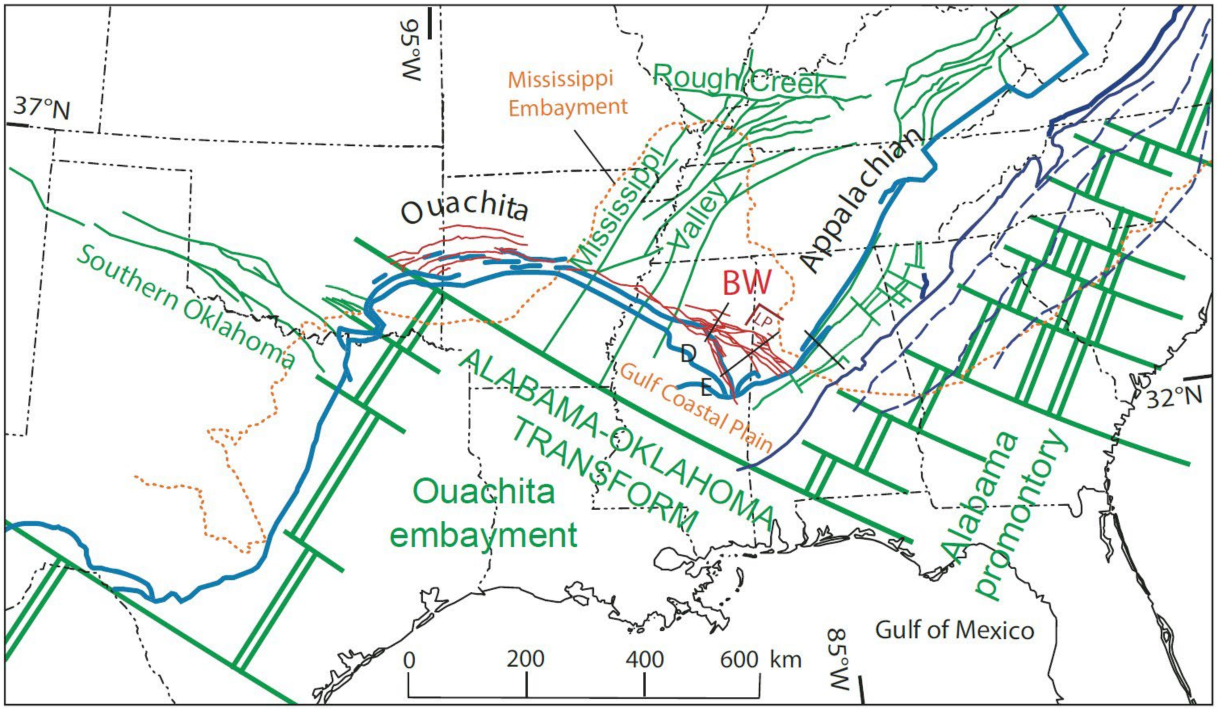


Figure 12. Map showing the Ouachita-Appalachian orogeny superimposed on the Laurentian margin post Iapetan rifting with intracratonic fault systems. The Black Warrior basin (BW) and the Lowndes-Pickens block (LP) are labeled in red. Red lines highlight basement faults in the orogenic foreland basins (Modified from Thomas, 2010).

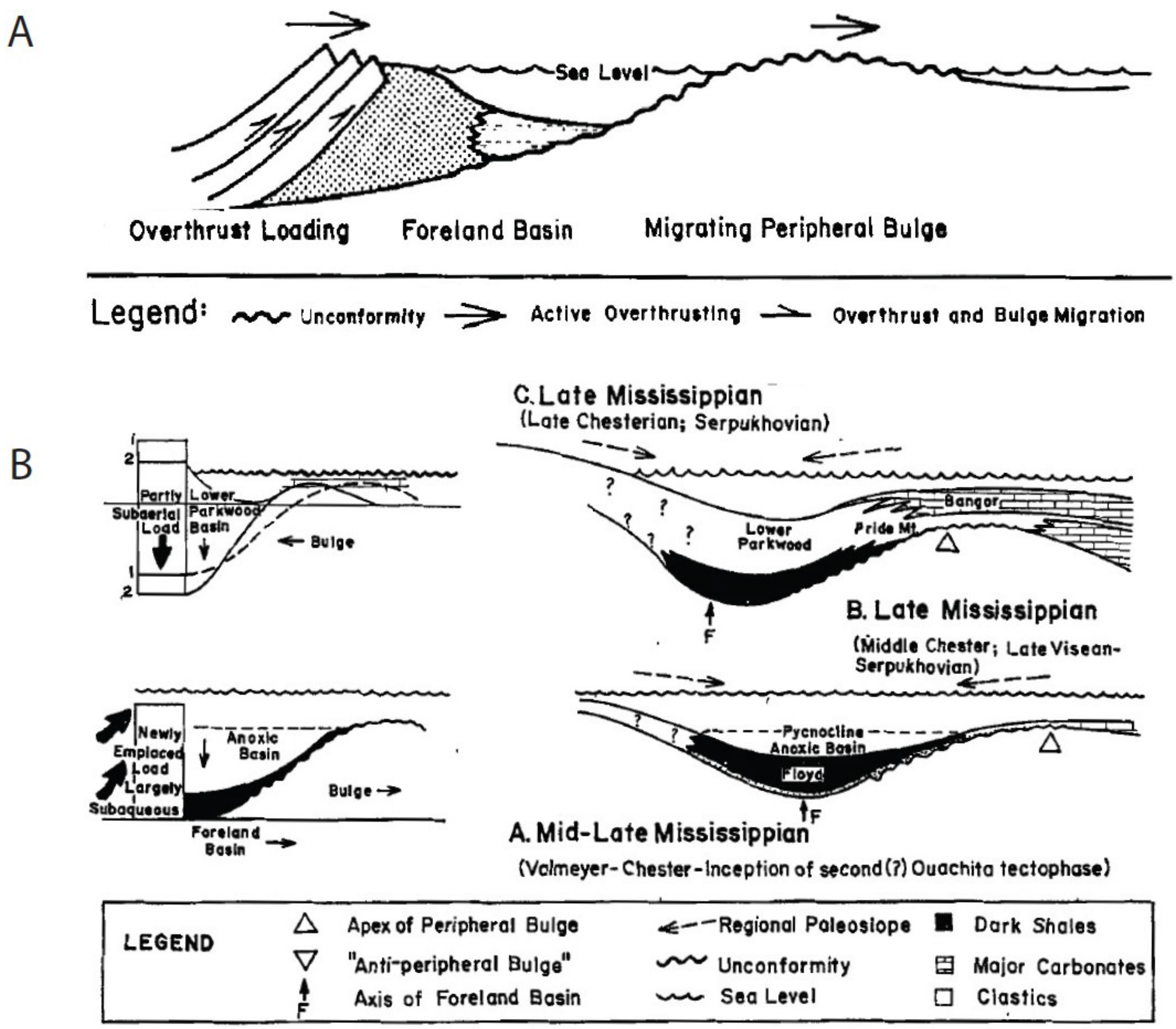


Figure 13. A) Schematic diagram displaying the formation of an orogenic foreland basin. B) Development of the Mississippian clastic wedge of the Black Warrior basin occured as migration of the Ouachita orogenic front continued to produce tectonic thrust loading and development of the peripherial bulge (figures adapted from Ettensohn and Pashin, 1993; Quinland and Beaumont, 1984).

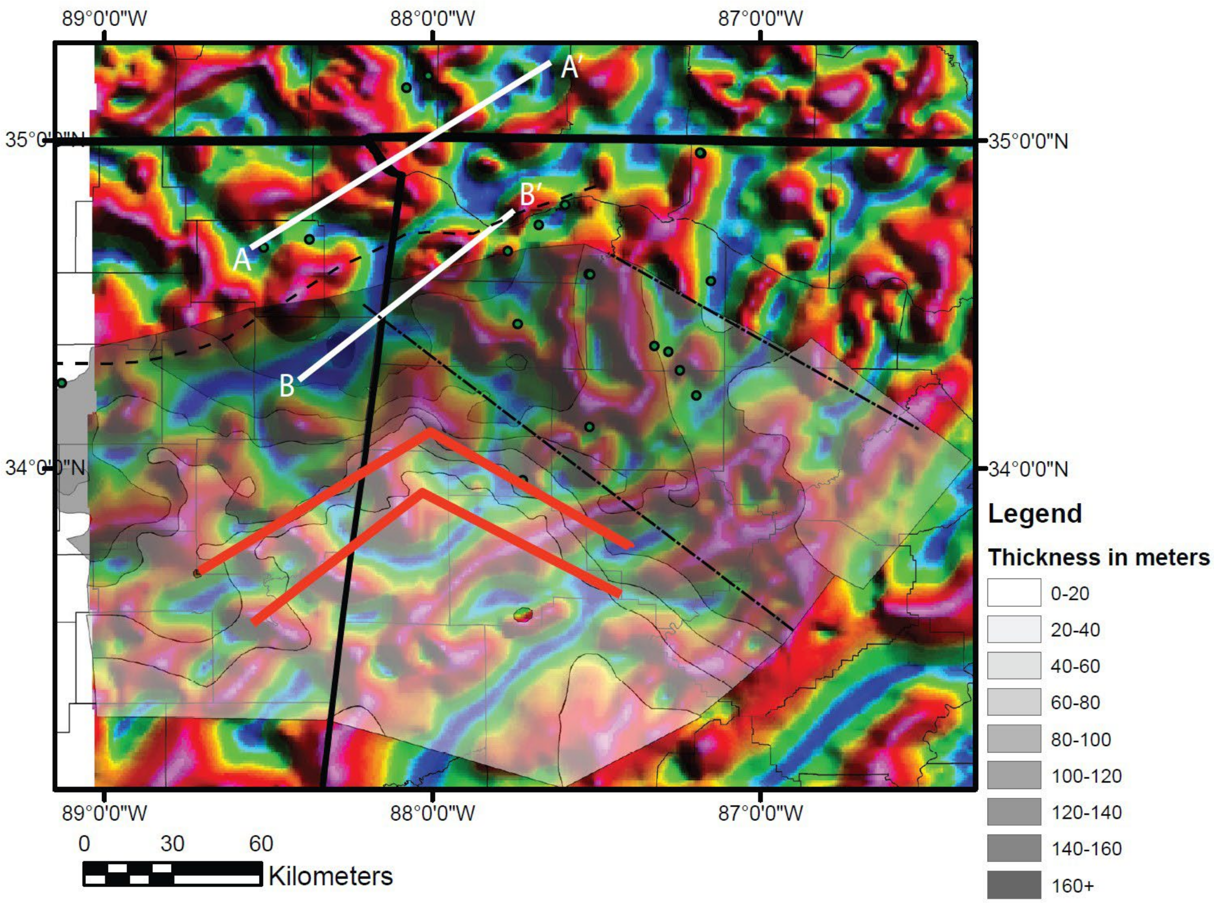


Figure 14. Vertical tilt derivative map with profiles A-A' and B-B'. Red mark the location of the trend from the Lowndes-Pickens block. Dashed line is the approximate erosional limit of the Pride Mountain Formation. Dotted and dashed line is the approximate limit of the Hartselle Sandstone barrier island depositional setting (isopach thickness modified from Kidd, 2008).

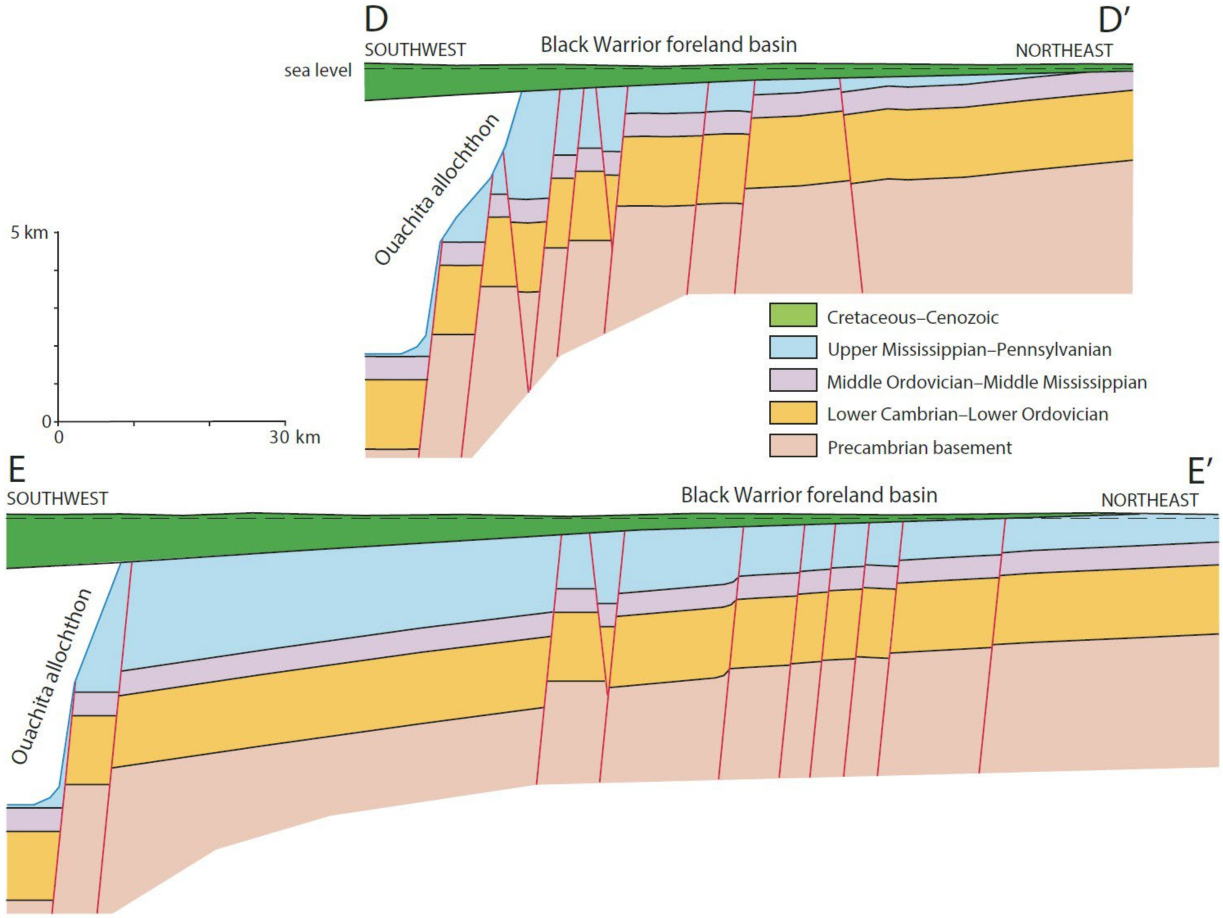


Figure 15. Cross section of the Black Warrior foreland basin and Ouachita from Thomas (1988), displaying faults in Precambrian basement faulting propagates into Paleozoic strata based on well logs and seismic data. The locations of the cross section are shown in figure 12.

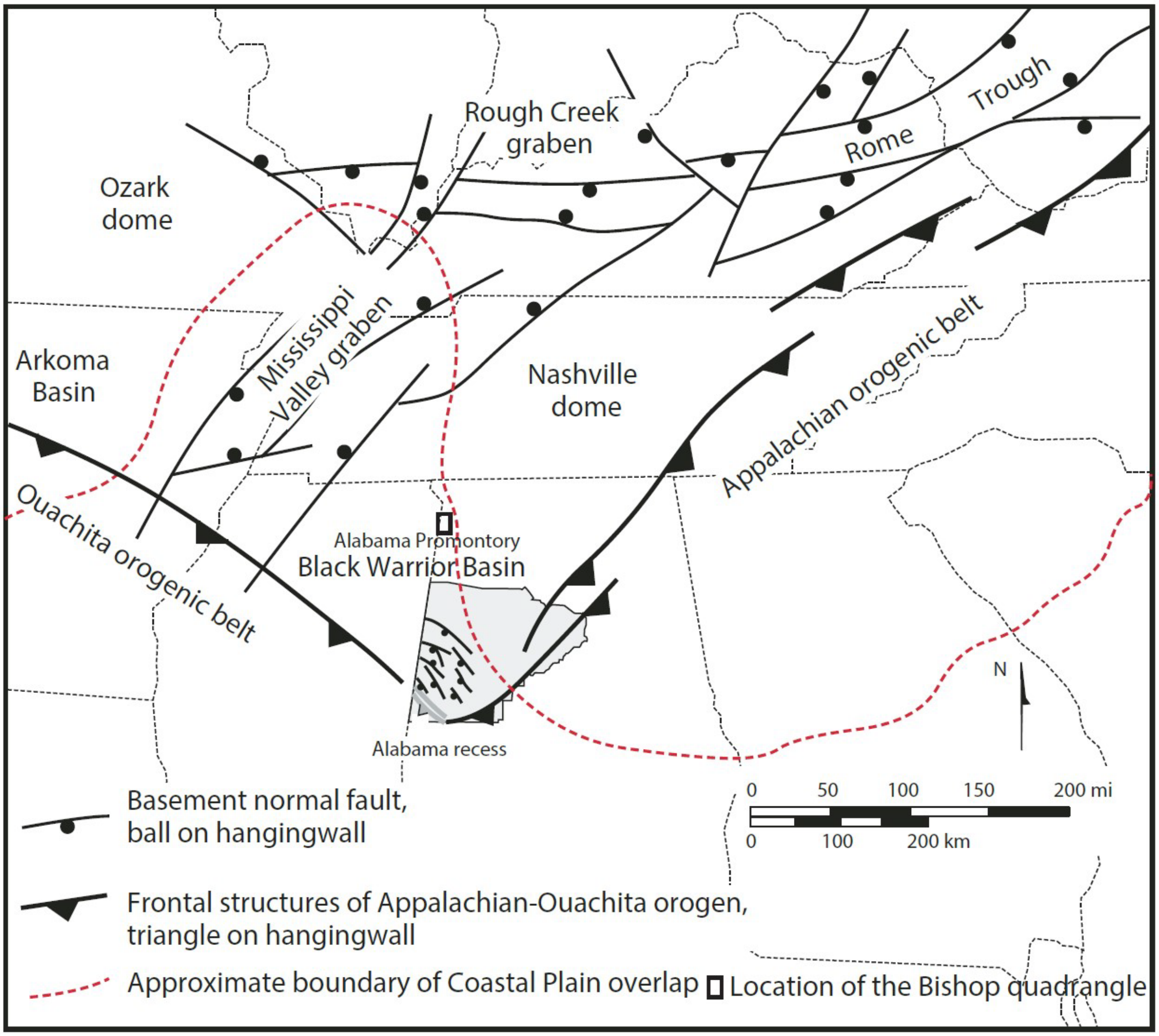


Figure 16. Normal faults within the southern Black Warrior basin with regional structural features of the mid-continent (modified from Groshong and others, 2010).

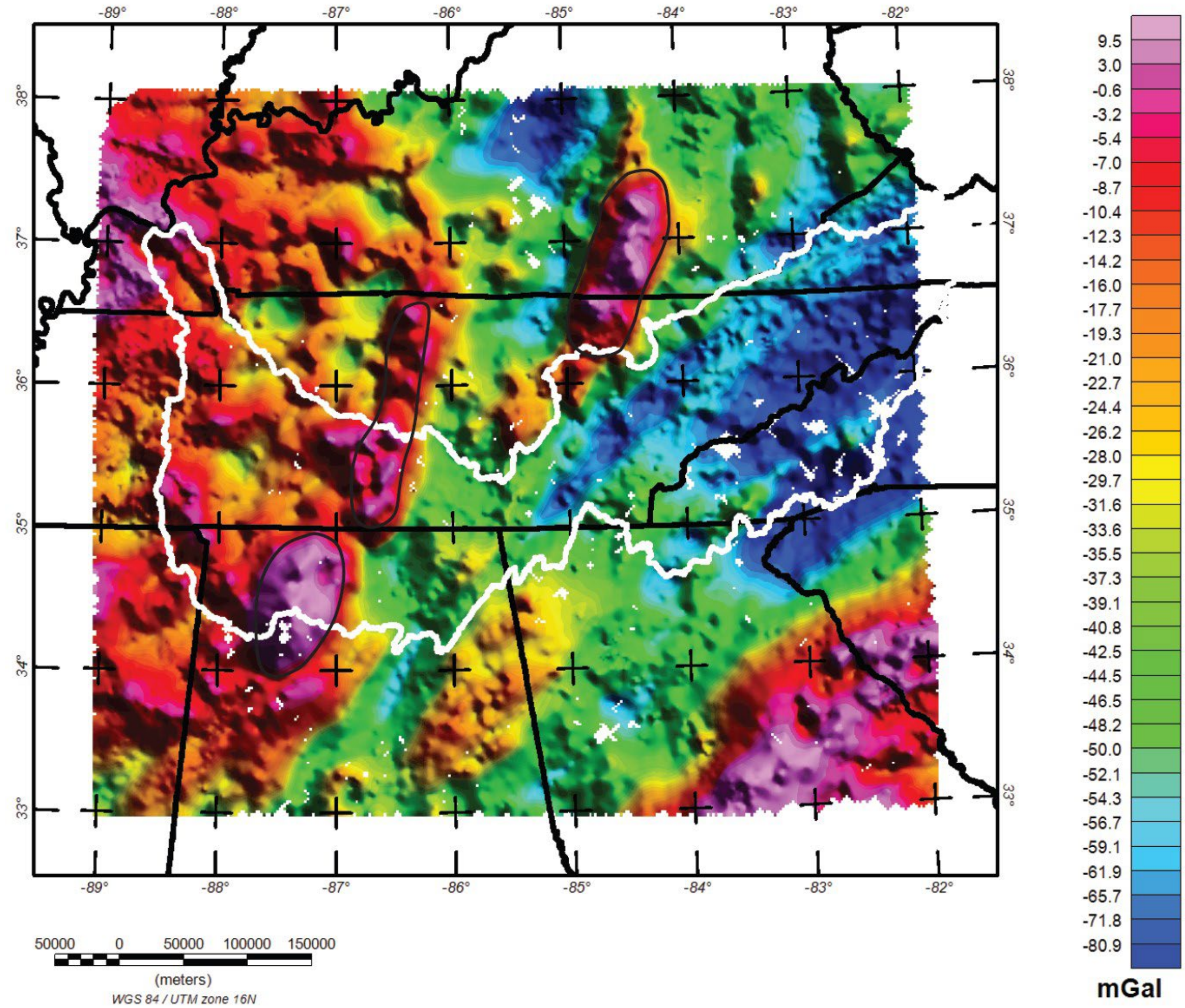


Figure 17. Complete Bouguer anomly map of the southeastern US. The East Continent Gravity high is outlined in black. The Tennessee River basin is outlined in white.

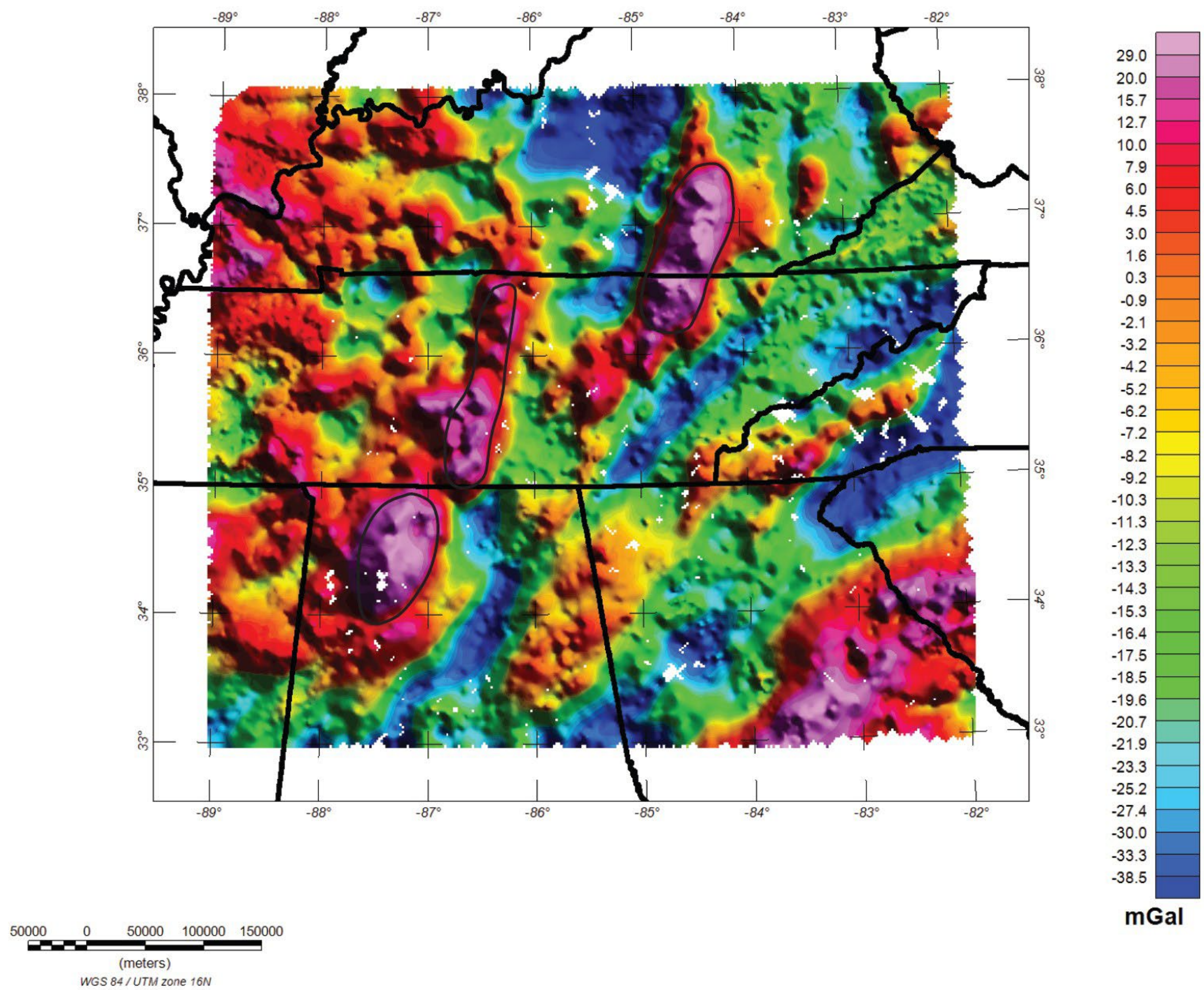


Figure 18. An isostatic residual anomaly map of the southeastern US. The East Continent Gravity high is outlined in black.

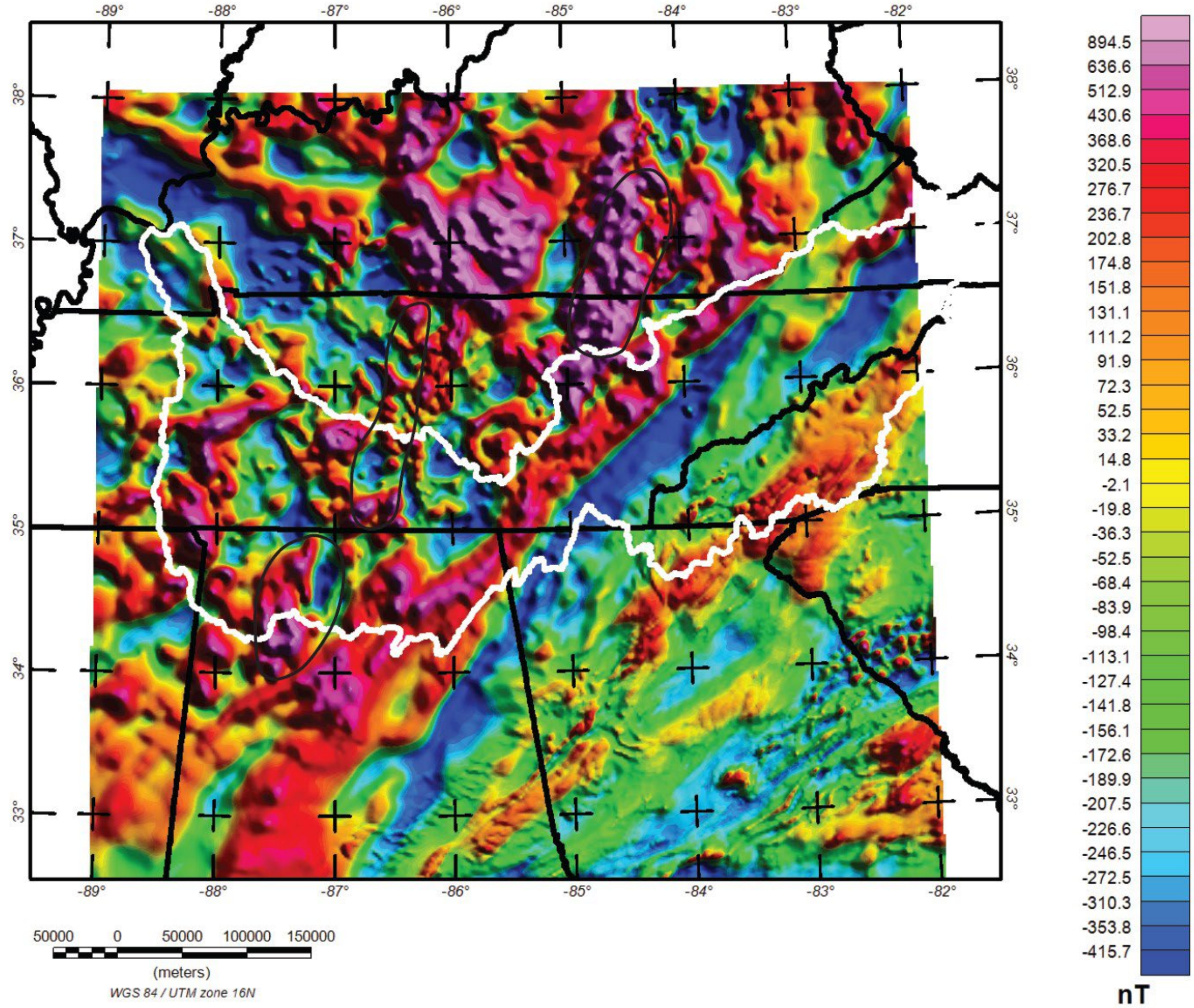


Figure 19. Reduced to poles magnetic anomaly map of the southeastern US. The Tennessee River basin is outlined in white.

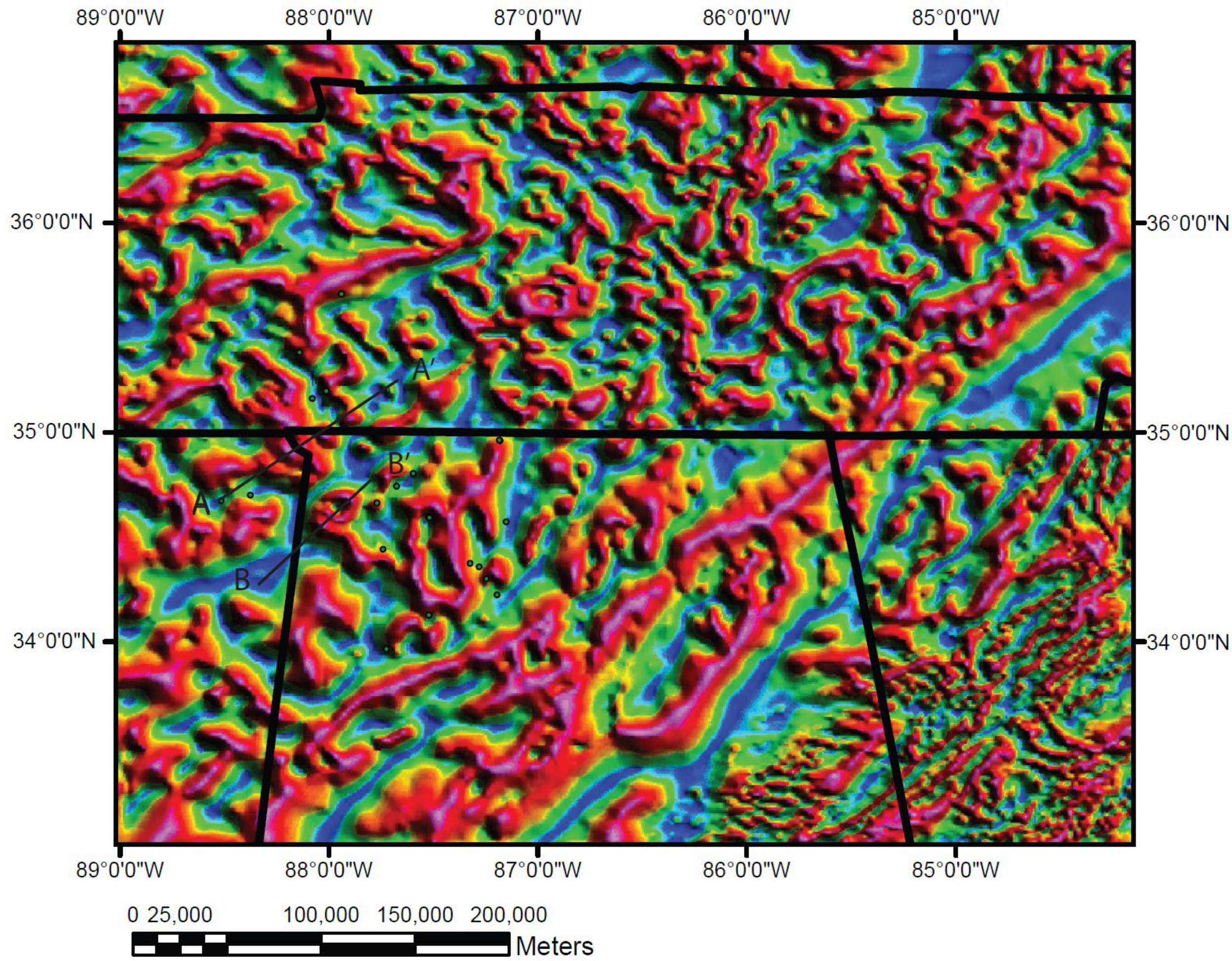


Figure 20. A vertical magnetic tilt derivative map anomaly map is shown above. Geophysical modeling profiles are labeled. Green dots are earthquake locations from the last 40 years.

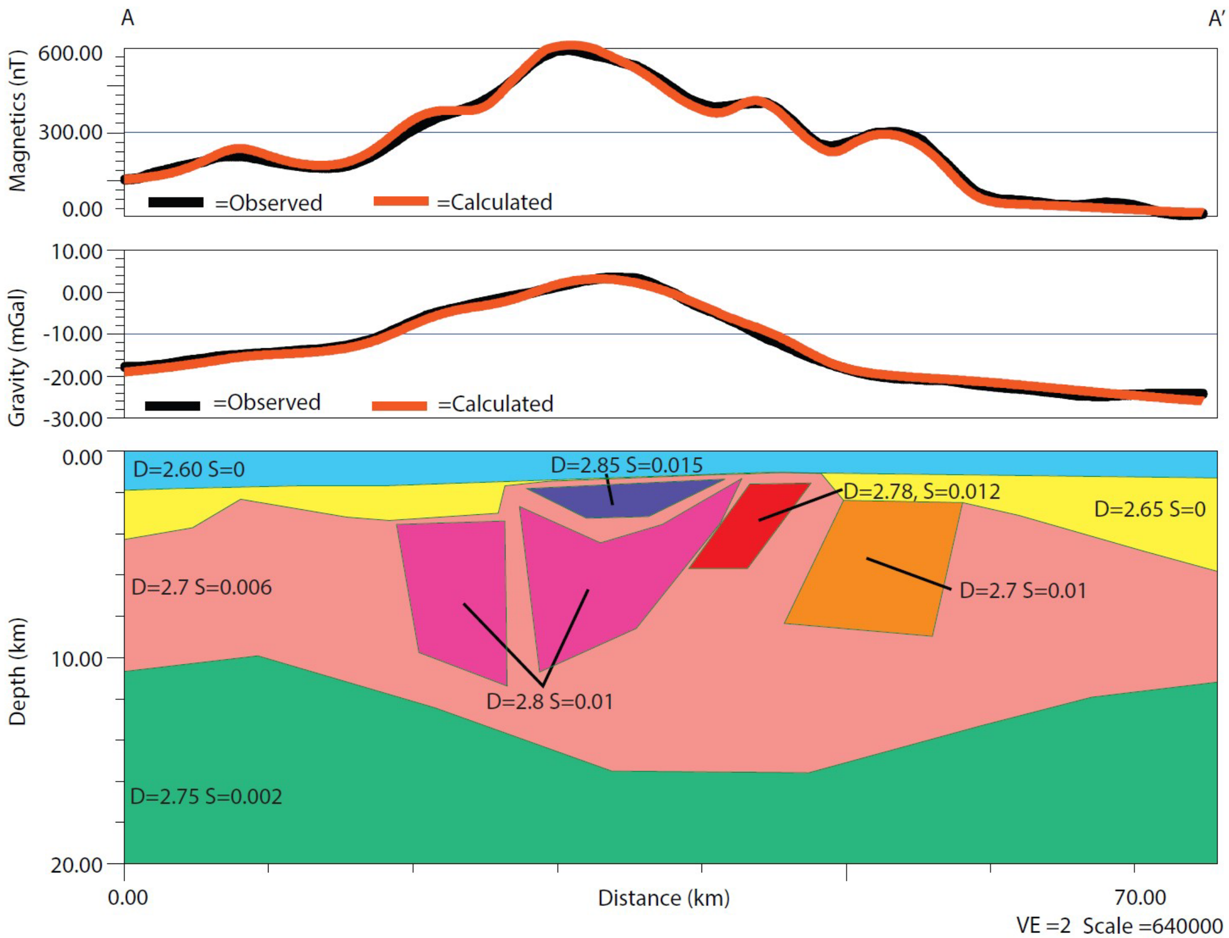


Figure 21. Model 1 of profile A-A' (Fig. 14) from 2D forward modeling using gravity and magnetic data with Geosoft Oasis Montaj GM-SYS program.

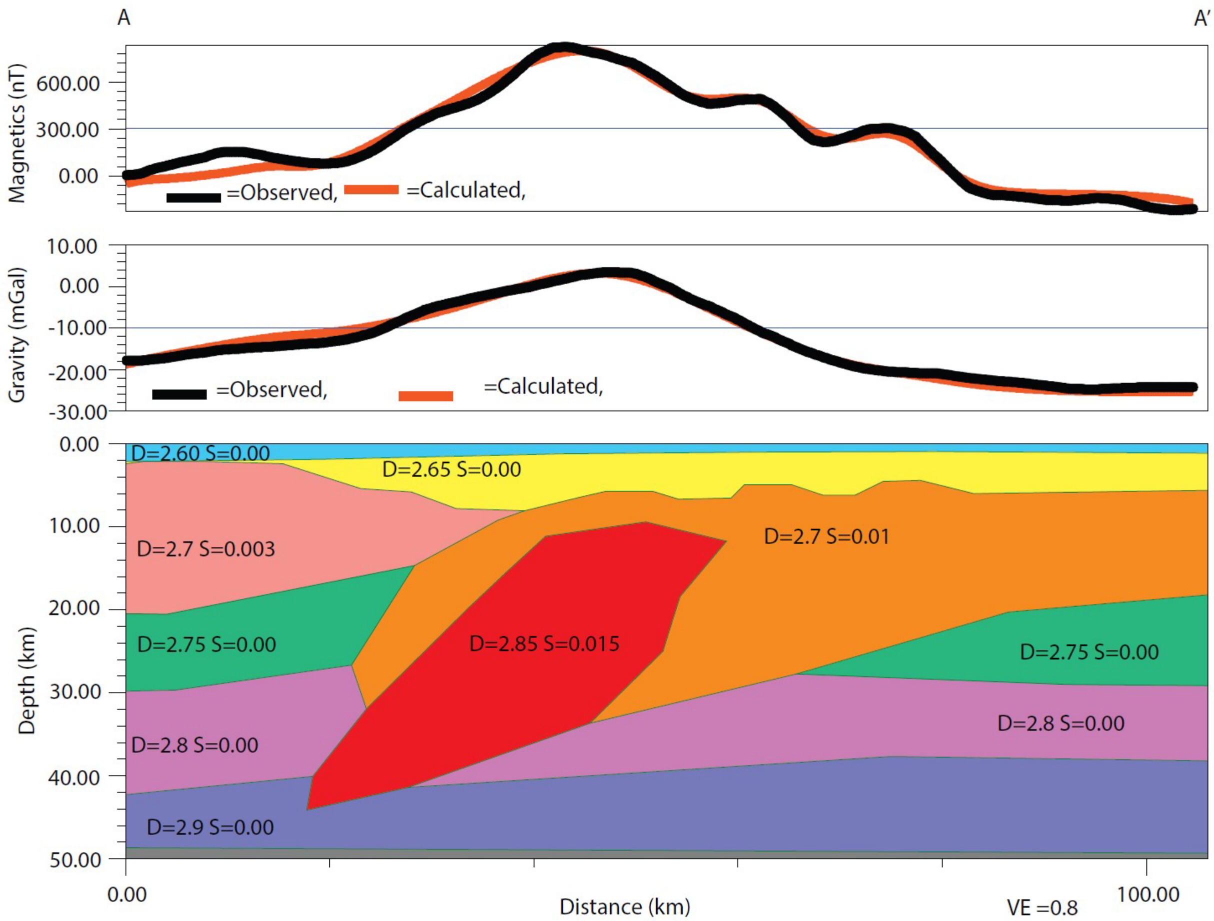


Figure 22.—Model 2 of profile A-A' (Fig. 14) from 2D forward modeling using gravity and magnetic data with Geosoft Oasis Montaj GM-SYS program.

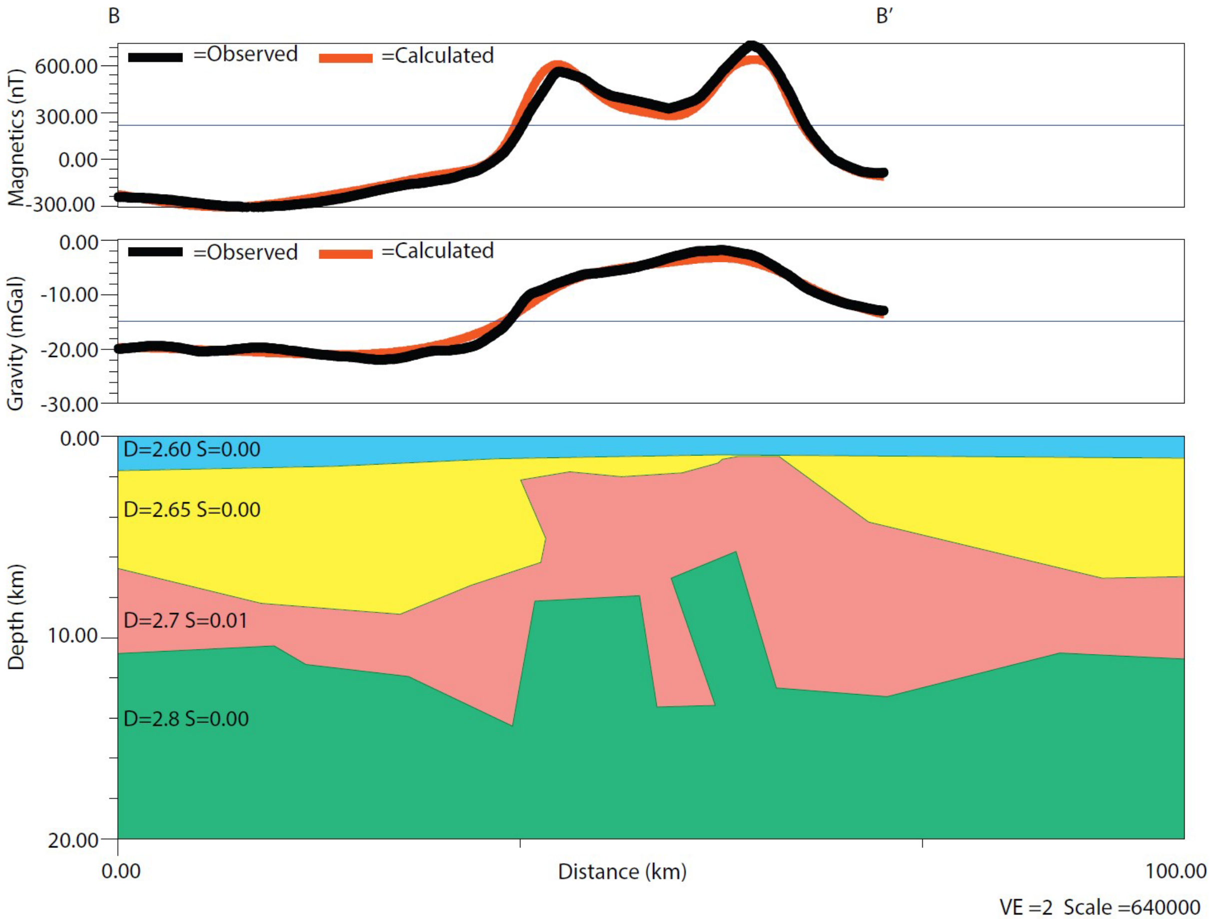


Figure 23. Model 1 of profile B-B' (Fig. 14) from 2D forward modeling using gravity and magnetic data with Geosoft Oasis Montaj GM-SYS program.

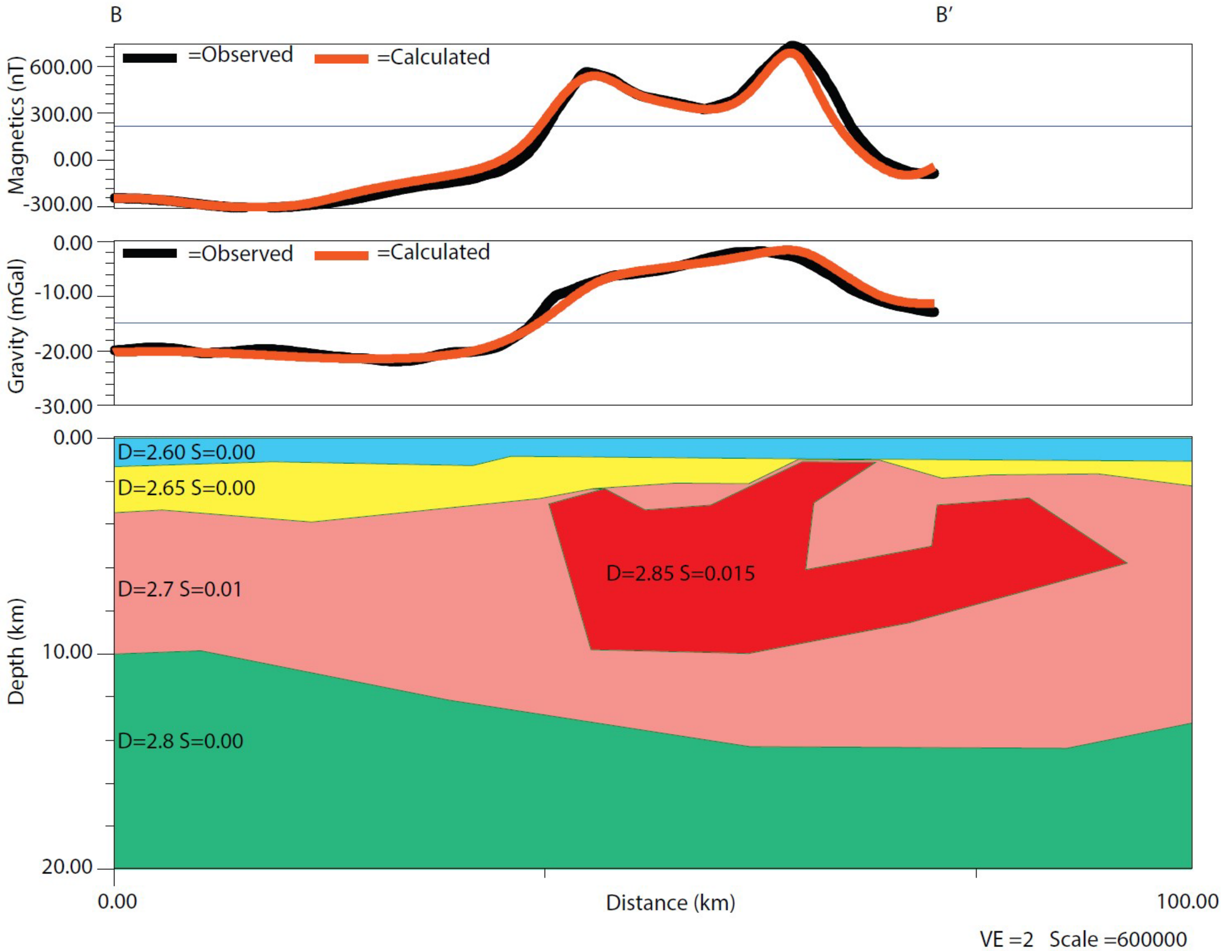


Figure 24. Model 2 of profile B-B' (Fig. 14) from 2D forward modeling using gravity and magnetic data with Geosoft Oasis Montaj GM-SYS program.

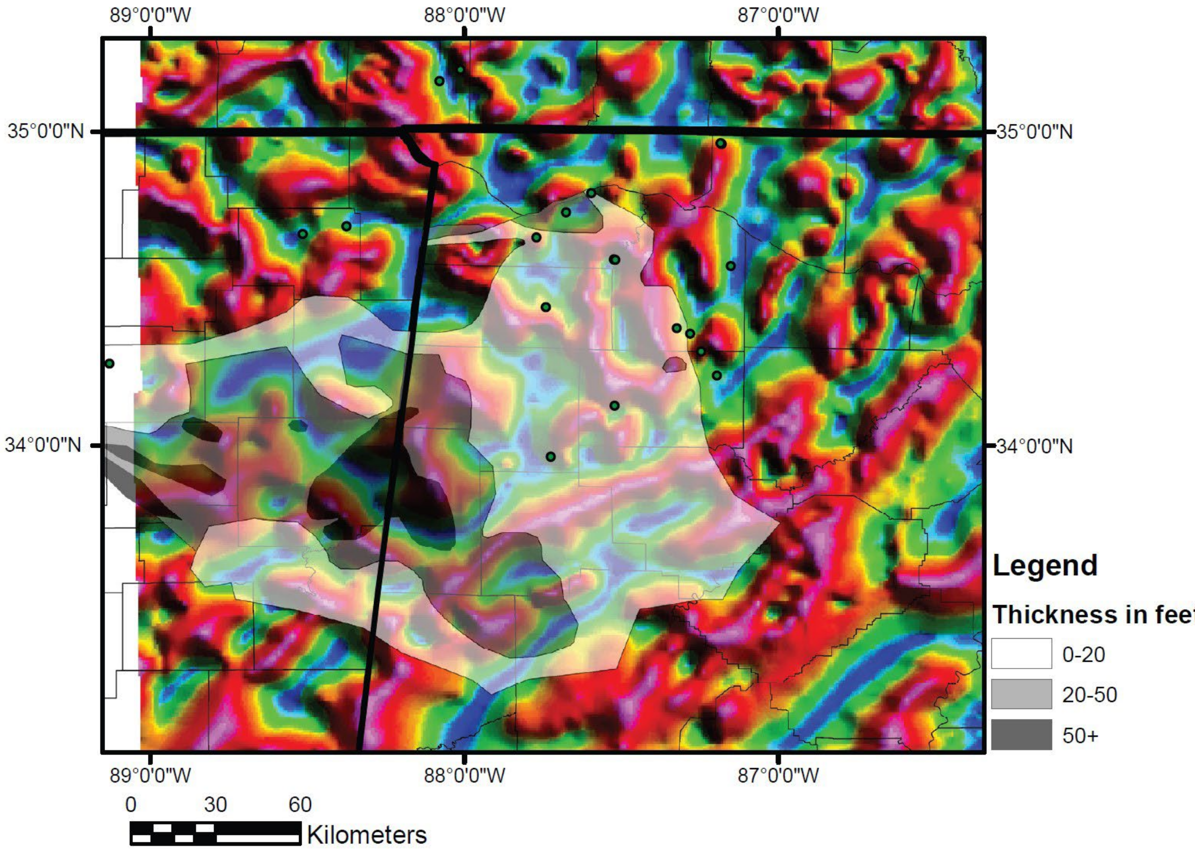


Figure 25. Isopach of the Lewis Sandstone of the Pride Mountain Formation overlain on the vertical tilt derivative anomaly map (isopach extent and thickness modified from Kidd, 2008).

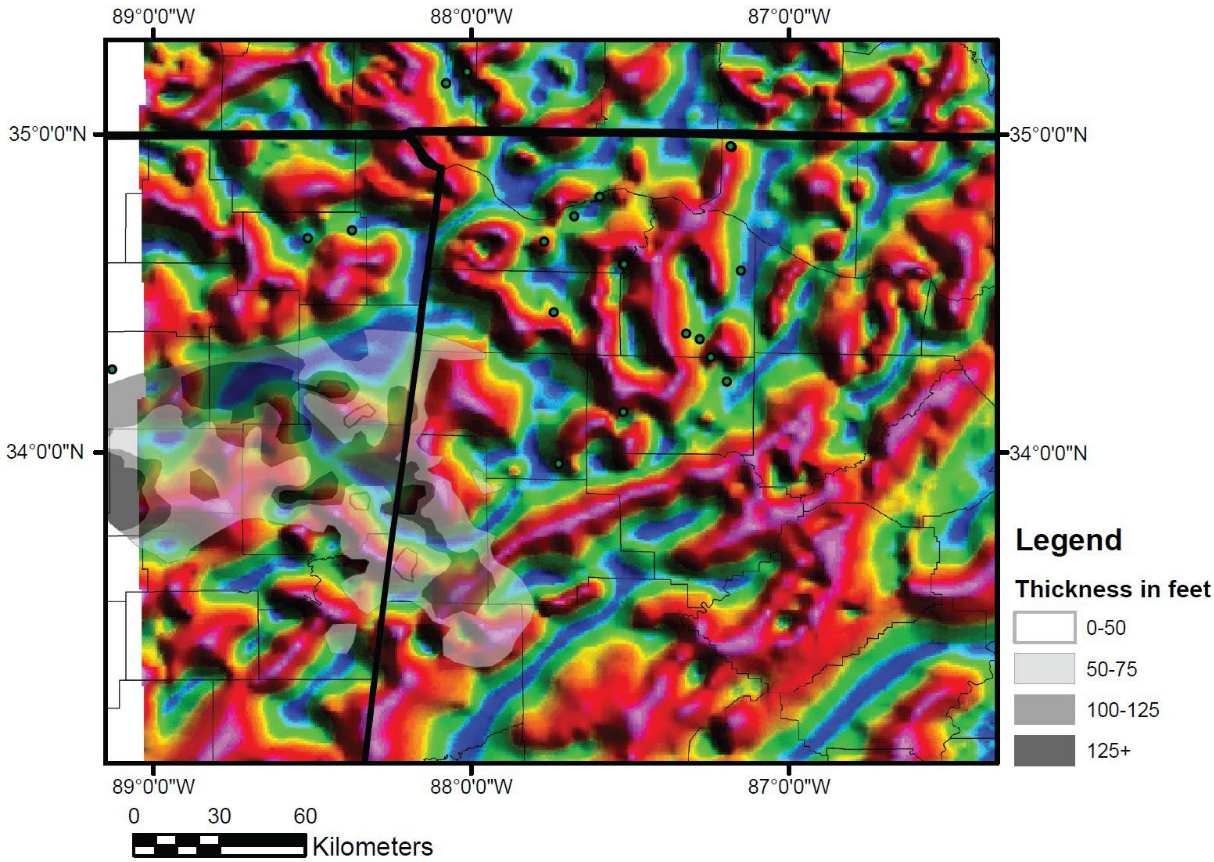


Figure 26. Isopach map of the Lewis cycle overlain on the vertical tilt derivative anomaly map (isopach extent and thickness modified from Canton, 2011).

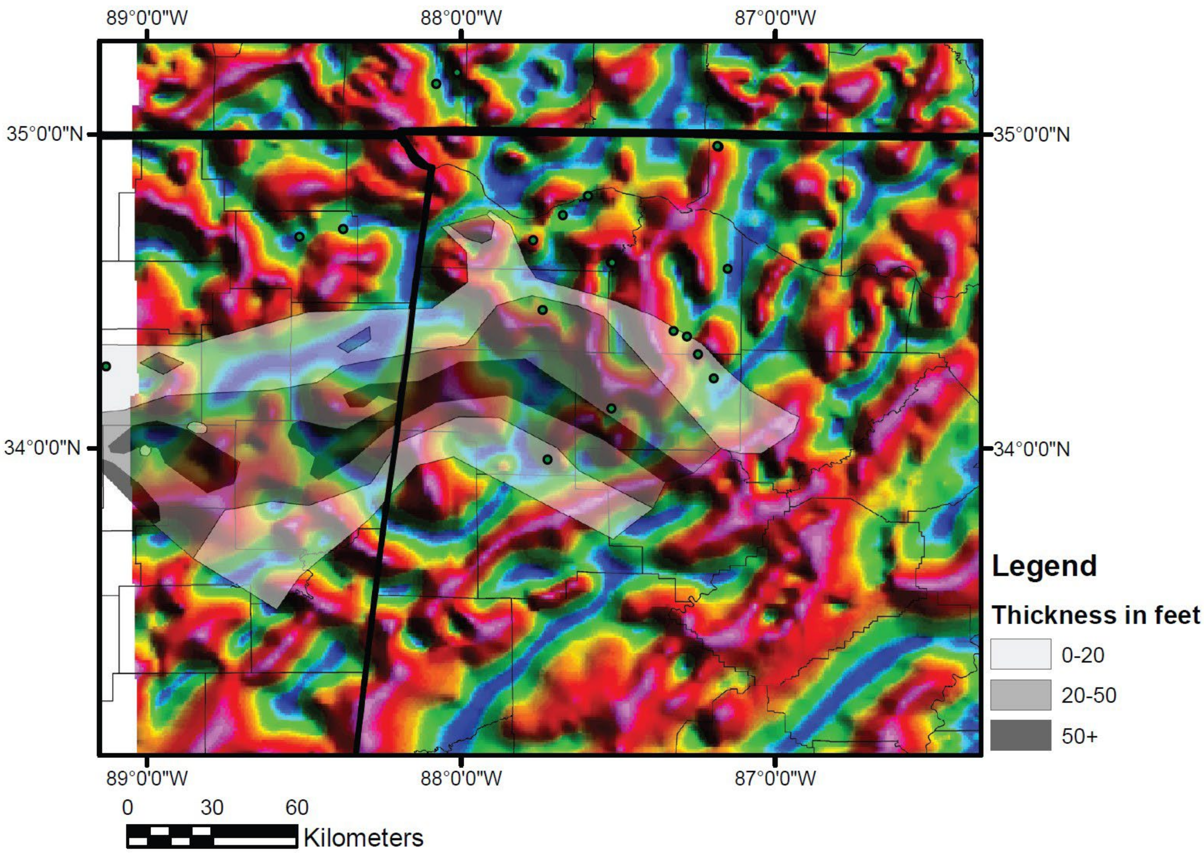


Figure 27. Isopach map of the Evans Sandstone overlain on the vertical tilt derivative anomaly map (isopach extent and thickness modified from Kidd, 2008).

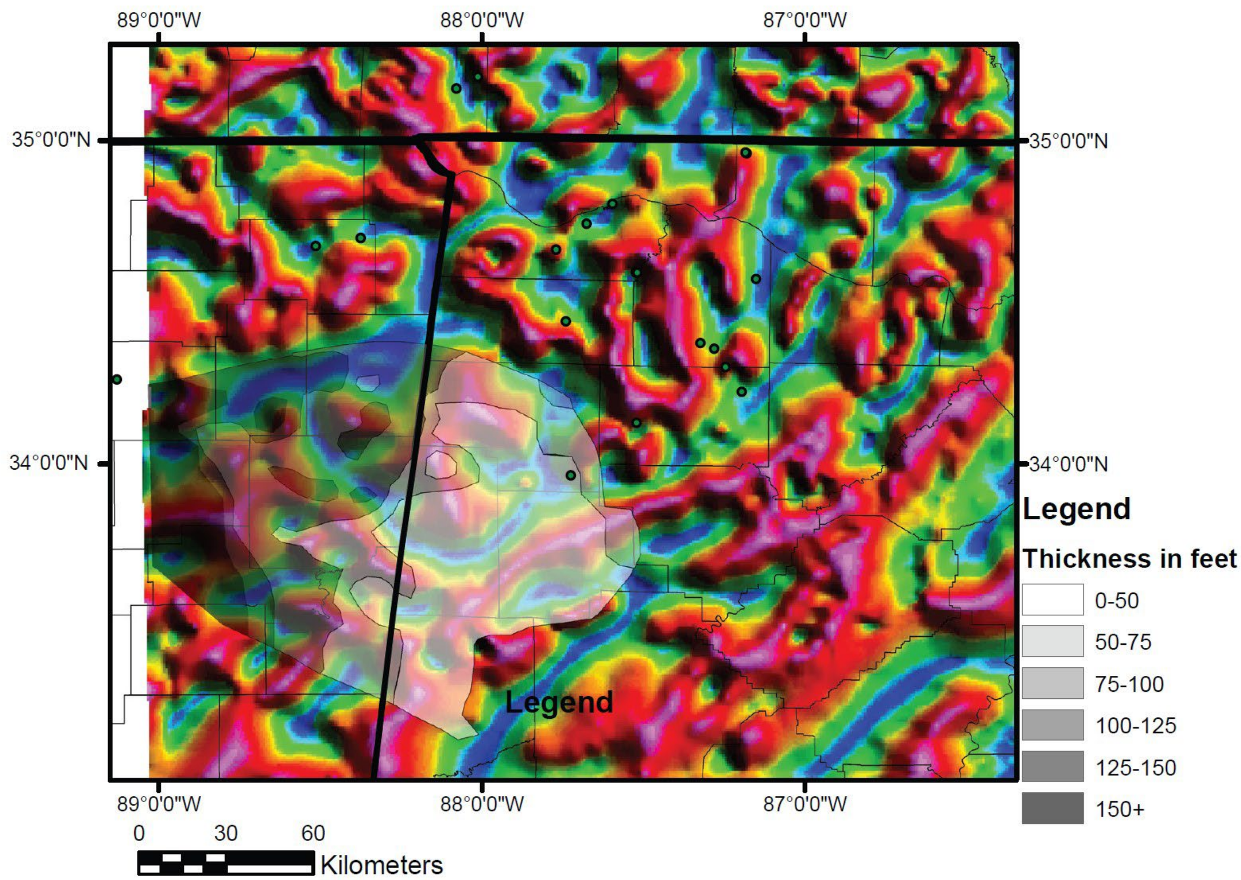


Figure 28. Isopach map of the Evans cycle overlain on the vertical tilt derivative anomaly map (isopach extent and thickness modified from Canton, 2011).

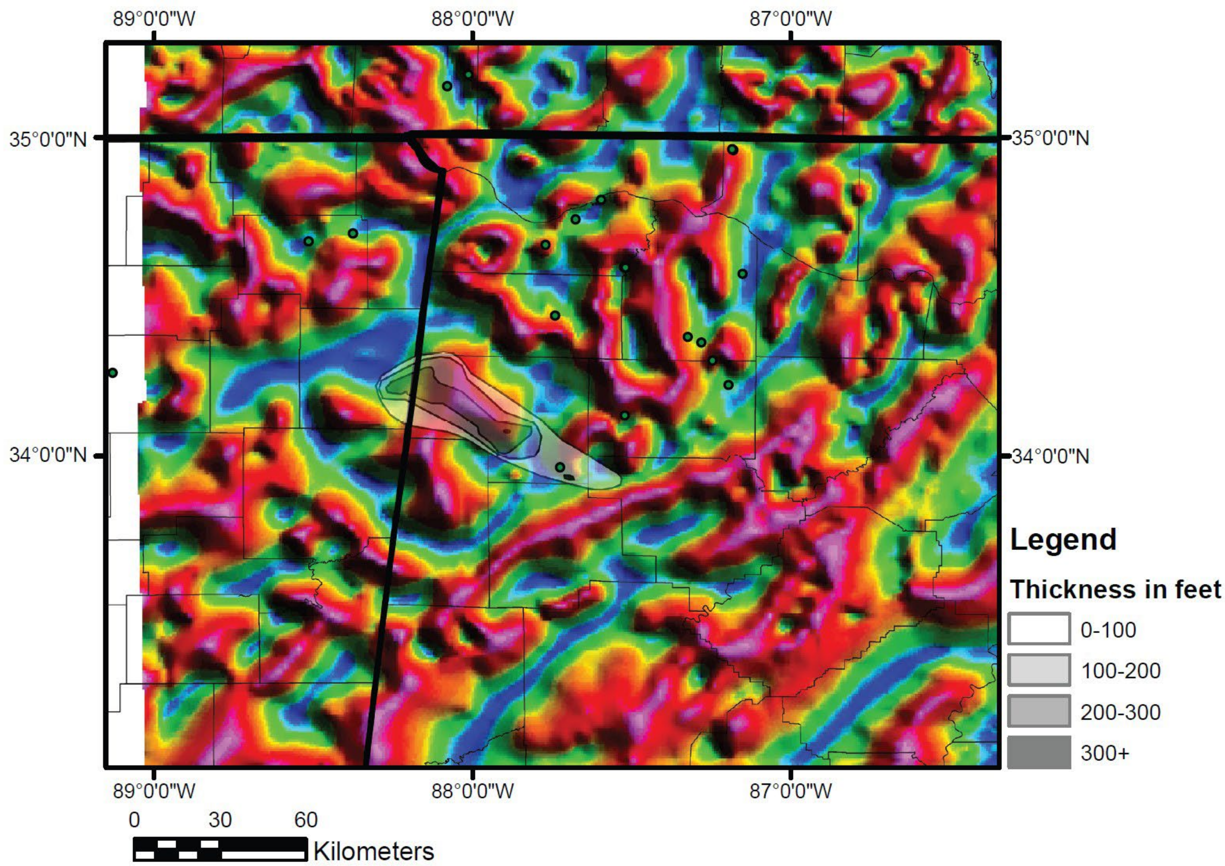


Figure 29. Isopach map of the Pearce Siltstone overlain on the vertical tilt derivative anomaly map (isopach extent and thickness modified from Kidd, 2008).

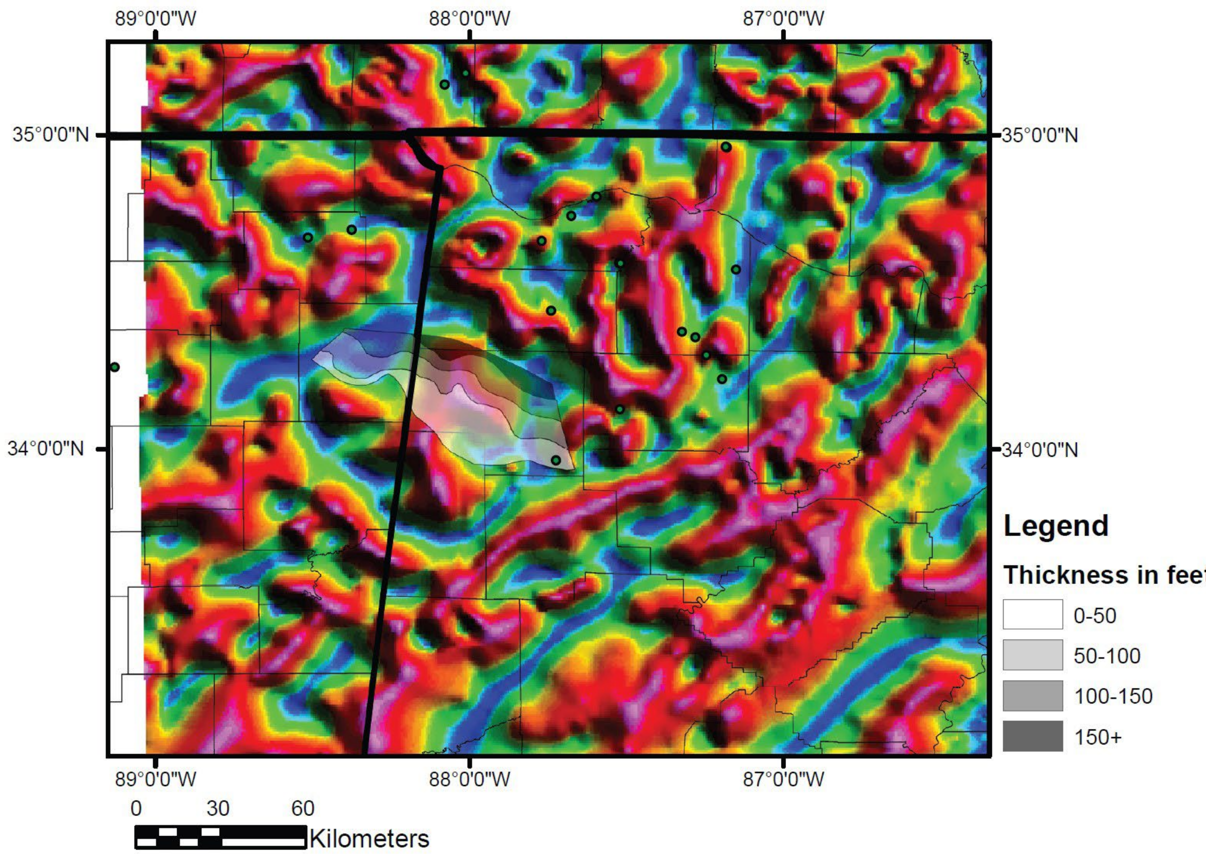
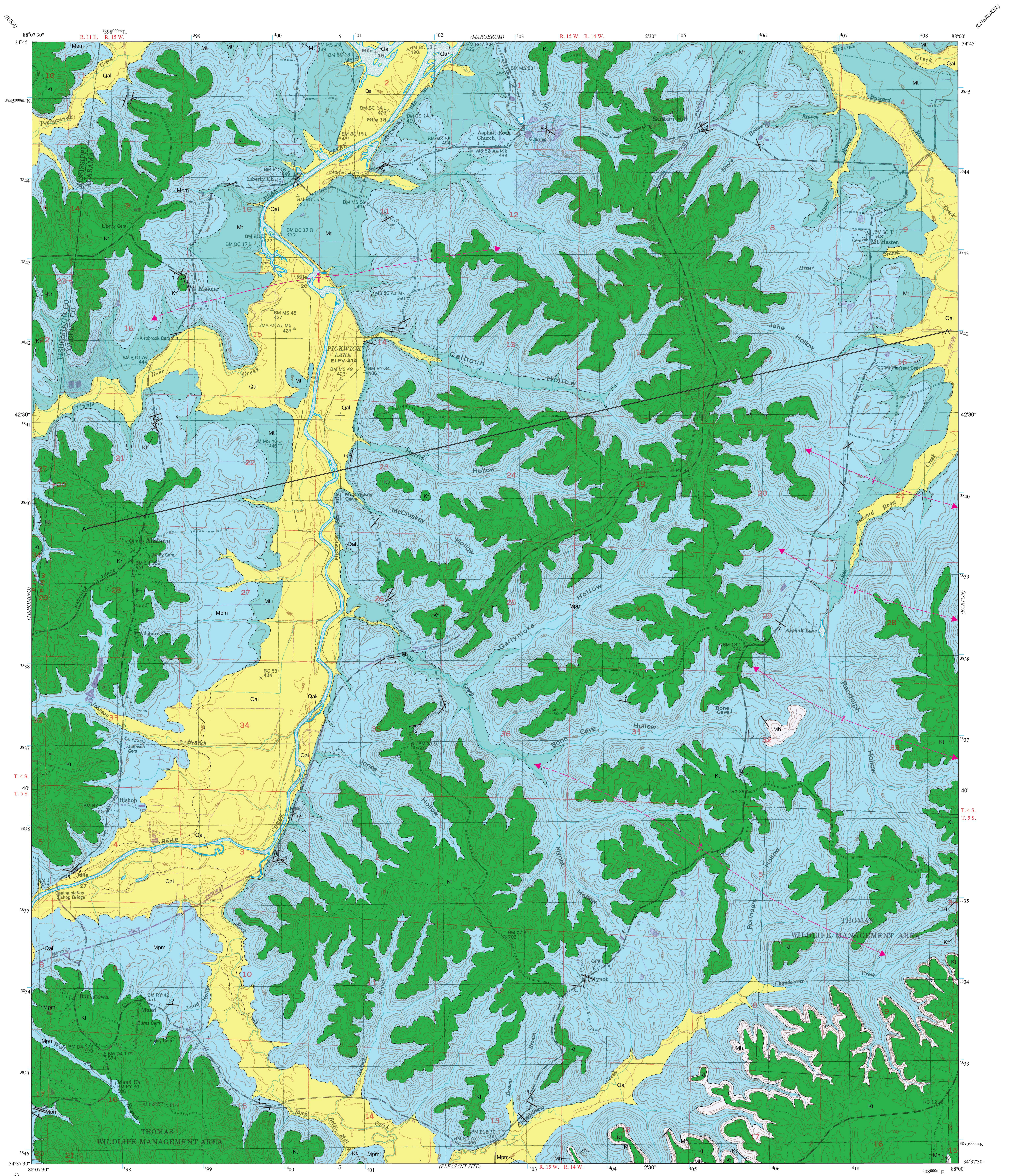


Figure 30. Isopach map of the Hartselle Sandstone overlain on the vertical tilt derivative anomaly map (isopach extent and thickness modified from Canton, 2011).

APPENDIX



CORRELATION OF MAP UNITS

Qal	Quaternary	CENOZOIC
unconformity		
Kt	Cretaceous	MESOZOIC
unconformity		
Mh, Mpm, Mt	Mississippian	PALEOZOIC

DESCRIPTION OF MAP UNITS

Qal **Alluvium and low terrace deposits (Quaternary)**—Unconsolidated deposits of sand, silt, clay, and gravel.

Kt **Tuscaloosa Group (Cretaceous)**—Unlithified clay, sand, and gravel; orange clay and sand is locally micaceous; gravel is matrix- to clast-supported and contains white chert clasts; sand grans are mostly quartz, and sandy intervals contain significant clay.

Mh **Hartselle Sandstone (Upper Mississippian)**—Thick-bedded to massive, light-colored, fine-grained, well sorted quartoze sandstone that commonly weathers grayish; grain size can range from coarse to silt with locally calcarous sands. Horizontal laminae, ripples and cross-beds are common. Fossils and plant fragments are locally uncomon but known to exist regionally.

Mpm **Pride Mountain Formation (Upper Mississippian)**—Medium- to dark-gray shale and mudstone, interbedded with distinct light- to medium-, gray- to brown sandstone and medium-gray limestone; base of unit defined by a gray, calcareous shale; up to two brown and gray sandstone intervals present in middle and upper part of formation; a limestone is present in the middle of the unit.

Mt **Tuscumbia Limestone (Lower and Middle Mississippian)**—Light- to dark-gray, micritic bioclastic, sparry limestone locally containing chert; lower portion of the Tuscumbia Limestone contains medium- to dark-gray, medium- to thick-bedded bioclastic limestone; upper portion is dominated by light- to medium-gray sparry micritic limestone, with bedding thickness varying from thin (<6") to thick (>10").

SYMBOLS FOR GEOLOGIC MAP

Contact, located very approximately

Contact, concealed beneath mapped units

Trace of anticline axis, located approximately, arrow showing direction of plunge

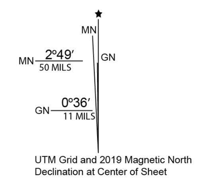
Trace of syncline axis, located approximately, arrow showing direction of plunge

15 Strike and dip of bedding

Base topographic map U.S. Geological Survey 1950. Photorevised 1984.
Produced in cooperation with the U.S. National Park Service

Polyconic projection. 1927 North American datum
10,000-foot grid based on Alabama (West)
rectangular coordinate system

Map rotated 0.6 degrees counterclockwise for display



UTM Grid and 2019 Magnetic North Declination at Center of Sheet

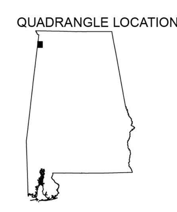
SCALE 1:24,000

1 — 1/2 — 0 — 1 MILE

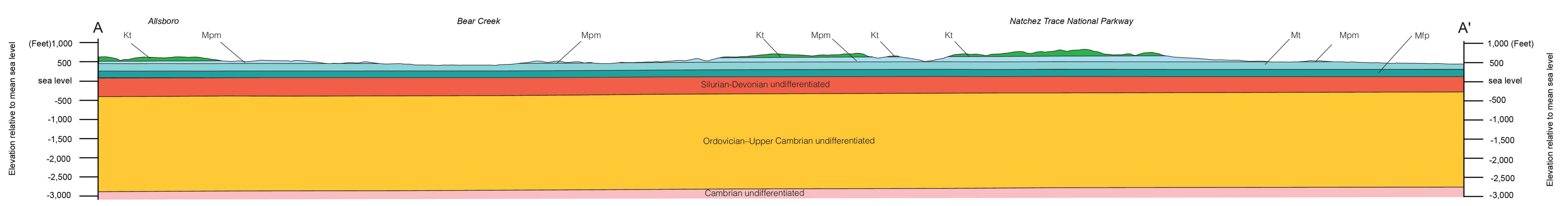
1 — 0 — 1 KILOMETER

CONTOUR INTERVAL 20 FEET
NATIONAL GEODETIC VERTICAL DATUM OF 1929

QUADRANGLE LOCATION



Digital database by M. P. McKay



SYMBOLS FOR CROSS SECTIONS A-A'

Stratigraphic contact

GEOLOGIC MAP AND CROSS SECTION OF THE BISHOP 7.5-MINUTE QUADRANGLE, COLBERT COUNTY, ALABAMA AND TISHOMINGO COUNTY, MISSISSIPPI

by
Joseph Lane, Matthew P. McKay, and Madeline Konopinski
2023