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Drew D. Burford Jr.

University of Kentucky, drewburfordjr@gmail.com

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Drew D. Burford Jr., Student

Dr. Edward W. Woolery, Major Professor

Dr. Edward W. Woolery, Director of Graduate Studies

INTEGRATED SEISMIC-REFLECTION AND MICROGRAVITY IMAGING
ACROSS THE SOUTHERN BOUNDARY OF THE CHARLESTON UPLIFT,
NEW MADRID SEISMIC ZONE, USA

THESIS

A thesis submitted in partial fulfillment of the
requirements for the degree of Master of Science in the
College of Arts and Sciences
at the University of Kentucky

By

Drew David Burford Jr.

Lexington, Kentucky

Director: Dr. Edward W. Woolery, Professor of Geophysics

Lexington, Kentucky

2019

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ABSTRACT OF THESIS

INTEGRATED SEISMIC-REFLECTION AND MICROGRAVITY IMAGING ACROSS THE SOUTHERN BOUNDARY OF THE CHARLESTON UPLIFT, NEW MADRID SEISMIC ZONE, USA

The Charleston Uplift (CU), a 30-km-long by 7-km-wide, N46°E-oriented subsurface geologic anomaly in the northern Mississippi embayment near Charleston, Missouri, exhibits up to 36 m of vertical relief across the Paleogene/Quaternary unconformity. Subsurface structural relief, along with the CU's coincident boundary alignment with contemporary microseismicity and the New Madrid North Fault (NMNF), suggest a structural origin. Subsequent seismic soundings indicate vertical structural relief is present in Cretaceous and Paleozoic horizons, supporting the fault-controlled origin. The southern boundary (CU-s) had not been investigated, nor had any direct fault images been acquired. Integrated microgravity and seismic-reflection methods across the inferred CU-s establish the first image of this fault.

Forward modeling indicated that the vertical variation of strata across the CU-s would induce a microgravity anomaly of 1.6 mGal. The observed microgravity anomaly survey across the southern boundary is $1.616 \pm .004$ mGal, and is consistent with the tectonic interpretation. A subsequently acquired seismic-reflection profile corroborates this interpretation. The imaged fault shows approximately 60, 35, and 35 meters of vertical down-to-the-south throw across the tops of Paleozoic, Cretaceous, and Tertiary horizons, respectively. This confirms the CU is not an erosional feature, but a structurally controlled extension of the NMNF.

KEYWORDS: Charleston Uplift, New Madrid Seismic Zone, New Madrid North Fault, seismic reflection, gravity, geophysics

Drew David Burford Jr.

7/25/19

Date

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By
Drew David Burford Jr.

Dr. Edward W. Woolery

Director of Thesis

Dr. Edward W. Woolery

Director of Graduate Studies

7/25/19

Date

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CHAPTER 1. INTRODUCTION

1.1 PROBLEM AND OBJECTIVES

The objective of this study was to locate and image the southern boundary of the subsurface Charleston Uplift (CU-s) and determine if its origin was tectonic or erosional in nature. To do so, integration of two different geophysical techniques was used: high-resolution microgravity and P-wave seismic reflection common-midpoint (CMP) imaging. Two microgravity anomaly transects serve to constrain the location of the CU-s. A coincident seismic survey provides a higher resolution image of the fault. The seismic image more precisely locates the CU-s and can be interpreted to infer the nature of the uplift's history.

The Charleston Uplift (CU) is a geologic feature with minimal surficial expression located near the New Madrid Seismic Zone (NMSZ) in Mississippi County, Missouri near the city of Charleston (Pryne et al., 2013; Rucker, 2017). It has been hypothesized that this subsurface stratigraphic high formed as a result of transpressional movement along pre-existing faults extending to the north-east from the New Madrid North fault (NMNF) (Pryne et al., 2013). However, an erosional origin has not been ruled out, since boundary faults have never been directly imaged.

1.2 SIGNIFICANCE

Recent midcontinent earthquakes in the central United States are concentrated within the New Madrid seismic zone (NMSZ) (USGS and CERI

earthquake catalogs), which dominates seismic hazard assessment in the region (Johnston and Schweig, 1996). Additionally, the NMSZ has a historic and prehistoric record of clustered large-magnitude earthquakes (Tuttle et al., 2002), leading to major earthquake hazard for much of the central United States (Johnston and Schweig, 1996). The NMSZ is the source of the great 1811-1812 earthquake sequence. On December 16, 1811 the first occurred, followed by earthquakes on Jan. 23 and Feb. 7 of 1812 (Johnston and Schweig, 1996). They estimate their moment magnitudes (**M**) at $8.1 \pm .31$, $7.8 \pm .33$ and $8.0 \pm .33$, respectively. Based on liquefaction data, Tuttle et al. (2002) estimate recurrence intervals for such events, between 200 and 800 years. Evidence for paleoseismic events is based on studies of paleoliquefaction (e.g. sand blows) (Tuttle, 2001; Tuttle and Schweig, 1995), geoarchaeological findings (Saucier, 1990), geophysical surveys of deformed Quaternary strata (e.g. Pratt, 2012; Ali and Woolery, 2014; Rucker, 2017), and geological trenching (Kelson et al., 1996).

The NMSZ earthquake mechanics, such as slip rate, displacement, strain accommodation, and fault locations, are poorly understood. These poorly constrained properties create a high degree of uncertainty and lack of consensus in regional hazard assessment (e.g., Pratt, 1994, 2012; Schweig and Ellis, 1994; Newman et al., 1999; Cox et al., 2000; Van Arsdale, 2000; Tuttle et al., 2002; Calais et al., 2005, 2010; Smalley et al., 2005; Calais and Stein, 2009; Pryne et al., 2013). The sparse surface deformation and poor preservation of tectonic-based geomorphic signatures is thought to be due to the NMSZ young

seismicity, migrating seismicity, broadly distributed strike-slip faulting, and active Mississippi River fluvial dynamics. These poorly defined factors diminish the physical surface manifestations and make seismic-hazard models more difficult to calculate (Pratt, 1994, 2012; Schweig and Ellis, 1994; Newman et al., 1999; Stein and Newman, 2004). The CU is no exception; along its boundaries there are no known surficial fault expressions. On the uplift itself however, Pryne et al. (2013) point out the elevation of the surface is slightly higher than the adjacent area.

Pryne et al. (2013) first recognized the CU using borehole data to investigate the shallow subsurface of Mississippi County. From these data they have constrained the CU to within this Missouri county, between the north-east terminus of the NMSZ and fault complexes further along strike to the north-east. If proven to be tectonic in origin, this feature may be structurally connected to faults within the NMSZ. To hypothesize the existence of NMSZ extensions to the north-east is tenable based on evidence in recent studies (McKay and Bonora, 2001; Baldwin et al., 2002, 2005; Csontos et al., 2008; Shumway, 2008; Pratt, 2012; Pryne et al., 2013; Ali and Woolery, 2014; Rucker, 2017). Physical evidence of such structural linkages is extremely pertinent to the determination of seismic hazard assessment on a regional scale.

1.3 GEOLOGIC BACKGROUND

This study investigates the subsurface in Mississippi County, Missouri, where Pryne et al. (2013) and Rucker (2017) have suggested the southern boundary of the CU is located. This south-eastern Missouri county lies in the

northern end of the Mississippi Embayment (ME). The essentially flat surface at the site (Pryne et al., 2013 and MSDIS LiDAR data) reflects little of the complex history below. This physiographic region is entrenched into the central United States, from southern Illinois to the northern Gulf Coastal Plain (Cushing et al., 1964).

The processes that created the ME began during the Precambrian, when the formation of Rodinia induced uplift in most of what would become the southeastern United States (Ervin and McGinnis, 1975; Van Arsdale, 2014). The subsequent breakup of Rodinia created a passive margin along the southeastern United States, bordered by the Iapetus Ocean (Van Arsdale, 2014; Kolata and Nelson, 1991) (Figure 1). Throughout Cambrian times, multiple aulacogens formed across this margin creating deep regional rift basins throughout North America (Kolata and Nelson, 1991). Locally, the Reelfoot Rift (RR) formed along with the Rough Creek Graben (RCG), the Rome Trough (RT) and the Southern Oklahoma Aulacogen (SOA). By the Late Cambrian regional down-warp enabled the formation of a cratonic embayment, within which deposition occurred through most of the Paleozoic (Kolata and Nelson, 1991; Ervin and McGinnis, 1975).

Figure 1: This figure shows a map of the Cambrian aulacogens (grey) that developed along the lapetan passive margin after the breakup of Rodinia. Specifically, the Reelfoot rift (RR), the structural precursor to the NMSZ, the Rough Creek Graben (RCG), the Southern Oklahoma Aulacogen (SOA), Rome Trough (RT), Ottawa Rift (OR), Saguenay Rift (SR) (from Wheeler, 1997).

The RR, shown in Figure 2, originally formed in response to tensional forces during the early Paleozoic. The RR was uplifted during the formation of Pangea at the close of the Paleozoic. Subsequently, the stress field changed again, returning to a tensional regime but with a north-west to south-east orientation. This occurred during the Mesozoic in response to the breakup of Pangea and caused subsidence of what would become the Gulf Coastal Plain (Kolata and Nelson, 1991; Ervin and McGinnis, 1975).

Figure 2: This figure shows an elevation map of the top of the Precambrian unconformity in the Reelfoot Rift area. The Reelfoot Rift is bounded by the Eastern Rift Margin faults (EM), Western Rift Margin fault (WM), and bisected by the Axial fault (AF). RF, Reelfoot fault; NM, New Madrid, Missouri (from Csontos et al., 2008).

Another cycle of uplift and subsidence occurred when the North American plate moved over the Bermuda hot spot. Hot, low density mantle material intruded below the RR during the mid-Cretaceous. As it cooled and sank, so too did the RR. This final act enabled the formation of the ME and the Mississippi River (Van Arsdale, 2014).

1.4 STRATIGRAPHY OF THE MISSISSIPPI EMBAYMENT

Figure 3: This figure shows a stratigraphic column of the northern Mississippi Embayment (from Pryne et al., 2013).

The ME consists of Late Cretaceous through Quaternary sediments (Cox and Van Arsdale, 2002). A stratigraphic column from Pryne et al. (2013) is shown in Figure 3. The deepest stratigraphic layer encountered in this study was the Paleozoic. This layer consists of limestone, dolomite, and sandstone (Ross, 1963). Cretaceous strata unconformably overlies the Paleozoic strata. These sediments from deepest to shallow are: the Coffee Formation, consisting of well-sorted sand interbedded with carbonaceous clays; the Demopolis Formation, containing thick marl beds; the McNary Sand, which contains fine- to coarse-grained sand and is interbedded with silty clays; lastly, the thin Owl creek formation, consisting of thick glauconitic, micaceous clay (Crone, 1981). Cretaceous strata are especially useful in seismic investigation due to the strong, coherent reflections generated at its boundaries (Rucker, 2017).

The base of Tertiary sediment is defined by the unconformable interface on which the Midway Group is laid. The Midway Group begins with the glauconitic, sandy clays of the Clayton Formation at the bottom (Ross, 1963; McBride et al., 2003), followed by the Porters Creek Clay. Unconformably above that is the Old Breastworks Formation (Crone, 1981). The Wilcox Group lies conformably above and its extent is defined by the base of the Fort Pillow Sand and ends at the top of the Flour Island Formation. These contain fine- to very coarse-grained quartzite sand and silty clays, respectively (Crone, 1981).

The Claiborne Group unconformably overlies the Wilcox Group, contains sand, silt, and clay and marks the beginning of Eocene strata (Crone, 1981). The last layer of Eocene strata is the Jackson Formation. The Jackson Formation

has medium- to very-fine grained silty sand and is interbedded with clayey silt (Crone, 1981). Woolery et al. (2003) report that since these Eocene strata are of similar lithologies, the boundaries between individual formations are not well observed in seismic exploration.

Van Arsdale et al. (2007) indicate that the subsequent Upland Complex was deposited by the ancestral Mississippi and Ohio rivers during the Pliocene. Followed by loess which can be comprised of three different ages depending on location within the ME (Kolata et al., 1981; Ross, 1963). The following alluvial sediments, namely the Charleston Fan, were deposited by ancestral Mississippi river floodwaters during deglaciation (Porter and Guccione, 1994).

1.5 SEISMOTECTONIC SETTING

Shown in Figure 4, the midcontinent stress field is primarily compressive; based on in-situ measurements, the principal stress direction is east-west (Cox et al., 2000). Regional compression is induced by westward movement of the North American plate away from the mid-Atlantic spreading center (Kolata and Nelson, 1991; Cox et al., 2000; Zoback and Zoback, 1980; Nelson and Bauer, 1987). The NMSZ and its faults can be generalized as two northeast-oriented dextral strike-slip fault zone segments offset by a central northwest-oriented left-stepping restraining-bend thrust (Odum et al., 1998; McClay and Bonora, 2001; and Pratt 2012). These fault planes are reactivated in response to the compression, inducing the observed seismicity. The northeast-trending arms: the New Madrid North Fault (NMNF) and the Axial Fault (AF), exhibit right-lateral slip at depths of 3 to 15 kilometers along a near-vertical fault surface. The step-over

arm: the Reelfoot Fault (RF), exhibits thrust fault displacement along southwest trending faults to similar depths (Csontos and Van Arsdale, 2008; Csontos et al. 2008).

Figure 4: This figure is a map of the Mississippi embayment showing NMSZ seismicity from (1979 to 2006) color coded by depth to hypocenter. The hypothesized boundaries of the CU are shown with dashed dark blue lines. Primary faults of the NMSZ are shown with dashed white lines. Solid white arrows approximate the orientation of the regional compressive stress on the fault network (Kolata and Nelson, 1991; Cox et al., 2000; Zoback and Zoback, 1980; Nelson and Bauer, 1987). The locations of 1811-1812 earthquakes are shown with yellow stars (Johnston and Schweig, 1996) (modified from Pryne et al., 2013).

McClay and Bonora (2001) test a generalized fault model similar to the NMSZ. In kinematic sandbox models, strike-slip movement was induced along two parallel, offset plates at the base of the sandbox. The results of these experiments revealed the faulting patterns shown in Figure 5. Pratt (2012) correlated their results with seismicity and known faulting in the NMSZ (Figure 6).

Figure 5: This figure shows the McClay and Bonora (2001) map of faults developed in a kinematic sandbox model where strike-slip movement was induced along two parallel, offset plates at the base of the sandbox (modified from Pratt, 2012).

Figure 6: In this figure Pratt (2012) has projected the faulting patterns seen in the McClay and Bonora (2001) kinematic sandbox model onto a map of established faults and contemporary microseismicity within the NMSZ. The Charleston Uplift is shown with a red star. The line labeled C shows the location of seismic profile from Pratt (2012). Memphis is labeled as M (from Pratt, 2012).

Pratt (2012) shows that the modeled faults correlate with the observed contemporary microseismicity and established faults, except for those in the northern end of the NMSZ. While the northern-most strike-slip faults in the model approach the Fluorspar Area fault complex in Kentucky and Illinois, known NMSZ faults terminate in south-eastern Missouri. To illustrate, Figure 6 has been overlain with a map of regional faults in Kentucky, Illinois, and Indiana in Figure

7. Pratt (2012) uses Figure 6 to predict that NMSZ tectonic processes should extend further to the north-east. The CU may prove to be one of these features.

Figure 7: This map of regional fault complexes has been overlain onto Figure 6. The location of the CU is shown with a red star (modified from Woolery, 2002 and Pratt, 2012).

1.6 PRIOR INVESTIGATIONS OF THE CHARLESTON UPLIFT AREA

While, seismic network analysis can reasonably constrain the primary faults in the generalized model, precisely locating and accurately determining characteristics like displacement, dip, strike and hazard potential of these faults

and other minor faults within the NMSZ requires implementation of higher-resolution geophysical techniques.

The NMNF is defined by the north-east trend of seismicity that originates at its conflux with the New Madrid West Fault and the Reelfoot Fault (Chiu et al., 1992; Csontos et al., 2008; Csontos and Van Arsdale, 2008). The NMNF experiences oblique compressional stress due to its north-east trending fault plane. This results in reverse dextral movement along the NMNF (Shumway, 2008). Baldwin et al. (2002, 2005) interpret a seismic profile that transects the NMNF at Farrenburg, MO. Their interpretation locates multiple near-surface faults including the Farrenburg Lineaments, which define the north-eastern terminus of the NMNF. A seismic profile is shown in Figures 8a and 8b and the profile location is shown in Figure 9.

Figure 8: This figure shows the Dow Chemical Vibroseis line M-8 oriented north–south along Sikeston Ridge A–H are faults interpreted by Baldwin et al. (2005). 8a) is the uninterpreted seismic line; triangular sections marked “nd” are areas of incomplete data recovery because of road overpasses intersecting the line. 8b) Is the interpreted seismic line. K, Cretaceous; P, Paleozoic horizons. See Figure 9 for profile location (from Pryne et al., 2013).

Figure 9: In this figure the NMNF is shown with an orange dashed line (Chiu et al., 1992; Csontos et al., 2008; Csontos and Van Arsdale, 2008). The Farrenburg Lineaments at the terminus of NMNF are represented by the black lines near Farrenburg (F), as interpreted by Baldwin et al. (2002, 2005). The fault near Olmstead (O) is also shown with a black line (Bexfield et al., 2005). Asterisks are the locations of the 31 October 1895 Charleston and the 2 February 2012 earthquake epicenters. The focal mechanism for the 2012 earthquake is shown with beach ball diagram: N40°E, 59° SE-trending fault plane with reverse dextral displacement. Black dots show earthquake epicenters from the Center for Earthquake Research and Information catalog (1979–2006). B and H, locations of faults B and H in Figure 8b. BA, Ballard, Kentucky; CA, Cairo, Illinois; C, Charleston, Missouri; CO, Commerce, Missouri; S, Sikeston, Missouri; HI, Hickman, Kentucky; N, New Madrid, Missouri (modified from Pryne et al., 2013).

To the north-east of the Farrenburg Lineaments, near Charleston

Missouri, in Mississippi County, the Center for Earthquake Research and

Information (CERI) and the United States Geological Survey (USGS) have recorded seismic activity (Figures 4, 9 and 37). Shumway (2008) has identified a group of these earthquakes within Mississippi County that lie along a north-east trend that she suggests have similar fault plane solutions, all striking to the north-east, with right lateral displacement.

Furthermore, a **M** 6.6 earthquake, the largest known earthquake in the NMSZ since the 1811-1812 series, occurred in the area in 1895 (Stover & Coffman 1993, Hooper and Algermissen, 1980) (Figure 9). They report that the highest isoseismal values associated with this earthquake followed a north-east trend all the way through southern Illinois into Indiana. Conversely, Bakun et al. (2003) interpret the same data from Hooper and Algermissen (1980) to conclude that the epicenter for this event was located ~100 km north of Charleston, in Illinois. Regardless, both interpretations show that this event did not occur on a NMSZ primary fault and neither location negates the possibility of this event occurring on an extension of the NMNF. More recently, a **M** 3.9 earthquake occurred in 2012 at the location shown in Figure 9. While lesser in magnitude than the 1895 earthquake, the nature and location of this event was far better understood since seismic networks were in place to record the seismic waves it produced. Pryne et al. (2013) describe its focal mechanism using a fault plane strike of N40°E, 59°SE with reverse dextral faulting.

Pryne et al. (2013) have proposed the CU as a structurally controlled seismogenic source. Other than the Reelfoot scarp (Van Arsdale et al., 1995), Crowleys Ridge (Van Arsdale et al., 1995), Farrenburg Lineaments, and possibly

Sikeston Ridge, most seismogenic structures in the NMSZ have no surface manifestation. Thus the lack of a surface fault expression along the boundaries of the CU would not be unusual. Pryne et al. (2013) use a 3D lithological model (Figure 10) made from 517 well logs in the area to define its overall structure and location (well locations shown in Figure 11). They indicate that a N46°E-striking subsurface high, extends through Mississippi County to as far as Cairo, Illinois. While this model serves to roughly constrain the near-surface location and size of the CU, its origin as a structural or erosional feature remained equivocal. They propose that the uplift could be an erosional high (paleodivide), a structurally folded ridge, but favored a fault-bounded structural uplift interpretation.

Figure 10: This figure shows a stack of 3 m interval lithology maps containing 10 intervals from 12 m to 42 m elevation w.r.t modern sea level. The black dashed lines show the hypothesized boundaries of the Charleston Uplift. View is oblique to the northwest with a vertical exaggeration of 75x. CA, Cairo; C, Charleston; F, Farrenburg; N, New Madrid (from Pryne et al. 2013).

Figure 11: In this figure the black dots show the locations of 300 ft deep boreholes across southeastern Missouri and neighboring portions of Kentucky and Illinois. MP-35 and MP-80 are locations where seismic reflection P-wave sounding data were obtained by Pryne et al. (2013). The yellow dots are locations where seismic reflection P-wave sounding data were obtained by Rucker (2017). N, New Madrid; F, Farrenburg; C, Charleston; CA, Cairo; CO, Commerce; HI, Hickman; O, Olmstead; S, Sikeston (from Pryne et al., 2013).

Subsequently, Rucker (2017) conducted eleven P-wave and six SH-wave seismic walkaways soundings in a series of transects across the Pryne et al. (2013) borehole-derived uplift boundaries. Seismic-reflection techniques are not sensitive to detection of strike-slip movement along the fault planes; however, they are useful for observing vertical depth (or two-way-travel-time) variation associated with any vertical component of transpressional fault displacement. The seismic soundings identified vertical offset of Tertiary, Cretaceous and

Paleozoic horizons across the northern boundary of the CU. Rucker's (2017) North Wyatt (NWY A through E transect) P-wave soundings were conducted with 48 40 Hz geophones at 2 m spacing. Between the NWYC and NWYB soundings (Figures 12a and 12b, respectively), Rucker (2017) reports variation in P-wave arrival times for corresponding reflectors (Table 1).

These two soundings serve to constrain the location of the northern boundary of the CU and support Pryne et al.'s (2013) interpretation that the CU is tectonically controlled. If erosional, the deeper horizons would not exhibit offset.

Figure 12: This figure shows the seismic soundings in time and the reflection horizons therein, as interpreted in Rucker (2017). These soundings show vertical offset between Tertiary (Te), Cretaceous (K), and Paleozoic (Pz) horizons across the northern boundary of the CU (sounding locations shown in Figure 11 and 19). (Figure modified from Rucker, 2017).

Table 1: This table lists the zero-offset times (t_0), stacking velocities, thicknesses, and depths of the Tertiary (Te), Cretaceous (K), and the Paleozoic (Pz) horizons observed in Rucker (2017) within each seismic sounding in the North Wyatt transect (modified from Rucker, 2017).

	t_0 (ms)	Stacking Velocity (m/s)	Thickness (m)	Depth (m)	Unit
NWYA	160	1425	114	114	Te
	220	1525	53	167	K
	275	1625	54	221	Pz
NWYB	160	1425	114	114	Te
	220	1525	53	167	K
	280	1625	58	225	Pz
NWYC	200	1425	143	143	Te
	280	1525	70	213	K
	355	1625	73	286	Pz
NWYD	185	1425	132	132	Te
	250	1525	58	190	K
	325	1625	72	262	Pz
NWYE	155	1425	110	110	Te
	205	1525	45	155	K
	280	1625	70	226	Pz

Tuttle et al. (2002) locate numerous liquefaction features near Charleston (Figure 13). During seismic data collection, the owner of farmland immediately adjacent to our first microgravity profile informed us that his fields were spotted with sand patches that may be liquefaction deposits. Using Google Earth, a potential sand blow was identified less than 4.5 km to the south-west of this profile, shown in Figure 14 (location of feature shown in Figure 19b).

Figure 13: On the left of this figure is a map of sand blows in the NMSZ. The section of the map near Charleston Missouri is enlarged on the right. Sand blows near Charleston are in close proximity to the hypothesized fault (modified from Tuttle et al., 2002).

Figure 14: A potential sand blow deposit less than 5 km to the south-west of the first microgravity profile (from Google Earth).

CHAPTER 2. METHODS

Seismic and gravitational exploration methods were chosen for their individual advantages in subsurface exploration and their ability to work in concert to produce a more complete picture of the subsurface than could be created through just one of these methods.

Microgravity measurement and analysis can cover a greater extent in less time than the seismic methods employed in this study. A series of high precision gravity field measurements at observation stations spaced in close proximity constitutes a microgravity profile. Subsurface conditions can be inferred from these data. Depending on the objectives of the study, microgravity data require processing, or reduction, to isolate the gravitational signature of the intended source. Specific data correction procedures are outlined in the following sections and discussed in depth in Appendix B. Microgravity data were observed across the hypothesized location of the CU-s and reduced to create Complete Bouguer Anomaly (Δg_{CBA}) profiles. The gravitational variation in the resulting anomaly profiles were interpreted to infer a likely fault location. Such analyses are, however, too low in resolution to precisely locate and image the hypothesized fault.

In order to precisely locate the CU-s, higher resolution seismic exploration was necessary. First, a P-wave seismic reflection and refraction walkaway sounding was performed adjacent to the fault. The purpose the seismic sounding was to optimize the array geometry, recording window, and the location of the subsequent CMP P-wave reflection profile. The seismic profile was conducted

coincident with the first microgravity profile (GP1), encompassing the section that microgravity analysis suggested the fault was located. The seismic data were then processed and interpreted to produce an image of the CU-s.

2.1 MICROGRAVITY METHODS

2.1.1 Forward Modeling

Prior to microgravity data collection, an expected Δg_{CBA} was calculated using the Oasis Montaj program. This program requires the digital construction of a subsurface model that represents the expected conditions of the subsurface, which it uses to calculate a Δg_{CBA} profile by evaluating horizontal density variations with the modeled subsurface. An example subsurface model and the calculated anomaly profile is shown in Figure 15. In the case of a transpressional fault, stratigraphic layers are vertically and horizontally displaced. Since this study's microgravity data are linear across the hypothesized fault plane, any horizontal displacement is ignored in model construction. In Figure 15, the vertical displacement of the high density layer in grey creates lateral density contrast across the fault plane. Pryne et al. (2013) and Rucker (2017) suggest CU-s exhibits similar horizontal density variations.

Figure 15: To illustrate the concepts of forward modeling, a simple two layer subsurface model, with a single vertical fault was created and the anomaly was calculated in the Oasis Montaj program.

This type of analysis is referred to as forward modeling. Forward modeling serves this study in multiple purposes. First, the anomaly calculated from the forward model indicates if the anticipated subsurface conditions would induce a Δg_{CBA} of a magnitude sufficient to be observed with the Scintrex Autograv CG-5 Gravimeter. This instrument is very durable, measures gravitational field values in mGal to 7 significant figures, constantly monitors electronic tilt sensors to automatically compensate for errors in gravimeter tilt and yields very consistent data. Specifically referring to the later, the CG-5 user manual claims that “Over

many 10's of field readings, the CG-5 will repeat to within a standard deviation of 0.005mGal" (Geomatrix, 2015).

Secondly, the calculated Δg_{CBA} yielded by the forward model is later compared with the observed Δg_{CBA} profiles and iteratively revised to interpret the subsurface. Variation between observed and calculated anomaly magnitudes indicates that the subsurface model parameters are inaccurate in either depth or density of strata, or both. Figure 16 depicts how varying the density of subsurface features influences a calculated gravitational anomaly. Figures 17a and 17b illustrate how varying the depth of subsurface features influences a gravitational anomaly.

Figure 16: This figure illustrates the effects of varying subsurface density parameters on the calculated microgravity anomaly. The magnitude of the anomaly changes in response to density. However, the anomaly gradient maxima locations remain the same showing that density contrast is simply a scale factor for the same model geometry and depth (from Hinze et al., 2013).

Figure 17: These models and their calculated anomalies illustrate the effect of varying subsurface model depth parameters on the calculated microgravity anomaly. 17a shows how increasing the depth of an anomalous block changes the calculated anomaly. 17b shows how increasing the depth of an anomalous stratigraphic layer changes the calculated anomaly (from Hinze et al., 2013).

Figure 18: These models depict the non-uniqueness problem that can arise in potential field modeling. Since the gravitational response profile is influenced by both density and depth, similar anomalies can be produced from very different subsurface conditions (from Hinze et al., 2013).

When considering vertical offset of multiple strata, each with varying density, the inversion is further complicated. In fact, potential field theory states that an infinite set of subsurface conditions could create the exact same anomaly, this principle is referred to as non-uniqueness (Hinze et al., 2013). Figure 18 shows three subsurface models that vary in both structure and density. Despite the stark differences between the subsurface models, their anomalies are strikingly similar. This inherent ambiguity limits the interpretation of gravity data. However, the high density of microgravity measurements within this study's gravity profiles reduces the ambiguity of subsurface conditions. The short-wavelength variations observed therein could not be produced by deeper features without unrealistically large density contrasts (Hinze et al., 2013).

To calculate the gravitational signature of a given subsurface model, Oasis Montaj interprets the model based on the two dimensional analysis methods described in Talwani et al. (1959) and calculates an anomaly using algorithms derived in Won and Bevis (1987). To create such a model, two types of subsurface parameters must be constrained: density contrasts and source geometry. This study estimates these parameters from data presented in localized investigations of the subsurface (e.g. Rucker, 2017; Bexfield et al., 2006).

2.1.1.1.1 Subsurface Forward Modeling: Model 1

Figure 19: Due to the close proximity of the NWYB and C soundings to this study's area of interest, the seismic horizon depths observed therein were used in microgravity forward modeling. The yellow dots show the show locations of the Rucker (2017) North Wyatt seismic soundings (NWYB and NWYC) w.r.t. where this study investigated the CU (black outlined box). Charleston is indicated with a black star. Red circles, epicenters of seismicity in the northern ME. Dashed orange line, NMNF. Dashed blue lines, Pryne et al. (2013) hypothesized locations of CU bounding faults. Green lines delineate Missouri counties (modified seismicity map generated from the CERI New Madrid Catalog).

From Rucker (2017) seismic investigations of the CU, most pertinent for the purposes of forward modeling were the seismic soundings conducted nearest to this study's area of investigation (Figure 19). The soundings nearest

to our site were the North Wyatt soundings A through E (NWYA through NWYE). Rucker (2017) interprets the horizons observed in each sounding and reports the depth to the top of each impedance boundary. By comparing the seismostratigraphic changes in depth between soundings, she suggests that the northern delineating fault of the CU (CU-n) strikes through a location between the NWYB and NWYC soundings. Table 2 shows the depth to each impedance boundary as interpreted by Rucker (2017) in the NWYB and NWYC soundings. These data enabled the structure of the subsurface model to be determined.

Table 2: This table lists the depth values in meters of stratigraphic boundaries observed in Rucker (2017) NWYB and NWYC soundings. These depths were used to calculate the expected microgravity anomaly across the hypothesized fault. Te: Tertiary, K: Cretaceous, Pz: Paleozoic.

Horizon	Depth to hanging-wall horizon (NWYB) (m)	Depth to foot-wall horizon (NWYC) (m)	Offset (m)
Te	114	143	39
K	167	213	46
Pz	225	286	61

As discussed above, the second type of model parameters required to calculate a Δg_{CBA} are density estimates. Since Rucker (2017) interpreted the impedance boundaries to be the tops of the Tertiary (Te), Cretaceous (K), and Paleozoic (Pz) formations, density estimates of corresponding strata were required. Bexfield et al. (2006) conducted a study in Olmstead, Illinois, less than 40 km away from our investigation site (Figure 11), wherein they report estimated densities of the Te, K, and Pz strata. These values are shown in Table 3 and assigned to their corresponding strata in Model 1.

Table 3: This table lists the density values from Bexfield et al. (2006) for strata that correspond to those observed in Rucker (2017). These density values were necessary to calculate the expected microgravity anomaly across the fault.

Stratigraphic layer:	Density (g/cc):
Quaternary	1.7
Tertiary	2.0
Cretaceous	2.4
Paleozoic	2.7

Model 1, shown in Figure 20, was constructed with the parameters described above. The fault is modeled to be vertical, as suggested in Rucker (2017) and Pryne et al. (2013). Oasis Montaj calculated that a microgravity survey that intersects the modeled fault normal to its strike would record a gravitational anomaly of ~1.6 mGal across the boundary. Such an anomaly was sufficient in magnitude to warrant investigation with an Autograv CG-5 gravimeter.

Figure 20: Model 1 - shows the microgravity anomaly calculated for a profile across a fault dividing vertically displaced strata. The Oasis Montaj program calculated that this subsurface model would produce a microgravity anomaly with a magnitude of approximately 1.6 mGal. Since Bexfield et al. (2006) and Rucker (2017) data were used to create the subsurface model, the calculated anomaly represents the anomaly expected to be observed in GP1.

2.1.2 Data Collection

Figure 21: These two maps show GP1, GP2, the location of the in-situ seismic sounding, SP1 and the hypothesized southern boundary of the CU. Figure 21b (right) encompasses the area outlined in yellow box in Figure 21a (left). Microgravity observation stations are shown with black triangles. The location of the P-wave reflection sounding is shown with a yellow star. The extent of SP1 is illustrated with a solid blue line. Hypothesized CU faults, with dashed blue lines. Epicenters of seismicity (1969 - present) shown with red dots (USGS). These data are plotted on a digital elevation model (w.r.t. modern sea level), represented using a color gradient between 84 meters (black) to 113 meters (white) (1 m resolution LiDAR data). Roads and, in Figure 21b, road names are shown with black lines. Elevation and road data from Missouri Spatial Data Information Service (MSDIS).

Microgravity values were observed and recorded along two profiles: GP1 and GP2, shown in Figure 21a and 21b. These profiles were intended to intersect the CU-s based on the location and strike of the CU suggested by Pryne et al. (2013). During data collection, the location of a station was marked with an orange spray paint dot prior to any measurements. The center of the gravimeter base was placed on this dot. Marking stations was necessary for

reoccupation measurements and to perform the elevation survey for each profile. To measure microgravity data, the gravimeter was placed at the station and leveled along both its x and y-axes using the CG-5 built in electronic levels. Next, the distance between the surface and a consistent point at the base of the instrument was recorded. Since gravity is inversely proportional to the square of the distance between two objects (Newton, 1687), it was necessary to account for the height of the gravimeter w.r.t. to the surface in the height correction calculation.

At each station the gravimeter recorded gravitational field values continuously for one minute. The gravimeter then reported an average of these values and a time stamp of the measurement. Immediately after recording the first value and its time stamp in the field book, and without moving the instrument, gravity was continuously observed again for one minute and the outputs recorded. The reported times of measurement cycles were recorded to correct these data for instrumental drift effects. Since these two measurements were done in rapid succession, the average of these two measurements, both gravity and time, constitute a single station's microgravity measurement and time stamp. After both observation cycles, a Garmin GPSMAP 64 was placed on top of the gravimeter and the coordinates at that station were recorded.

Prior to the first survey, the gravimeter was measured continuously to measure the instrumental drift constant. The meter was also set to remove theoretical tidal attraction of the Sun and the Moon in the survey region. The remaining residual drift was removed using a drift correction determined by

repeated gravity measurements at a base station. The first station measured during data collection was the base station. Base station measurements were not included in either microgravity survey. After measurement at the base station, the gravimeter was moved to the first observation station, within the actual profile. Periodically throughout data collection, the gravimeter was returned to the base station and placed in the exact same location, as closely as possible, and a measurement was conducted adhering to the procedure described above. No more than five observation station measurements were made without returning to the base station. During microgravity data collection the minimum and maximum duration between base station reoccupations was 42 minutes and 1 hour, 15 minutes, respectively. After the measurement of all stations within a microgravity profile, an observation station within the profile was reoccupied and measured. This was necessary to calculate the magnitude of error for each profile. Lastly, the base station was reoccupied and measured again. The data were recorded in the field book and later transcribed into excel to create the unprocessed data tables in Appendix A.

2.1.2.1 GP1

GP1 consisted of 13 microgravity observation stations of variable spacing. A 14th station, the base station, was further to the south-east. The 2.3 km microgravity profile transected the fault hypothesized by Pryne et al. (2013) along Mississippi County Road 325. This road intersects the hypothesized fault at a near exact 90-degree angle. This was necessary because the microgravity

anomaly in the forward model was calculated for a profile that intersected the fault plane normal to its surface. GP1 raw data are shown in Table A1.

2.1.2.2 GP2

About 4 km to the south-west of GP1, a second microgravity survey was conducted: GP2 (Figure 21). GP2 observations were performed along County Road D. The location and extent of GP2 was planned based on the fault location constraint suggested by GP1 results and the N46°E hypothesized fault strike (Pryne et al., 2013). GP2 consisted of 14 microgravity observation stations of variable spacing. A 15th station to the east of the profile was chosen as a base station and reoccupied periodically during data collection. It is important to note that GP2 was conducted at a ~45° angle to the hypothesized fault plane. This was done because there were no other roads close to GP1 that had an ideal azimuth. GP2 raw data are shown in Table A2.

2.1.2.3 Elevation Surveys

An elevation survey for each microgravity profile was necessary for microgravity data corrections. This survey yielded measurements of relative change in surface elevation at each observation station within GP1 and GP2. The elevation survey was done with a BenMeadows Company B2A magnetically damped auto-level elevation survey kit. Data Tables 9 and 10, respectively, show the change in surface elevation at each station w.r.t. the surface at a specific station within each microgravity profile.

2.1.3 Data Reduction

The following types of data reduction were implemented in processing these data: drift, latitude, elevation of gravimeter w.r.t. to the reference geoid, mass and terrain. The effects of the residual tidal and instrumental drift were removed from the data using base station reoccupation drift curve.

Despite the CG-5 precision, the quartz springs it uses to measure gravity experience some fatigue. This induces minute changes in how they respond to the gravitational field and, in turn, how the instrument quantifies that field. The purpose of the instrumental drift correction is to remove any of these effects. The latitude correction is intended to counteract the effect of the outward force induced by the rotation of the earth. The elevation correction works in concert with the mass correction. The first adjusts the data for any variation in height between the instrument and the equipotential geoid and the second removes the effect of the mass between the surface and the equipotential geoid. Lastly, the terrain correction takes into account changes in elevation of the surface surrounding the observation stations and how the mass therein influences measurements (Hinze et al., 2013; Hammer, 1939; Hubbert, 1948).

The raw microgravity data in Tables 9 and 10 were reduced to remove these extraneous influences in accordance with the procedures outlined in Appendix B. Processed microgravity data are presented in sections 3.1.1 and 3.1.2.

2.1.4 Observed Microgravity Data Analysis

Δg_{CBA} profiles were compared with data from the Pan American Center for Earth and Environmental Sciences (PACES). PACES data were collected over the entire United States and further to create gravitational field maps that can be implemented in various forms of potential field analysis. The data retrieved from PACES and this study's data were processed according to the same procedure to allow comparison of their Δg_{CBA} maps with our Δg_{CBA} profiles. While lower in spatial resolution in comparison to this study's microgravity surveys, the PACES data are implemented to evaluate the reliability of our microgravity data and to observe any off-line gravitational field trends surrounding the locations of the Δg_{CBA} profiles.

An understanding of microgravity analysis and its limitations reiterate the purpose of this study's microgravity data: to constrain the location of the hypothesized CU-s. Geophysical potential field theory enables a fault location to be approximated from a Δg_{CBA} profile (Hinze et al., 2013). Theoretically, a plane that separates vertically offset strata of increasing density with depth (e.g. a fault), is located along a profile where the anomaly's slope, $\frac{dg_{CBA}}{dx}$, is greatest (Blakely and Simpson, 1986). Therefore, after the prescribed microgravity data corrections, the slope of the observed anomaly was calculated. Further seismic investigation was planned based on the location of $\frac{dg_{CBA}}{dx}$ maxima within microgravity profiles.

2.2 SEISMIC METHODS

Seismic investigations of the subsurface rely on the principles of wave propagation. “Seismic wave” is a broad term that encompasses various forms of linear-elastic energy propagation through the subsurface (Sheriff and Geldart, 1995): longitudinal P-waves, transverse S-waves, and retrograde elliptical Rayleigh R-waves (Caufield, 2005). In a seismic survey an energy source is created at the surface and as the energy propagates through the subsurface, geophones record the frequency, amplitude, and relative arrival times of the propagating energy waves. For two dimensional studies, the geophones are arranged in a linear array collinear with the seismic source. The geometry of the array and the source, along with seismic wave arrival times, are used to calculate the velocity of a seismic wave as it permeates the subsurface media.

The seismic profile and soundings in this study were conducted with the intent to examine the P-wave seismic response of the subsurface. The energy dispersion associated with a P-wave causes compression at the wave front, normal to its surface, and dilation behind it in response. Due to the principle of conservation of energy, the kinetic energy (KE) of the wave attenuates, or decreases in amplitude, as it travels through the subsurface. The attenuation and seismic velocity of a specific geologic medium is affected by its density (ρ), bulk modulus (K), and shear modulus (μ) (Caufield, 2005). The amplitude of a seismic wave decreases due to geometric spreading of the wave front, partial transmission at impedance boundaries and the conversion of KE into heat at

frictional interfaces (Caufield, 2005). The second type is of particular importance in this seismic investigation.

Impedance boundaries are created between rock of differing mechanical properties (Caufield, 2005). When a seismic wave is partially transmitted across a planar boundary, a fraction of the KE refracts across the boundary and a fraction of the energy is reflected (Caufield, 2005). For non-planar impedance boundaries, such as a fault, the inconsistencies in reflection and refraction angles can cause inconsistent diffraction of incident energy. For these reasons, reflected seismic signals can be used to observe depth changes of seismostratigraphic interfaces and the actual fault between them. Thus, these data provided higher resolution constraints on the location of the CU-s.

The resolution of a P-wave exploration refers the minimum necessary separation between two features required to distinguish between them (Sheriff and Geldart, 1985). Vertical resolution can be determined using the Rayleigh quarter-wavelength ($\lambda/4$) criterion, where wavelength (λ) refers to the dominant wavelength of the seismic waves used to image the features. The Rayleigh $\lambda/4$ criterion suggest that any two features separated by less than $\lambda/4$, will not appear as two distinct bodies in reflections of seismic waves with a wavelength equal to λ (Sheriff and Geldart, 1985). Additionally, a minimum $\lambda/8$ thickness is necessary to detect a feature. However, Sheriff and Geldart (1985) point out that this limit can thin to $\lambda/30$. Thus, recording shorter λ seismic waves yields better resolution. Since λ is inversely proportional to f , increasing the recording frequency also provides better resolution.

Vertical and horizontal resolution are not equivalent. The equation for horizontal resolution, or minimum length (L) of a feature along profile, of a 2D seismic study is:

$$\frac{L}{2} = \sqrt{\frac{\lambda h_0}{2}} = \frac{V}{2} \sqrt{\frac{t}{f}}.$$

where h_0 is the depth of the feature, t is the arrival time of the seismic wave, V the average seismic velocity, and f is the dominant frequency. This equation is derived from the first Fresnel zone equation (Sheriff and Geldart, 1985).

2.2.1 Data Collection

Seismic data were collected coincident with the GP1 profile, on County Road D (Figure 21). This location was determined based on results of microgravity anomaly analysis. For the most part, the chosen site had barren fields on either side of the road. However, towards the end of the CMP profile, County Rd. D was flanked by a thick tree stand. If it had been windy during data collection the rustling of the trees could have added noise to the seismic data, but the air was still that day and there was no noticeable increase in noise in that section of the CMP profile.

For all seismic data a 4.5 kg hammer strike on a 20 cm x 20 cm steel plate was used to generate the P-wave seismic source. The subsurface response was measured by 24, 40 Hz vertical component 7.1 cm spike plant receivers (Mark Products™), spaced 4 m apart. Lower frequency geophones were considered but since higher frequency seismic waves provide better vertical resolution, higher frequency geophones were preferred. Two hammer strikes were vertically

stacked at each shot point in an effort to improve the signal-to-noise ratio of these data. Geophones were connected to two, 4 m take-out cables, which linked them to the recording device: a 48-channel NZXP Geometrics StrataVisor engineering seismograph. The recording window length was 1.024 seconds. Sampling rate was 0.25 ms. During collection, a 60 Hz notch filter was applied to the data to remove unshielded power line interference. There were no observed sources of seismic noise; there were no active farm equipment in adjacent fields and no data were collected while vehicles were driving on County Road D.

2.2.1.1 Seismic Sounding

One P-wave seismic reflection and refraction sounding or walkaway, was performed within the extent of the subsequent CMP profile (see figure 21) using the methods described above. Initial soundings were collected with 48 geophones, 4 m apart. Two seismic sources, one at the first geophone and a second, 84 m to the south-east of the first geophone, were used to image the extent within which the Te, K, and Pz horizons were expected. Two separate shot points created two seismograms. Offsetting the second shot by 84 m meant that the last 24 traces in the first seismogram had reflection mid-points in common with the first 24 traces of the second seismogram. Therefore, traces 25 through 48 were deleted from the first seismogram and traces 1 through 48 of the second were appended. The resulting composite seismogram had 72 traces. This process is illustrated in Figure 22. Assuming laterally homogeneous, flat strata allows the records to be combined in this manner (Vincent et al., 2006).

Figure 22: This figure illustrates how two seismographs combined to create a 72 trace seismic sounding.

Soundings are not intended to examine lateral variation within the subsurface. Their purpose was to determine an optimal CMP array geometry, specifically to determine the necessary offset of the shot point to capture specific seismic signals of interest. An individual P-wave seismogram typically captures four different types of perturbations: the air blast, surface waves, refractions, and reflections. These features are outlined in the P-wave seismogram in Figure 23.

Figure 23: This seismic sounding is used to identify the four types of seismic waves observed in this study's seismic data.

Figure 23 illuminates the purpose of the sounding technique. If a 0 m offset were used in the subsequent CMP profile, each 24 trace seismogram would look similar to traces 1 through 48 of the sounding. Since this study was

concerned with the P-wave reflection signals, it was decided to offset the source from the array by 84 m. In doing so, more of the perturbations recorded at each geophone would be induced by the seismic energy reflecting off of impedance boundaries of interest. Based on Rucker (2017) soundings, these boundaries would be the tops of the Te, K, and Pz strata. If the CMP profile intersected the hypothesized transpressional fault plane, these boundaries would abruptly change in depth across the profile.

2.2.1.2 CMP

The P-wave seismic reflection CMP profile in this study (SP1) was conducted with the equipment described above, with an array geometry similar to the sounding and at a location determined by, and coincident with, GP1 (see Figure 21). Two 4 m take-out cables were placed along the profile and 48 geophones were connected at each take out. Conversely, only 24 geophones were recording during each shot. The energy source was offset 84 m from the first active geophone. Two shots were performed at each source location. The shot point was advanced 4 m at a time, up-line. Before moving to the next shot point, the nearest geophone was turned off and the geophone immediately up-line to the last active geophone was activated to maintain optimal array geometry. After 24 shots, the near-field takeout cable and its 24 geophones were relocated to the far end of the up-line takeout cable. This process was repeated to create the continuous CMP profile comprised of seismic recordings from 192 shot points.

Such a survey records seismic reflections of different ray-path geometries off of common midpoints which increases the signal-to-noise ratio (Burger et al., 2006). The fold of a survey refers to the number of times a single common midpoint is sampled. A higher fold is indicative of a higher signal-to-noise ratio. To calculate fold, the following equation is used:

$$\text{Fold} = (RS \times NR) \div (2 \times SS),$$

where RS is receiver spacing, NR is the number of receivers and SS is the shot spacing. Following the data collection procedure above resulted in a 12-fold CMP profile. Processing these data produced the first image of the hypothesized southern boundary fault of the CU.

2.2.2 Data Processing

2.2.2.1 Seismic Sounding

Seismic sounding data were processed in VISTA 13.0. First, the field files were converted into the SEG-Y data format used by the program. This was followed by two gain functions. The first, a spherical divergence gain, is used to counteract the seismic signal attenuation due to geometric spreading of the wave-front (Sheriff and Geldart, 1982). The second, a mean gain, normalizes the amplitude of all seismic waves based on an average amplitude of all recorded signals.

This was followed by the application of a bandpass filter. A bandpass filter removes seismic signals above and below the intended frequencies. Based on our P-wave data, the frequencies of interest were between 40 and 120 Hz. To reduce ringing effects associated with bandpass filters with discrete cutoff

frequencies, the true extent of the bandpass filter ranged from 30 Hz to 160 Hz. The seismic signals with frequencies between 30 to 40 Hz and 120 to 160 Hz were tapered. For example, suppose trace 1 exhibits purely 38 Hz seismic signal and trace 2 exhibits purely 32 Hz signal. Both of their amplitudes would be reduced but the amplitude of the signal in trace 2 would be reduced by more than that of trace 1

Next, another gain function was used: automatic gain control (AGC). The AGC function is similar to the mean gain function, except that it normalizes signal amplitudes within specific time windows. The time window used to process these data was set to 100 ms, so all seismic signals between 0 and 100 ms were normalized in amplitude using a scaling factor based on the average amplitude of all recorded signals within the 0 to 100 ms recording window. This process is automatically repeated for subsequent time windows. These filters and gains increase the amplitude of seismic reflection signals and decrease the amplitude of all other noise, either random or coherent.

Subsequently, each seismic record for a given geophone, or trace, was then inspected to determine if a mute was necessary. The purpose of this process is to delete the seismic signals recorded by a geophone that was poorly planted in, or coupled with, the surface. Muting such traces improves the overall signal to noise ratio of the data (Baker, 1999). Next, it was necessary to create a geometry file. Creating a geometry file spatially correlates the seismic data. The defined locations of each shot point and each geophone were applied to data

headers. These processes were performed on all sounding data. The resulting seismograms were spliced together as described in section 2.2.1.1.

2.2.2.2 CMP Profile

Seismic sounding data were process first in VISTA 7.0 and then again in VISTA 13.0. The first processing steps of the CMP profile were the same as those described above; however, additional steps were required to produce the composite image consisting of all seismograms recorded along the profile.

For a CMP profile, following the creation of a geometry file, top and bottom mutes were applied. Since refraction signals arrive at each geophone before reflection signals in these data, the top mute is used to delete most of the refraction signals (see Figure 23). Since it arrives at any given geophone after reflection signals, most of the surface wave signals (see Figure 23) were removed with the bottom mute. These two mutes were applied to all data, 24 traces at a time.

The following processing procedure is referred to as a normal moveout correction, or NMO. This correction stretches or compresses seismic signals within each trace to account for variation in arrival time relative to the distance between a source and the recoding geophones (Burger et al., 2006). To do so, semblance analysis was performed. Where each 24 trace segment was examined in the VISTA interactive velocity analysis window. This subroutine displays each trace segment next to a semblance cross-correlation plot. The first step was to select the arrival time of the first observed reflection signal with each 24 trace segment. The curving reflection signals in the data are then overlain

and compared to the hyperbolic x^2 over t^2 , or velocity², curves to determine how well they correlate. VISTA uses the selected hyperbolas that correlate with a reflection signal to calculate the normal moveout time for each signal within a trace. The normal moveout time is then subtracted from the two-way-travel-time (TWTT) for each trace, yielding the zero-offset time for each reflection. More specifically this equation is:

$$t_{\text{TWTT}}^2 - \frac{x^2}{v^2} = t_0^2,$$

where t_{TWTT} is the TWTT, t_0 is the zero-offset time, x is the distance between the source and a given geophone and v is the calculated velocity of the sediment between the surface and the reflecting impedance boundary (Yilmaz, 2001). This step effectively flattens the first set of reflection signals within each 24 trace segment, to represent the first impedance boundary's response as if it were imaged with a seismic wave that was incident to the boundary at a 90° angle.

Since seismic velocity changes across impedance boundaries, the seismic waves reflected from subsequent impedance boundaries travel at different velocities. To take this fact into account when representing the data, VISTA uses the velocity² curves of the subsequent impedance boundaries to determine what is referred to as an interval velocity, or the velocity of a specific layer of sediment. For example, the interval velocity of the second layer of sediment is determined by the Dix equation:

$$V_i^2 = \frac{V_n^2 t_n - V_{n-1}^2 t_{n-1}}{t_n - t_{n-1}},$$

where V_i is the seismic velocity of the second layer, V_n is the root mean square seismic velocity of the first and second layers, t_n is the zero-offset time of the

second impedance boundary, V_{n-1} is the seismic velocity of the first layer and t_{n-1} is the zero-offset time of the first impedance boundary (Yilmaz, 2001).

This step effectively flattens the second impedance boundary's reflection signals within each 24 trace segment, to represent each reflection signal as if it were imaged with a seismic wave of a constant velocity that was incident to the boundary at a 90° angle. VISTA automatically performs this step for the second and any subsequent layers of sediment. This process is repeated for each 24 trace segment.

After the NMO correction, the seismic signal in each data trace is sorted by the common midpoint from which it reflects. The sorted traces are then stacked. Processing and plotting the seismic data in this manner improves the signal-to-noise ratio and enables individual horizons to be identified along the extent of the profile.

The final step was a depth conversion. The calculated root mean square velocity of all sediment layers and the calculated zero-offset times can be used to determine interval thickness (h_n) (Yilmaz, 2001):

$$h_n = V_n \left(\frac{t_n - t_{n-1}}{2} \right).$$

Calculation of interval thickness enables the seismic signals within each trace to be appropriately plotted in depth. This is particularly useful in determining and quantifying the magnitude of seismic horizon vertical displacement across a fault. Additionally, these data provided localized constraints on horizon depths which were implemented in the revision of the microgravity anomaly model.

CHAPTER 3. RESULTS

3.1 GRAVITY DATA

3.1.1 GP1

The observed potential field values within GP1 were processed to produce the Δg_{CBA} profile shown in Figure 24, where GP1 has a maximum Δg_{CBA} of $1.616 \pm .004$ mGal (error calculation discussed in Appendix B). The processed microgravity data are shown in Table 3. The greatest $\frac{dg_{CBA}}{dx}$ in GP1 was between stations 3 and 4, suggesting the CU-s intersected the profile between these two stations (Hinze et al., 2013; Blakely and Simpson, 1986). This section of GP1 is highlighted in green in Figure 24.

Table 4: This table lists the GP1 processed microgravity values.

station	Δg_{CBA}
s0	0.001813
s1	-0.200000
s2	-0.073604
s3	-0.029272
s4	0.484662
s5	0.789996
s6	1.257254
s7	1.359833
s8	1.416103
s9	1.395005
s10	1.309795
s11	1.046497
s12	0.919579

Figure 24: This figure shows the GP1 Δg_{CBA} profile. Triangles show processed microgravity values at stations 0 through 12 (from right to left). The red lines between them are linearly interpolated. The green shade between stations 3 and 4 indicate the section of greatest slope in the anomaly.

To compare the forward model with GP1, first, the observed data were interpolated using the cubic spline method. Next, $\frac{dg_{CBA}}{dx}$ between interpolated points was calculated. Within GP1, the maximum slope of the interpolated anomaly was located 0.73 km from station zero. The locations of observed and calculated anomaly slope maxima were equated by adjusting the x-axis location of the fault in the subsurface model to 0.73 km. In Figure 25, the observed and interpolated anomaly (green line) is compared with the calculated anomaly (black line) from Model 1.

Figure 25: At the top of this figure, the observed (cubic spline interpolated) GP1 microgravity anomaly and the calculated anomaly are plotted. The subsurface model parameters were derived from Rucker (2017) North Wyatt soundings B and C (NWYB, NWYC) and Bexfield et al. (2006) density data.

The magnitudes of the calculated and observed anomalies were nearly equal: ~ 1.6 mGal and $1.616 \pm .004$ mGal, respectively, suggesting that GP1 had intersected the hypothesized southern fault of the CU. The greatest differences between observed and calculated anomalies, shown therein, is derived primarily from three observed data points at the locations furthest from the fault. These

endpoint discrepancies could be induced by slight changes in stratigraphic depths not accounted for in our single fault model. Such changes are expected based on the isopach maps derived from Rucker's (2017) seismic soundings. Anomaly discrepancies could also arise from density variations within individual strata, also not accounted for in this model. These factors considered, our analysis of microgravity data focus on the extent of GP1 closest to the modeled fault. In this section of GP1, the slopes of the observed and the calculated anomalies are similar. These analyses further serve to suggest that GP1 intersected the CU-s.

3.1.2 GP2

The observed potential field values in the second microgravity profile (GP2) were processed to produce the Δg_{CBA} profile shown in Figure 26. The processing steps taken to correct GP2 data were the same as those implemented in processing GP1 data. The processed microgravity data are shown in Table 5. The calculated error of this survey was ± 0.015 mGal. While still small in magnitude compared to the overall magnitude of the anomaly, this error was more than three times greater than that of GP1. We attribute this to windy conditions during GP2 data collection. The gravimeter was shielded with an umbrella during data collection but this was not a perfect solution.

Table 5: This table lists the GP2 processed microgravity values.

station	Δg_{CBA}
s0	1.081970
s1	0.632826
s2	0.360466
s3	0.040463
s4	0.565563
s5	0.089672
s6	0.000000
s7	0.473071
s8	0.498048
s9	0.657990
s10	0.627701
s11	0.590973
s12	0.747920
s13	0.815798

Figure 26: This figure shows the GP2 Δg_{CBA} profile. Triangles show processed microgravity values at stations 0 through 13 (from right to left). The red lines between them are linearly interpolated.

The GP2 anomaly did not match our model, both in character and magnitude of the expected anomaly. We explored multiple explanations as to why. First, GP2 was conducted at a $\sim 45^\circ$ angle to the hypothesized fault plane. Our model calculates the gravitational response for a profile that intersects the fault normal to its fault plane. At an angle, the calculated anomaly would appear as if it were stretched along its x-axis. However, relative changes in microgravity values would be the same in magnitude. Therefore, the calculated anomaly still does not correspond to GP2 measurements.

A second possibility is that GP2 did not transect the CU-s. Boundary traces of the CU likely express sinuosity along strike (Rucker, 2017), a characteristic noted in Figures 5, 6 and 7 (Pratt, 2012). This implies that a perfectly linear singular fault strand may be an oversimplification of the actual CU character. If the CU-s does express such sinuosity, the fault strike may bend slightly to the south-west before its intersection with GP2. If this is the case, it could explain the upward trend in GP2 measurements seen in the last 2 km of the profile. This section of GP2 may be the first half of the anomaly in Model 1. Since no seismic data were collected coincident with GP2, such an interpretation is purely speculation.

3.1.3 Data analysis

Figure 27: These figures compare this study's anomaly profiles with cross-sections of the Bouguer gravitational field from the PACES database. 27a (top), GP1 and corresponding PACES cross-section. 27b, (bottom) GP2 and corresponding PACES cross-section. GP1 and GP2 data points are linearly interpolated.

Because GP2 did not match Model 1, we used gravitational field data from the Pan American Center for Earth and Environmental Sciences (PACES) database to confirm the validity of our microgravity measurements in GP1 and GP2. Figures 27a and 27b show cross-sections of the Δg_{CBA} field from the PACES database compared to GP1 and GP2 observations at their respective locations.

We interpret differences in anomaly magnitudes, specifically between GP1 and the cross-section of the PACES's potential field, to be an effect of data resolution. The PACES database had two actual observation points within GP1, near its first and last stations (shown in Figure 28). Since GP1 has far more points of observation along the profile than PACES, our data have a much better ability to incorporate the effects of abrupt potential field changes. A short period variation in gravity, like that in the GP1 anomaly, induced by an abrupt change in lateral density (i.e. vertical displacement across a fault), would not be as apparent from only two observations at the start and terminus of the profile. Since the two coincident observation points in the gravitational field cross-section from PACES almost match our observations at these points, our microgravity data in GP1 are judged to be reasonable.

The overall trends of the gravitational field cross-sections are similar to both GP1 and GP2 observed anomalies, except for one data point. At station 4 in GP2 we observed a microgravity value 0.520 mGal greater than the value reported by PACES. The width of the gravitational high centered at this station is 633 m. Sharp anomalies like this are attributed to proportionally sharp density

contrasts within the shallow subsurface (Figure 17a). Based on GP2 microgravity data and an understanding of the regional history and geology, we speculate that the anomalous high at station 4 could be induced by either a near surface mafic dike or a pyrite mineralization.

We consider the possibility of a mafic dike in this region to be plausible since Hildenbrand and Hendricks (1995), using magnetic and gravity data, show that at least 8 plutons have intruded below the ME. While plutons are much larger, their presence in the area indicates that the development and intrusion of a mafic dike would also be possible. Hinze et al. (2013) provides a depth approximation formula for a narrow dike:

$$Z_t \approx 0.7 \times X_{1/2},$$

where Z_t is the depth to the top of the dike and $X_{1/2}$ is half the width of the gravitational anomaly. Using this formula, we calculate the depth to top of the potential dike to be ~222 m below the surface. Indicating the top of the dike would intrude into the Cretaceous sediment in Model 1, which Bexfield et al. (2006) estimates to have a density of 2.4 g/cc. The Bloomfield Pluton, located less than 50 km to the west of our study area, at a depth of ~2 km below the surface, consists of igneous rock ranging in density from 2.7 to 3.0 g/cc (Ravat et al., 1987). Thus the intrusion of similarly dense rock would create a horizontal density contrast that could be used to explain the anomalous spike at station 4. However, a mafic dike intrusion this far into the shallow subsurface has not been observed in other studies of the NMSZ.

Alternatively, the same gravitational signature could be produced by a pyrite deposit. Analysis of cores drilled in the ME revealed extensive and large pyrite deposits within the McNairy Sand stratigraphic layer (Crone, 1981). In Model 1, the McNairy Sand is in the Cretaceous stratigraphic layer, which has a density of 2.4 g/cc (Bexfield et al., 2006). The density of pyrite is 5.0 g/cc (Stevens et al., 1980), and thus the horizontal density contrast would be stark. If the GP2 survey encountered such a deposit, it could explain the anomalous high at station 4.

Both of the potential sources used to speculate on the GP2, station 4 microgravity measurement would be inherently narrow. Thus, its gravitational influence would not be well quantified in microgravity measurements from a distance similar to those used to create the PACES potential field map. This is why the anomalous high in GP2 is not represented in the cross-section of the PACES potential field map. Otherwise, GP2 and the PACES cross-section closely match. Therefore, we again interpret this to be an effect of data resolution and judge GP2 microgravity data to be reasonable.

Figure 28: This map shows the Bouguer gravity field in the study area. GP1 location plotted with a black line. GP2 location plotted with a white line (data from PACES).

Because the objective for the microgravity surveys was to constrain anomalous uplift targets for higher-resolution seismic reflection imaging and since GP1 exhibits an anomaly that correlates more closely with the hypothesized CU-s intersection, the target for seismic investigation was decidedly GP1. Slope analysis of this transect was used to determine the location of seismic investigations and is compared with subsequent models.

3.2 SEISMIC DATA

3.2.1 Seismic Soundings

To further constrain the intersection of the CU-s with GP1, seismic methods were employed as described in Chapter 2. The seismic sounding performed coincident with GP1 was spliced together as described in section 2.2.1.1. The interpreted Te, K, and Pz horizons in this sounding are also shown in Figure 29. The identification of each impedance boundary as a Te, K, or Pz horizon was determined by comparing sounding data to other P-wave soundings done within the region (i.e. Rucker, 2017 and Pryne et al., 2013).

Figure 29: Prior to collecting SP1 data, this seismic sounding was conducted in situ, to determine optimal acquisition parameters for SP1. Arrows indicate the interpreted Tertiary (Te), Cretaceous (K) and Paleozoic (Pz) horizons.

Rucker (2017) NWYC sounding shows similar TWTT for the Te, K, and Pz horizons. Both the NWYC sounding and the sounding in Figure 29 were conducted off-uplift. The spliced and interpreted seismographs are compared in Figure 30. The TWTT to the Te horizons in both seismograms was about 210 ms. However, the TWTT to K, and Pz horizons in our sounding were greater than those in Rucker (2017) NWYC sounding, by 30 ms and 20 ms, respectively. This indicates that the Te stratigraphic layer at our location is thicker than that of the NWYC site. The horizon interpretations in our seismograph are reasonable since in Rucker (2017) the other off-uplift soundings of the same transect (NWYD and NWYE) showed greater variation in TWTT of the K and Pz horizons than that of the discrepancy between Rucker (2017) NWYC sounding and this study's (see Table 1).

Figure 30: a) is the NWYC off-uplift seismic sounding with Te, K, and Pz horizons identified as interpreted by Rucker (2017). b) is this study's sounding, also conducted off-uplift, with the interpreted Te, K, and Pz horizons identified. Perfect correlation between the two soundings was not expected. Rucker (2017) NWYC through E, all off-uplift, showed comparable variations in TWTT of corresponding strata. These data indicate that Tertiary strata at our location is thicker than Tertiary strata at the NWYC site (30a modified from Rucker, 2017).

3.2.2 SP1

The location and extent of the SP1 profile (shown with a solid blue line in Figure 21b) was intended to encompass stations 3 and 4 of GP1. Seismic data were collected and processed as described in section 2.2. The first processed (in VISTA 7.0) and uninterpreted seismic image of the SP1 profile is shown in

Figure 31a. No depth conversion was performed on these data, which is why they are plotted in the time domain. The same data were reprocessed in VISTA 13.0 and the resulting image is shown in Figure 31b. A depth conversion was performed on these data. Figure 31b is used in all subsequent analyses of the SP1 profile.

Figure 31: This figure shows the uninterpreted CMP seismic reflection images, SP1 profile. For reference, x-axes are consistent with GP1 profiles, in that distances show are w.r.t. Station 0 of GP1. Locations of microgravity observation stations coincident with SP1 are indicated by triangles. 31a (top) and 31b (bottom) were processed similarly, the main difference is that no depth conversion was performed on data in 31a.

Since the acquisition parameters of SP1 were chosen specifically to resolve the Te, K, and Pz horizons, expected between 0.1 km and 0.4 km depth, seismic data above and below this depth section is not interpreted. Above 0.1 km, the optimized acquisition array can distort the compressional seismic waves due to the wide angle of seismic wave incidence in the near surface. In addition to increased susceptibility to lateral velocity variations, amplitude anomalies and

phase changes, as defined by Zoeppritz equations, can degrade overall signal quality for the final stacked image.

In addition to surface wave interference and attenuation effects, any direct reflections from impedance boundaries below 0.4 km were too weak in amplitude to be observed. The tops of the K and Pz strata are stark impedance boundaries, meaning that most of the incident seismic source energy is reflected. The remaining seismic energy that is transmitted and reflects off of subsequent impedance boundaries is too low in amplitude to be observed by a survey of this design. The Dow Chemical seismic profile also exhibits these effects (Figure 8a).

Figure 32: Interpreted CMP seismic reflection image, SP1 profile. For reference, x-axis is consistent with GP1 profiles, in that distances shown are distances w.r.t. GP1 Station 0. Locations of microgravity observation stations coincident with SP1 are indicated by triangles. The Te, K and Pz horizons are overlain with a yellow dashed line at both ends of the profile. They show vertical variation across the profile. The green shaded section of SP1 is the section of GP1 that had the maximum slope in the observed anomaly. See Appendix D for a more detailed fault interpretation.

Table 6: This table lists the depth values in meters to the top of the horizons identified in SP1. Te: Tertiary, K: Cretaceous, Pz: Paleozoic.

Reflector	Depth to hanging-wall horizon (SP1) (m)	Depth to foot-wall horizon (SP1) (m)	Vertical Offset (m)
Te	115	140	35
K	190	225	35
Pz	310	380	60

The Te, K and Pz horizons are identified in Figure 32. Their interpreted depths and vertical offsets are listed in Table 6. The interpreted horizons identified in SP1 have similar depth values to those used in microgravity anomaly Model 1 (from Rucker, 2017 NYWB and C soundings in Table 2). Overall, the depth to the top of each strata in SP1 was slightly greater. SP1 horizon interpretations are reasonable since Rucker (2017) soundings at different locations did not show consistent stratal depth or thickness. However, each horizon shows similar magnitudes of vertical offset across SP1. The magnitude of the difference between depth and offset values are shown in Table 7.

Table 7: Lists the magnitude of difference in depth and offset for the Te, K, and Pz horizons between those observed in Rucker (2017) NWYB and C soundings and those observed in SP1.

Reflector	Difference in depth to hanging-wall horizon (m)	Difference in depth to foot-wall horizon (m)	Difference in offset (m)
Te	1	3	4
K	23	12	11
Pz	85	94	1

The green shaded section of SP1 in Figure 32 is where maximum $\frac{dg_{CBA}}{dx}$ was calculated between observed microgravity values in GP1. Within that section of SP1, these horizons show noticeable vertical deformation. Furthermore, the interpreted fault in Figure 32 intersects this region, indicating that the fault location estimated through GP1 slope analysis correlates well with the location of our seismic interpretation at the CU-s intersection location with SP1. As expected, the fault is nearly vertical (i.e., apparent dip = 86° from horizontal).

The identified horizon depths in SP1 were used to calculate the thickness of each strata. Table 8 shows the results of these calculations. We attribute most of the difference in layer thicknesses across the fault to be a result of syndepositional faulting, indicating that the CU-s experienced some vertical component of movement during both the Tertiary and Cretaceous periods. Since no horizons above 0.1 km depth could be confidently interpreted as offset, the data in SP1 cannot be used to comment on the activity of this fault during the Quaternary.

Table 8: This table lists the thickness of strata observed in SP1.

Strata	Thickness in Hanging-wall (SP1) (m)	Thickness in foot-wall (SP1) (m)
Te	75	85
K	120	155

3.3 GRAVITY MODEL REVISIONS AND ANALYSES

The parameter values used to create the subsurface of Model 1 were revised following seismic investigation. Since these data were collected coincident with GP1, they provided far more localized constraints on the subsurface structure. In all subsequent models, these constraints are applied.

The discrepancy between the depths to the Te, K, and Pz horizons seen in Rucker's (2017) NWY soundings and the depths of the same strata observed in SP1 have implications for our microgravity anomaly model. As illustrated in Figure 17, changing the depth of a stratigraphic layer changes the magnitude of the calculated anomaly. Additionally, the dip angle of the fault would slightly skew model results (Hinze et al., 2013). Using the localized depth estimates and the 86° fault dip observed in SP1, Model 2 was produced, shown in Figure 33. This model does not match the anomaly observed in GP1. The magnitude of this calculated anomaly was ~1.5 mGal, yielding a $.116 \pm .004$ mGal difference in magnitude between the observed and calculated anomaly.

section of the model, from 0 to 0.1 km depth. When compiled, these data yielded lithological columns at each observation station, and a more precise picture of the shallow subsurface than could be determined from SP1.

Pryne (2012) identified and described 6 lithological classes. However, they did not report density estimates for these classes. In order to augment the revised anomaly model, such estimates were required. Using a variety of sources (Bodman and Constantin, 1965; Crone, 1981; USGS water table data; Aslan and Autin, 1998; Osipov, 2011; Mehdiratta and Triandafilidis, 1978; Lo Presti, et al., 1992) likely densities were assigned and microgravity Model 3 was produced (Figure 34). The process of determining model structure and specific densities is covered in depth in Appendix C.

Figure 34: Model 3 - At the top of this figure, the plot shows the microgravity anomaly signature that was calculated from the shallow subsurface model in the middle of the figure. This subsurface model was created with Pryne (2012) lithology data. At the bottom are color coded density estimates that correspond to lithologies in the shallow subsurface model.

The resulting subsurface model contained horizontal density contrasts. The Oasis Montaj program calculated the total variation in the gravitational field induced by these contrasts for the extent of GP1. The magnitude of the calculated anomaly was ~ 0.25 mGal. The calculated anomaly in Model 2 does not account for the shallow subsurface density variation in Model 3. The gravimeter, however, recorded their influence on the gravitational field. In order to make more accurate inferences about discrepancies between observed and calculated anomalies, the subsurface in Model 2 was augmented to incorporate Model 3's shallow subsurface lateral density contrasts. The result is shown in Figure 35: Model 4.

Figure 35: Model 4 - On top, the observed (cubic spline interpolated) GP1 microgravity anomaly and the calculated anomaly are plotted. This subsurface model combines the subsurface parameters of Models 2 and 3. Model parameters determined from Pryne (2012) data constitute 0 to ~0.1 km depth. Below ~0.1km depth, model parameters reflect the horizon depths along with the fault dip and location observed in SP1.

Including Pryne (2012) shallow subsurface data in the model, along with horizon depth and fault geometry determined from SP1, produced a calculated anomaly that more closely matched our observations in GP1. Figure 36 quantifies the differences therein; this difference curve is referred to as the residual anomaly. While less in magnitude, the model still poorly correlates to GP1 at end points. For reasons described in section 3.1.1, we do not speculate as to the causes of potential field variation at end points.

Figure 36: The residual anomaly in this figure was calculated by subtracting the calculated anomaly values, in Model 4, from the interpolated microgravity values in GP1.

However, variation between Model 4's calculated anomaly and the GP1 anomaly between stations 2 through 6, is of importance. The discrepancy between these stations could be rooted in the interpretation of the fault in SP1,

the interpretation of the magnitudes of offset between corresponding seismic horizons, the density and structure estimates made pertaining to Pryne (2012) shallow subsurface data, or all of the above. However, if all of our estimates and assumptions accurately reflect the subsurface below GP1; it may be the case that, between stations 2 and 6 (Figure 36) the sinusoidal section of the residual anomaly may be the result of an abrupt lateral change in density in the subsurface not accounted for in the model due to missing or incomplete borehole data from Pryne (2012).

The narrowness of the sinusoidal section in the residual anomaly suggests that lateral density contrasts exist at a proportionally shallow depth within the subsurface along GP1 (Hinze et al., 2013). In order to induce a similar residual anomaly at depth, the density contrasts would be greater than reasonable, based on Bexfield et al. (2006), and the overall correlation of GP1 with the calculated anomaly in Figure 35. Supposing this density contrast is the result of vertical displacement along a shallow transpressional extension or a fault-splay of the CU-s, the depth approximation equation for a vertical fault:

$$Z_c \leq X_{1/2},$$

can be applied to the residual anomaly to calculate an approximate maximum depth to the center of the supposed fault (Z_c), where $X_{1/2}$ is half of the anomaly's width at half of the anomaly's maximum magnitude ($A_{1/2}$) (Hinze et al., 2013).

Figure 37 illustrates how to determine the variables presented in this equation.

Figure 37: This figure illustrates the method used to assign values to variables used in the shallow fault depth approximation equation above (from Hinze et al., 2013).

By applying this method to the residual anomaly we approximate the depth of a fault that would be necessary to correlate the observed and calculated anomalies. Determination of values for these variables based on the residual anomaly is shown in Figure 38. It shows that $X_{1/2} = .08$ km; indicating that lateral density contrasts responsible for the discrepancy between the observed and calculated anomalies, between stations 2 and 6, could be induced by transpressional movement along a vertical fault with a depth-to-center equal to 80 m.

Figure 38: This figure shows how the residual anomaly was used to determine the values of variables for the depth approximation equation. $X_{1/2}$ was equal to 0.08 km. Substituting 0.08 km for $X_{1/2}$ in the equation above yields $Z_c \leq 80$ m. This means that the maximum depth to the center of a vertical fault would be 80 m below the surface. The horizontal density contrasts associated with such a fault would reduce the difference between the observed and calculated anomalies.

Since Model 4 incorporated Pryne (2013) shallow lithology data (0 to 100 m deep) and residual anomaly analysis suggests faulting occurs at 80 m depth and shallower, the question arises: Why doesn't the shallow subsurface model include the horizontal density contrasts associated with such a fault? To answer this, calculation of $\frac{dg_{CBA}}{dx}$ along the residual anomaly was used to determine the location of the potential shallow fault predicted through residual anomaly

analysis. The residual anomaly has a gradient maxima between GP1 Stations 3 and 4, specifically at 0.7 km from GP1 Station 0. The cross-section of Pryne (2012) lithology data show that there are horizontal density contrasts at 80 m deep, between Stations 3 and 4 but they cannot be conclusively identified as a fault. However, closer examination of the uninterpreted Pryne (2012) lithology cross-section in Table C1 (Appendix C) reveals that in order to produce Model 4, assumptions were made about lithology data reported as missing or damaged within the cross-section. A continuous section of missing or damaged lithology data was reported by Pryne (2012) between GP1 Stations 1 through 5, from 27 to 55 m below the surface. The center of the shallow fault predicted by residual anomaly analysis is located 35 m directly below this extent of missing data. 80 m is the depth approximation value for the center of the inferred shallow extension of the CU-s. This implies that faulting may have occurred closer to the surface and that the associated horizontal density variations (i.e. vertical offset of lithological layers) were not incorporated in the Pryne (2012) data and therefore not reflected in Models 3 and 4. The spatial correlation of the inferred shallow fault and the missing data serve to explain why correlation of the observed and calculated anomalies decreases between GP1 Stations 2 through 6.

CHAPTER 4. CONCLUSIONS

We conclude that the CU southern boundary fault intersects GP1 between microgravity observations stations 3 and 4 based on the slope of the observed Δg_{CBA} profile. In Figure 39a the section of GP1 with the maximum $\frac{dg_{CBA}}{dx}$ is highlighted in green. The green section of GP1 is projected onto SP1 in Figure 39b. Furthermore, we conclude that SP1 has directly imaged this fault at the location shown in Figure 39b. The fault identified in SP1 intersects the projected green shade, indicating that the microgravity slope analysis method provided a reliable constraint on the location of the CU-s. The fault interpretation is reinforced by the correlation between the measured and calculated anomalies in Figure 39a. Their correlation also indicates that GP1 has a near perpendicular intersection with the CU-s. This is consistent with the strike hypothesized by Pryne et al. (2013). Additionally, the seismic horizons associated with the tops of the Te, K, and Pz strata all show greater depth on the south-east side of the fault. This indicates that the CU is a tectonically controlled structure.

Figure 39: This composite figure is a spatial correlation of GP1 and the interpreted P-wave seismic reflection CMP profile, SP1. Figure 39a compares the observed (GP1) and calculated microgravity anomaly (from Model 4) and highlights the section of maximum observed anomaly slope with green shading. The blue line along the x-axis of 39a shows the location of SP1 relative to GP1. Figure 39b is the 12-fold CMP interpreted image from SP1. SP1 acquisition parameters were optimized for identifying the Tertiary, Cretaceous and the Paleozoic horizons. The seismic horizons that correlate with the tops of these strata are shown with dashed yellow lines in 39b. In 39b, the section of maximum slope within the observed anomaly is projected onto SP1 with green shading. This region of SP1 shows vertical variation of reflector depths and is intersected by the CU-s.

From its location in SP1, a linear extrapolation of the imaged CU-s, at a N46°E strike (Pryne et al., 2013; Rucker, 2017) to the south-west, shows that it intersects contemporary seismicity, the Farrenburg Lineaments and the NMNF (Figure 40). These observations along with the fact that the rate of seismicity is

elevated in this region with respect to the midcontinent surrounding the NMSZ (Chiu et al., 1992; Petersen et al., 2014), lead us to conclude that the CU is structurally connected to the NMNF. This discovery extends the NMSZ a minimum of 30 km to the north-east; from the currently defined NMNF terminus at the Farrenburg Lineaments (Baldwin et al., 2005), directly through East Prairie, to 6.5 km south-east of Charleston, Missouri.

The structural connection between the CU and the NMNF has potent implications for seismic hazard assessment in the region. Johnston and Schweig (1996) suggested that the 23 January 1812 earthquake occurred on the NMNF. Since the CU and the NMNF are linked, seismicity from a similar event originating in the NMNF could propagate along the CU and would pose greater risk further to the north-east than previously estimated.

Figure 40: The fault imaged in SP1 (CU-s) is projected to the southwest using the N46°E fault strike suggested by Pryne et al. (2013). The projected fault plane intersects contemporary seismicity (USGS), the Farrenburg Lineaments (Baldwin, 2005) and seismicity associated with the NMNF (Chiu et al., 1992; Csontos et al., 2008; Csontos and Van Arsdale, 2008). Seismicity data from USGS earthquake catalogue. Road data from Missouri Spatial Data Information Service.

APPENDICES

APPENDIX A: OBSERVED MICROGRAVITY DATA TABLES

Table A1: This table lists the GP1 unprocessed input data. t: observation time, g_obs: observed microgravity value, lat: latitude at the station, lon: longitude at the station, h of CG-5: height of the gravimeter w.r.t. the surface, elv_s w.r.t. s3: elevation of the surface w.r.t. the surface at Station 3.

t	station	g_obs (mGal)	lat	lon	h of CG-5 (m)	elv_s w.r.t. s3 (m)
18:43:51	b1_-2	5979.348	36.8549	-89.2573	0.142	
18:47:18		5979.347				
18:18:51	s-3	5979.465	36.8671	-89.2747	0.158	0.076
18:24:59	s-3	5979.469				
18:05:32	s-2	5979.469	36.8683	-89.2764	0.142	-0.330
18:08:01	s-2	5979.473				
17:55:48	s-1	5979.599	36.8695	-89.2781	0.150	0.076
17:58:14	s-1	5979.599				
17:42:00	s0	5979.762	36.8707	-89.2798	0.137	0.000
17:44:30	s0	5979.769				
17:19:56	b1_-1	5979.339	36.8549	-89.2573	0.140	
17:22:41	b1_-1	5979.334				
17:13:07	b1_1	5998.528	36.8549	-89.2573	0.164	
17:17:01	b1_1	5998.625				
17:35:00	s0	5998.436	36.8707	-89.2798	0.162	0.000
only one cycle of g measurement						
17:43:45	s1	5999.032	36.8717	-89.2813	0.159	0.051
17:45:42	s1	5999.024				
18:06:19	s2	5999.404	36.8723	-89.2822	0.168	0.006
18:07:58	s2	5999.407				
18:26:41	b1_2	5998.608	36.8549	-89.2573	0.165	
18:28:30	b1_2	5998.606				

0.78016204	s3_1	6000.17	36.8754	-89.2865	0.154	-0.060
0.78188657	s3_1	6000.17				
18:52:23	s4	6000.402	36.8771	-89.2889	0.167	0.013
18:54:10	s4	6000.402				
19:02:08	s5	6000.534	36.8782	-89.2905	0.171	0.111
19:04:09	s5	6000.539				
19:21:30	s6	6000.551	36.8790	-89.2917	0.178	0.261
19:23:58	s6	6000.556				
19:32:47	s7	6000.543	36.8796	-89.2925	0.179	0.156
19:34:36	s7	6000.547				
19:50:49	b1_3	5998.627	36.8549	-89.2573	0.170	
19:52:52	b1_3	5998.633				
20:10:59	s8	6000.491	36.8811	-89.2947	0.168	-0.054
20:12:44	s8	6000.497				
20:20:04	s9	6000.442	36.8822	-89.2962	0.154	0.089
20:21:44	s9	6000.425				
20:35:48	s3_2	6000.226	36.8754	-89.2865	0.165	-0.060
20:38:18	s3_2	6000.231				
20:51:46	b1_4	5998.678	36.8549	-89.2573	0.170	
20:53:40	b1_4	5998.685				

Table A2: This table lists the GP2 unprocessed input data. t: observation time, g_obs: observed microgravity value, lat: latitude at the station, lon: longitude at the station, h of CG-5: height of the gravimeter w.r.t. the surface, elv_s w.r.t. s0: elevation of the surface w.r.t. the surface at Station 0.

t	station	g_obs (mGal)	lat	lon	h of CG-5 (m)	elv_s w.r.t. s0 (m)
19:17:28	b2_1	5976.038	36.8486	-89.3081	0.155	
19:20:58	b2_1	5976.034				
19:33:10	s0	5976.874	36.8490	-89.2882	0.148	0.000
19:38:11	s0	5976.882				
19:50:41	s1	5976.457	36.8488	-89.2910	0.122	-0.102
19:56:22	s1	5976.454				
20:03:56	s2_1	5976.126	36.8588	-89.2940	0.138	0.1016
20:06:58	s2_1	5976.125				
20:16:20	s3	5975.855	36.8487	-89.2970	0.1565	-0.102
20:19:19	s3	5975.839				
20:27:59	b2_2	5976.03	36.8486	-89.3081	0.149	
20:30:24	b2_2	5976.025				
20:24:57	s4	5975.704	36.8487	-89.3000	0.157	2.654
20:46:08	s4	5975.671				
20:57:51	s5	5975.679	36.8489	-89.3026	0.139	0.711
no time recorded, discarded						
21:08:56	s6	5975.769	36.8487	-89.3041	0.137	-0.051
21:11:03	s6	5975.769				
21:23:34	s7	5976.136	36.8485	-89.3118	0.141	0.292
21:25:58	s7	5976.151				
21:35:20	s8	5976.306	36.8485	-89.3161	0.146	-0.241
21:38:52	s8	5976.273				
21:49:41	b2_3	5975.981	36.8486	-89.3081	0.149	

21:51:23	b2_3	5975.981				
22:00:58	s9	5976.31	36.8484	-89.3191	0.142	0.241
22:04:54	s9	5976.313				
22:13:08	s10	5976.354	36.8484	-89.3211	0.142	-0.102
22:14:58	s10	5976.355				
22:25:08	s11	5976.247	36.8480	-89.3225	0.142	0.114
22:26:52	s11	5976.26				
22:35:20	s12	5976.357	36.8483	-89.3265	0.125	0.279
22:38:28	s12	5976.367				
22:47:25	s13	5976.439	36.8483	-89.3291	0.144	0.178
22:49:26	s13	5976.438				
23:03:15	s2_2	5976.007	36.8588	-89.2940	0.104	0.1016
23:05:22	s2_2	5976.009				
23:13:44	b2_4	5975.911	36.8486	-89.3081	0.144	
23:16:06	b2_4	5975.909				

APPENDIX B: GRAVITY DATA PROCESSING

This appendix will go into detail about the data processing steps taken to produce the microgravity anomaly data used in this study. Since the CG-5 gravimeter measured gravitational acceleration to seven significant figures, at least seven significant figures were maintained through all calculations.

Appendix B1: Instrumental Drift Correction

Instrumental drift effects were removed from these data. During data collection the base station was returned to periodically where a measurement was taken and recorded. Since the gravimeter could not be placed at the exact same height during each base station reoccupation, this effect of its height was removed from all base station readings prior to the following steps. This correction calculation is discussed further below. Otherwise, any changes in the reading at the base station could be used to determine an average instrumental drift factor of the gravimeter between base station observations since this location was exactly the same each time and since lunar and solar tidal forces were accounted for by the gravimeter.

Excel was used to calculate linear equations relating the change in observed gravity at the base station with respect to time. We did not assume a constant linear drift for the entire survey, which is why a different equation was used to account for the effects of instrumental drift between each base station reoccupation. For example, in GP1, the first and second base station measurements (averaged and corrected for gravimeter height) were 5979.380

mGal and 5979.391 mGal, respectively. The duration between these two measurements was (in decimal format) 0.05851852 days, roughly equal to about 1 hr and 24 min. Decimal format was used for simplification of unit conversion. Dividing the change in base station gravity (Δg_B) by the change in time (Δt_B) between base station measurements yields the instrumental drift correction factor used to adjust observation stations values. For example, between the first and second base station measurements the instrumental drift correction factor was calculated as follows:

$$\frac{\Delta g_B}{\Delta t_B} = \frac{0.0116172 \text{ mGal}}{0.05851852 \text{ days}} = 0.19852177 \text{ mGal/day}.$$

GP1 stations 0 through 3 were measured between the first and second base station measurements. Thus, the duration between the observation station measurement time (t_{Sx}) and base station measurement time (t_{B0}) was calculated. Station 0 duration calculation is used as an example:

$$\Delta t = t_{S0} - t_{B0} = 0.7383681 \text{ days} - 0.7231308 \text{ days} = 0.0152373 \text{ days}.$$

Δt was multiplied by the correction factor determined above:

$$\Delta t \times \frac{\Delta g}{\Delta t} = 0.0152373 \text{ days} \times \frac{0.0116172 \text{ mGal}}{0.05851852 \text{ days}} = 0.003024930 \text{ mGal}.$$

Station 0 gravity measurement was reduced by the resulting value. This procedure was repeated for all stations in GP1 and GP2 using their respective base station measurements, time values and instrumental drift correction factors.

Appendix B2: Latitude Correction

To account for the force experienced by the gravimeter as a result of the rotation of the earth (i.e., centrifugal force in the rotating frame of reference), we use the equation given in Hinze et. al. (2013):

$$\Delta g_{CF} = .0008144 \frac{\text{mGal}}{\text{meter}} \times \sin 2\theta \times \Delta x_{NS},$$

where θ is the latitude in decimal degrees of the first point of observation (station 0) and Δx_{NS} is the change in distance in the north-south direction in meters between station 0 and subsequent observation stations. Δg_{CF} calculation between GP1 station 0 and station 1 is used as an example where θ of station 0 was 36.85486° and the distance between the two stations was 135.1175 m in the north south direction:

$$\Delta g_{CF} = .0008144 \frac{\text{mGal}}{\text{m}} \times \sin(2 \times 36.85486^\circ) \times 135.1175 \text{ m} = 0.1056383 \text{ mGal}.$$

The maximum centrifugal force felt by an observer occurs at the equator outward from the earth and decreases as an observer moves toward the north or south pole. Thus, the centrifugal force correction value at each station was subtracted from the observed gravity at that station to correct our gravity measurements.

Appendix B3: Elevation Correction

To account for the influence of elevation change at the surface on gravity readings, all measurements were corrected to adjust their values as if they were taken on a gravitationally equipotential surface. Zero elevation was set to the surface of the reference geoid. Three factors at each station were accounted for:

change in surface elevation w.r.t. the geoid, relative elevation change of subsequent stations and height of the gravimeter at that observation station. The sum of these quantities in meters is the h in the following equation from Hinze et. al. (2013):

$$\Delta g_h = \frac{2g}{R} \times h,$$

where Δg_h is the magnitude of adjustment that must be made to the measured gravity at a station in mGal, g is the gravitational acceleration of Earth in $\frac{m}{s^2}$, and R is the radius of Earth in meters. This equation simplifies to:

$$\Delta g_h = .3086 \frac{\text{mGal}}{\text{meter}} \times h.$$

For example, at GP1 station 1 the gravimeter was .150 m above the surface.

This value was added to the elevation of the station 0 w.r.t. to the geoid: 94 m.

And finally the change in surface elevation between station 1 and station 0:

0.076 m, was included in the sum:

$$h = .150 \text{ m} + 94 \text{ m} + .076 \text{ m} = 94.226 \text{ m}.$$

Substituting this value for h in the equation above yields:

$$\Delta g_h = .3086 \frac{\text{mGal}}{\text{meter}} \times 94.226 \text{ m} = 29.07814 \text{ mGal}.$$

Since the elevation of the gravimeter at all stations was greater than that of the reference geoid, the magnitude of the gravitational field would be less due to the inverse square of the distance factor inherent in the formula for gravity (Newton, 1686). Thus, the calculated Δg_h was added to the values of our survey stations, respectively.

Appendix B4: Mass Correction

To remove the influence of the mass between our measurement and the reference geoid the Bouguer Slab equation from Hinze et. al. (2013) was implemented:

$$\Delta g_{BA} = 2\pi G\sigma h,$$

where G is the gravitational constant in $\frac{\text{m}^3}{\text{s}^2\text{kg}}$, σ is the estimated average density of subsurface material in $\frac{\text{kg}}{\text{m}^3}$ and h is the change in height from the reference geoid in meters. The density used for all stations was estimated at $1500 \frac{\text{kg}}{\text{m}^3}$. For example, at station 3 the elevation w.r.t. to the geoid was 94.137 m:

$$\Delta g_{BA} = 0.00004193 \frac{\text{m}^3}{\text{s}^2\text{kg}} \times 1500 \frac{\text{kg}}{\text{m}^3} \times 94.137 \text{ m} = 5.920747 \text{ mGal}.$$

Positive extraneous mass between the geoid and the gravimeter would inconsistently increase a measurement of gravity. Thus, for our measurements, all made at an elevation higher than the geoid, Δg_{BA} was subtracted from the measured gravity.

Appendix B5: Terrain Correction

Lastly, to correct our data and list them as Complete Bouguer Anomalies (Δg_{CBA}), a terrain correction was necessary. This correction takes into account changes in elevation of the surface surrounding the gravity stations and how the surrounding mass influences measurements. Our profiles were conducted on flat roads that were slightly elevated compared to the surrounding flat farmland. However, between the roads and the fields there was a shallow ditch on both

sides. Overall, the change in elevation between the surface of the road and the fields was about one meter. The ditch between them dipped less than half a meter below the surface of the fields. The depth and width of the ditch did change; this was investigated to determine if a terrain correction was necessary. For example, if the ditch deepened along a profile, there would be less mass below and immediately adjacent to stations at the end of the profile. Therefore, the gravimeter would progressively record lower microgravity values along a profile. The same would be true for a profile adjacent to progressively widening ditch.

Using terrain effect estimation techniques in Hubbert (1948), it was determined that the variation in the gravitation field due to the adjacent ditches was negligible. Thus, the described corrections performed on these data constitute a Δg_{CBA} for all practical purposes.

Appendix B6: Microgravity Profile Error Calculation

This section will describe how the error values of each microgravity profile were calculated. In order to perform this calculation, reoccupation and measurement of a station within each profile, in addition to base station reoccupation and measurement, was necessary. For GP1, this station was station 3. In addition to measurement after station 2, station 3 was reoccupied and measured again after station 13 but prior to the final base station measurement.

The first step in this calculation was to remove the gravitational variation due to the change in height of the gravimeter, w.r.t. the surface, since the

gravimeter could not be placed at exactly the same height during s_{3_1} and s_{3_2} measurements. Next, instrumental drift corrections (see sec. 4.2.1) were applied to the s_{3_2} microgravity values. Since between s_{3_1} and s_{3_2} measurements multiple base station reoccupations were performed, it was necessary to apply each drift correction factor accordingly. Doing so removed all instrumental drift that had occurred in the duration between s_{3_1} and s_{3_2} measurements.

Theoretically, after these two correction factors: gravimeter height and instrumental drift had been applied to the s_{3_2} measurements, the values of s_{3_1} and s_{3_2} should be equal. Therefore, any difference between the two constitutes the error of a survey. GP1 error was calculated to be $\pm .004$ mGal. GP2 error was calculated to be ± 0.015 mGal.

APPENDIX C: AUGMENTING MODELS WITH PRYNE (2012) DATA

As described in 2.2.2, SP1 was not intended to resolve the shallow subsurface (0 to ~0.1 km depth). In an effort to take this section of the subsurface into account in microgravity analysis, we contacted Daniel E. Pryne, the author of Pryne (2012) and co-author of Pryne et al. (2013). The lithology models presented in these studies included data pertaining to 0 to ~0.1 km depth. For the purpose of augmenting this study's models, Pryne sent the lithological data that were used to hypothesize the existence of the CU. These data were loaded into ArcGIS software and were plotted in NAD_1983_UTM_Zone_15N to correlate these data with microgravity observation stations in this study. Data from Pryne (2012) consisted of 31 polygon gradient maps, each representing the observed lithology at specific depths, in 3 m depth sections. The lithology class at each station in GP1 was evaluated and recorded. This was repeated for each 3 m depth section. These data were used to create Table C1.

Table C1: This table shows the uninterpreted lithology profile along GP1, derived from the from Pryne (2012) model.

depth	GP1 observation stations												
to top (m)	12	11	10	9	8	7	6	5	4	3	2	1	0
0.00	5	5	5	5	4	4	4	4	4	4	4	4	4
-3.05	4	4	4	4	5	5	5	5	4	4	4	4	4
-6.10	3	3	3	3	3	3	3	3	2	2	2	2	2
-9.14	1	1	1	1	1	1	1	1	1	1	1	1	1
-12.19	1	1	1	1	1	1	1	1	1	1	1	1	1
-15.24	1	1	1	1	1	1	1	1	1	1	1	1	1
-18.29	1	1	1	1	1	1	1	1	1	1	1	1	1
-21.34	1	1	1	1	0	1	0	0	0	0	0	0	0
-24.38	1	1	1	1	1	1	1	1	1	1	1	1	1
-27.43													
-30.48													
-33.53	0	0	0	0	0								
-36.58	1	0	0	0	0	0							0
-39.62	1	1	1	0	0	0	0						0
-42.67	1	1	1	0	0	0	0						
-45.72	1	1	1	0	0	0	0						
-48.77	1	1	1	0	0	0	0						
-51.82	1	1	1	0	0	0	0						
-54.86	3	3	3	3	3	3	3	3	3	2	2	2	2
-57.91	5	4	4	4	4	4	4	3	3	3	3	3	3
-60.96	5	4	4	4	4	4	4	3	3	3	3	3	3
-64.01	5	5	5	5	5	5	5	5	5	5	5	5	5
-67.06	5	5	5	5	5	5	5	5	5	5	5	5	5
-70.10	5	5	5	5	5	5	5	5	5	5	5	5	5
-73.15	5	5	5	5	5	5	5	5	5	5	5	4	4
-76.20	4	5	5	5	5	5	5	4	4	4	4	4	4
-79.25	4	4	5	5	5	5	4	4	4	4	4	4	4
-82.30	4	4	5	5	5	5	5	4	4	4	4	4	4
-85.34	4	4	4	4	4	4	4	4	4	4	4	3	3
-88.39	4	5	5	5	5	5	4	4	4	4	4	3	3
-91.44													
-94.49													
KEY	Lithology												
	0	Gravel, low gamma and high resistivity values											
	1	Sand, low gamma and high resistivity values											
	2	interval containing equal parts of values 1 and 3											
	3	Silt, intermediate gamma & intermediate-high resistivity value											
	4	An interval containing equal parts of values 3 and 5											
	5	Clay, high gamma & low resistivity values											
		Null value assigned to missing or damaged data											

Between 27 and 52 m there was a considerable amount of missing data. Modeling the subsurface as shown in Table C1 would have created erroneous lateral density discontinuities and the calculated microgravity anomaly would be meaningless. At each station in GP1, between 27 and 52 m, where lithology data were missing the lithology surrounding the missing data points was sand. Therefore, it was assumed that the null values in Table C1 between 27 and 52 m consisted of the sand lithology. This assumption is reasonable considering that 9 to 52 m depth correlates to the Upland Complex in the stratigraphic column in Figure 3 and that the Upland Complex consists of fine- to very coarse-sand and gravel (Crone, 1981). Other sections of missing data from Pryne (2012) were laterally continuous across the profile. Therefore, these depth sections were assumed to consist of the same sediment used to model Quaternary soil in prior models.

Density values were assigned by considering a multitude of sources. First local water table depths were evaluated. The USGS reports that the depth to the water table in Mississippi County has fluctuated between 1 and 4 m in the past 5 years (Figure 38) This is of particular relevance since clays within the Mississippi River floodplain consist mostly of vertisols (Aslan and Autin, 1998). Vertisols, such as montmorillonite, are defined by their ability to absorb water and expand. Osipov (2011) shows that the density of this type of clay, when hydrated, is 1.40 g/cc. This density was assigned to the clay lithology identified by Pryne (2012) and modeled as such.

Figure C1: This graph plots the past 5 years of water table depth in feet w.r.t. the surface as recorded at the USGS East Prairie recording station in Mississippi County, Missouri.

The sand lithology was estimated to be 1.61 g/cc. This value was determined by averaging density values from two different sources. The first: 1.54 g/cc, is an average density of various well graded sands (Mehdiratta and Triandafilidis, 1978). The second value considered for sand lithology was estimated in an American Society for Testing and Materials (ASTM) study (Lo Presti et al., 1992). Mixtures of sand and silt particles at varying proportions were used to create a linear regression, which enabled calculation of a purely sand mixture density: 1.68 g/cc. Additionally, their 50/50 sand/silt mixture matches the description of the sand/silt lithology in Pryne (2012). The correlation justified the use of ASTM's 50/50 sand-silt soil density measurements (1.77 g/cc average) for the purposes of modeling. This estimate is supported by experimental

observations in Bodman and Constantin (1965). The same ASTM study reported an average density value of 1.99 g/cc for well graded gravel. This density is used in modeling of gravel lithologies.

For silt lithologies the density was estimated and modeled at 1.41 g/cc. Bodman and Constantin (1965) experimentally assessed the densities of soils with varying hydration levels. Since all of the recorded silt lithologies along GP1 intersected the minimum water table depth, the modeled silt was assigned the density value of a saturated silt soil reported in Bodman and Constantin (1965). The remaining lithology, “an interval containing equal parts of [clay] and [silt]” (Pryne, 2012) was estimated to have a density equal to the average of clay and silt lithology densities. The interpreted subsurface lithologies and corresponding densities are shown in Table C2.

Table C2: This table shows the interpreted lithology profile along GP1.

depth to top (m)	Observation stations in GP1												
	12	11	10	9	8	7	6	5	4	3	2	1	0
0.00	5	5	5	5	4	4	4	4	4	4	4	4	4
-3.05	4	4	4	4	5	5	5	5	4	4	4	4	4
-6.10	3	3	3	3	3	3	3	3	2	2	2	2	2
-9.14	1	1	1	1	1	1	1	1	1	1	1	1	1
-12.19	1	1	1	1	1	1	1	1	1	1	1	1	1
-15.24	1	1	1	1	1	1	1	1	1	1	1	1	1
-18.29	1	1	1	1	1	1	1	1	1	1	1	1	1
-21.34	1	1	1	1	0	1	0	0	0	0	0	0	0
-24.38	1	1	1	1	1	1	1	1	1	1	1	1	1
-27.43													
-30.48													
-33.53	0	0	0	0	0	1	1	1	1	1	1	1	1
-36.58	1	0	0	0	0	0	1	1	1	1	1	1	0
-39.62	1	1	1	0	0	0	0	1	1	1	1	1	0
-42.67	1	1	1	0	0	0	0	1	1	1	1	1	1
-45.72	1	1	1	0	0	0	0	1	1	1	1	1	1
-48.77	1	1	1	0	0	0	0	1	1	1	1	1	1
-51.82	1	1	1	0	0	0	0	1	1	1	1	1	1
-54.86	3	3	3	3	3	3	3	3	3	2	2	2	2
-57.91	5	4	4	4	4	4	4	3	3	3	3	3	3
-60.96	5	4	4	4	4	4	4	3	3	3	3	3	3
-64.01	5	5	5	5	5	5	5	5	5	5	5	5	5
-67.06	5	5	5	5	5	5	5	5	5	5	5	5	5
-70.10	5	5	5	5	5	5	5	5	5	5	5	5	5
-73.15	5	5	5	5	5	5	5	5	5	5	5	4	4
-76.20	4	5	5	5	5	5	5	4	4	4	4	4	4
-79.25	4	4	5	5	5	5	4	4	4	4	4	4	4
-82.30	4	4	5	5	5	5	5	4	4	4	4	4	4
-85.34	4	4	4	4	4	4	4	4	4	4	4	3	3
-88.39	4	5	5	5	5	5	4	4	4	4	4	3	3
-91.44													
-94.49													
	KEY												
ρ (g/cc)		Lithology											
1.99	0	Gravel, low gamma and high resistivity values											
1.61	1	Sand, low gamma and high resistivity values											
1.77	2	interval containing equal parts of values 1 and 3											
1.43	3	Silt, intermediate gamma & intermediate-high resistivity value											
1.42	4	An interval containing equal parts of values 3 and 5											
1.40	5	Clay, high gamma & low resistivity values											
1.70		Quaternary soil identified in Bexfield et al. (2006)											

APPENDIX D: SP1 IN DETAIL

Figure D1: Through multiple iterations of processing SP1 data and based on comments from reviewers, we suggest there are at least three faults captured by SP1, shown with white lines in this figure. The red shaded section indicates the main CU-s shear zone which divides the uplift and the adjacent down-dropped side to the south-east. The white lines on the far left and right of the image represent antithetic faults we have identified that we believe to be associated with the CU-s. The green shade indicates the extent of GP1 that had the maximum slope of that profile. This interpretation of the CU-s, in red, correlates with the slope analysis prediction. Since all of our gravity models only accounted for one fault, this more detailed interpretation was not implemented. However, incorporating all three faults in a similar model may yield a better correlation between the calculated anomaly and that of GP1.

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VITA

Drew David Burford Jr.

Education:

Rhodes College B.S. in Physics: 2013-2017

Professional Experience:

Teaching assistant in University of Kentucky EES Department:
2017-2019

Research assistant in Rhodes College Exoplanetary Imaging Lab:
2016

Honors:

Sigma Gamma Epsilon member: 2019

Southern Athletic Association's Fall Academic Honor Roll: 2013