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Interpretation of Crustal Structure from Regional Gravity Anomalies, Ouachita Mountains Area and Adjacent Gulf Coastal Plain[1]

J. M. KRUGER[2] and G. R. KELLER[3]

ABSTRACT

A gravity data base from more than 35,000 stations was used to generate a series of regional gravity maps of the Ouachita Mountains area including adjacent parts of the craton and the Gulf coastal plain. These maps were used in conjunction with information from 96 wells, data from preexisting geophysical and geological investigations, and computer models to interpret four gravity profiles that transect the study area (approximately lat. 30-37°N, long. 91.5-99°W). These models, gravity maps, and previous investigations were then used to analyze various regional gravity anomalies and to interpret the gross crustal structure of the region and its tectonic implications.

These data suggest that variably attenuated continental crust lies beneath the Gulf coastal plain, south of the Ouachita system gravity gradient, as opposed to "typical" continental crust of the craton north of this gradient. This variation in crustal structure probably reflects the complexity of Eocambrian and early Mesozoic rifting in the area. The Arkoma basin gravity minima may result from the combined effect of a late Paleozoic foreland basin and an Eocambrian northwest-trending, rift-related basin. The Ouachita system interior zone gravity maximum varies along strike of this orogenic belt. This anomaly appears to be a good indicator of the position of the Eocambrian continental margin and associated rift zone. Gravity anomalies in the Gulf coastal plain appear to be a combined effect of variable crustal attenuation, basins and uplifts, and mafic intrusions. Gravity maxima in the southern Oklahoma aulacogen result from uplifts and deep-seated mafic intrusions; gravity minima result from deep sedimentary basins.

INTRODUCTION

The Ouachita system (Figure 1) is a major late Paleozoic orogenic belt that extends more than 1,500 km from Mississippi to northern Mexico. Flawn et al (1961) provided a useful and still definitive summary of the geology of this feature. The Ouachita Mountains of Arkansas and Oklahoma, and the Marathon basin region of west Texas are the only two significant exposures of the Ouachita system, although many drill holes in Arkansas, Oklahoma, and Texas have encountered Ouachita facies rocks (e.g., Flawn et al, 1961; Nicholas and Rozendal, 1975; Denison et al, 1977).

Most of the Ouachita system is buried; therefore, much controversy exists as to its origin. Early plate tectonic models for its development (Walper and Rowett, 1972; Keller and Cebull, 1973) were simplistic, but they recognized that the tectonic histories of the Gulf of Mexico, Gulf coastal plain, and Ouachita system were intertwined.

Since these early models were introduced, a variety of more complex models has been proposed. It is beyond the scope of this study to review all of these models (see Pindell and Dewey, 1982; Pindell, 1985, for recent reviews); however, we have summarized the points regarding plate interactions that are of major interest in this study.

First, most workers agree that a major continental breakup occurred during the Eocambrian (late Precambrian and Early Cambrian). This rifting event established the structural framework for much of the south-central and southeastern United States (e.g., Keller et al, 1983). Passive continental margins formed along the resulting rifted continental margin during the early Paleozoic, and subsidence occurred in many interior basins, particularly along failed rift zones that extended into the continent from its margins.

Subduction zones signaling ocean closings began to form during the early to middle Paleozoic, but the type, location, number, and polarity of the subduction zone–volcanic arc complexes are controversial. Most of the areas where critical observations need to be made are deeply buried, particularly along the Ouachita orogenic belt. Thus, workers can interpret the available data in several ways, but most believe any middle to late Paleozoic subduction zone adjacent to the Ouachita orogenic system dipped to the south.

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[2]Department of Geological Sciences, University of Texas at El Paso, El Paso, Texas 79968. Present address: Marathon Oil Company, P.O. Box 120, Casper, Wyoming 82602.

[3]Department of Geological Sciences, University of Texas at El Paso, El Paso, Texas 79968.

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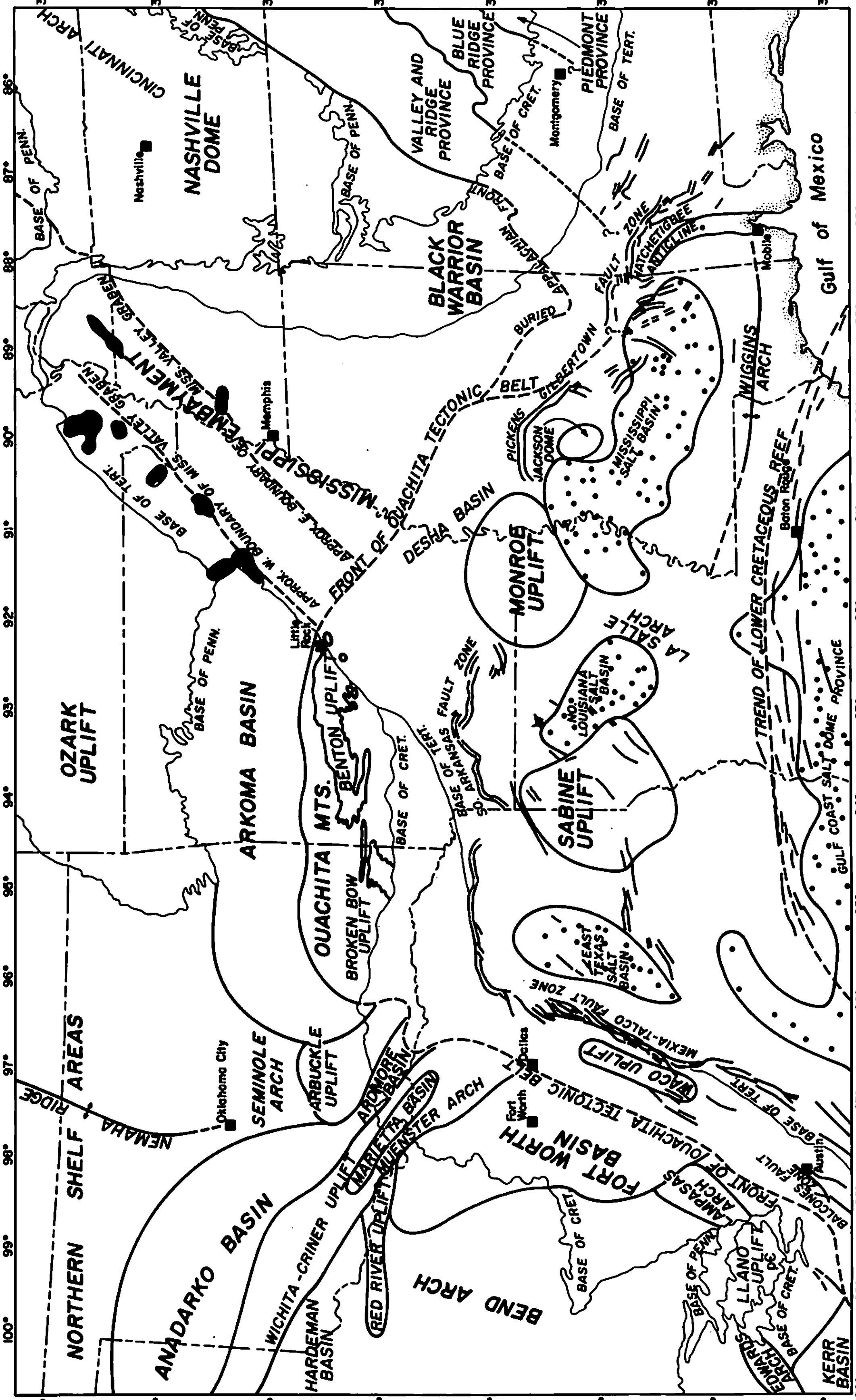

Figure 1—Index map of study area showing major tectonic features (modified from Miser, 1959; Cohee, 1962; Martin and Case, 1975; Nicholas and Rozendal, 1975; Walper et al, 1979; Hildenbrand et al, 1982). Black dots indicate salt domes; dark pattern indicates mafic to ultramafic intrusions in Mississippi Embayment.

The major phase of orogenic activity began during the late Paleozoic and seemed to migrate westward along the Ouachita system, and ended with early Permian deformation in the Marathon region of west Texas (Flawn et al, 1961; King, 1977). During this time, South America was in close proximity to North America (e.g., Van der Voo et al, 1976), and continent-continent collision was probably at least partly responsible for the Ouachita orogeny. However, owing to the irregular shape of the late Paleozoic continental margin of North America and the need to account for blocks such as Yucatan (Pindell and Dewey, 1982; Pindell, 1985), any scenario of collision must be complicated and possibly incomplete. Microcontinents also may have influenced this orogeny (Thomas, 1976, 1977).

The last major phase of tectonic activity was rifting that began during the Triassic as the Gulf of Mexico began to reopen. This phase of activity is documented by the presence of Triassic red beds deposited in grabens, and associated Triassic intrusions and volcanics (Vernon, 1971; Woods and Addington, 1973; Jackson and Seni, 1983). These features are generally analogous to the Triassic grabens exposed along the Atlantic Coast. This rifting event is a major cause of ambiguity in interpretations of geophysical data because it is difficult to determine if deeply buried features are the result of Eocambrian rifting, late Paleozoic orogenic activity, or Mesozoic rifting. This is particularly true of measurements of deep crustal structure because stretching models for rifting (McKenzie, 1978) predict transitional crustal structure whose continental affinities may be difficult to recognize.

This study uses gravity, drilling, and geologic data in an integrated analysis to resolve some questions regarding Ouachita system development, particularly in the area of the Ouachita Mountains (Figure 1). We compiled a large data base of gravity readings for this purpose (Figure 2), and where possible, our analysis relied heavily on drilling and geologic data to constrain gravity interpretations.

Previous geophysical studies in the area are limited but also provided valuable constraints on our interpretations. These studies are discussed in the context of their implications for the interpretations of individual features.

GRAVITY MAPS AND COMPUTER MODELS

The gravity data used in this study are part of a large data base maintained at the University of Texas at El Paso. In the study area, many of these data were provided by H. Brown (personal communication). Additional sources of data were the Defense Mapping Agency, the University of Texas at Dallas, the National Geodetic Survey, and the U.S. Geological Survey. This data base is edited and maintained in a standard format. It is also tied to a common gravity datum (IGSN-71; Morelli, 1976). Sea level was used as an elevation datum, and a density of 2.67 g/cm^3 was used in Bouguer corrections. All gravity data were reduced to Bouguer anomaly values using the 1967 formula for theoretical gravity (Morelli, 1976) and the reduction formulas of Cordell et al (1982).

Approximately 35,000 gravity readings were available for this study (see Figure 2, distribution). To enhance various anomalies and trends, we applied a variety of digital processing techniques to these data. First, we compiled a grid of the simple Bouguer anomaly values at a 4.0-km (2.5-mi) interval using the minimum curvature technique (Briggs, 1974). These values were contoured using a modified version of the Surface II graphics system (Sampson, 1978) to produce the Bouguer anomaly map shown as Figure 3.

The regional gravity field was modeled by fitting a second-order polynomial surface to the Bouguer anomaly values using the Lance (1982) technique. This surface (Figure 4) represents the application of a very low-pass filter to the data and generally depicts a decrease in gravity values from the Gulf Coast toward the craton. The difference between the second-order polynomial and the simple Bouguer anomaly map represents a residual (high-pass filtered) anomaly map (Figure 5; Figure 6, same map with tectonic features shown). This map provides better definition of many anomalies under consideration, because the interfering regional trend has been removed.

The trends of various anomalies intersect; therefore, we separated trends by strike filtering (i.e., filtering in which anomalies with a certain trend can be selectively attenuated). Figures 7 and 8 depict strike-filtered maps constructed using a two-dimensional Fourier transform–based technique (Coultrip, 1982). To enhance anomalies not associated with the dominant northwest-trending southern Oklahoma aulacogen, northwest-trending anomalies were rejected by strike filtering to produce Figure 7. The parameters used in this filter rejected linear trends from N28°W to N68°W in the spatial domain. The strike-filtered map in Figure 8 was developed using a filter designed to reject trends associated with the Ouachita system. This filter rejected linear trends from N0°E to N85°E in the spatial domain, and was particularly useful in analyzing the extent of structures associated with the southern Oklahoma aulacogen.

The anomalies in the study area were well suited for two-dimensional computer modeling. Thus, four gravity profiles (AA′ through DD′, Figure 3) were constructed from the Bouguer anomalies at gravity stations that were projected, parallel to the contour lines of the Bouguer anomaly map in Figure 3, onto the profiles. None of the stations were projected more than 7 km (4.4 mi), and most of them were projected much less.

Except for the part of profile AA′ that follows the COCORP seismic reflection line in Arkansas (Lillie et al, 1983) where detailed gravity data were gathered, the data base used for the profiles is shown in Figure 2. In each profile, the gravity value of each station was plotted against its projected distance (in kilometers) from the origin of the profile. Except for stations along the COCORP line, no terrain corrections were made. The position of stations along the COCORP line on profile AA′ was determined by projecting the station directly east or west to a north-south line through the area of the survey (dashed part of profile AA′ in Figure 3).

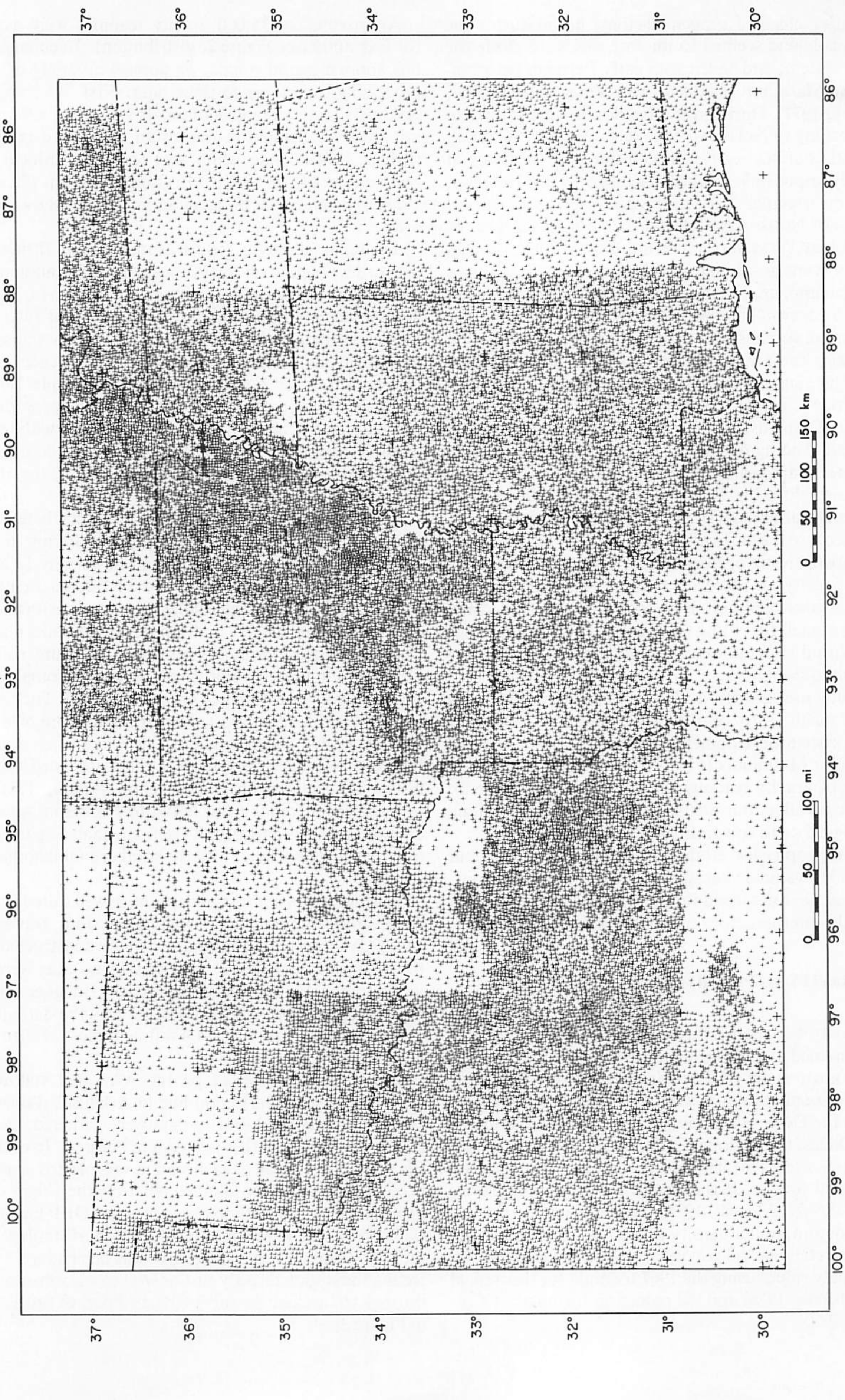

Figure 2—Index map showing distribution of gravity readings (+) used in this study.

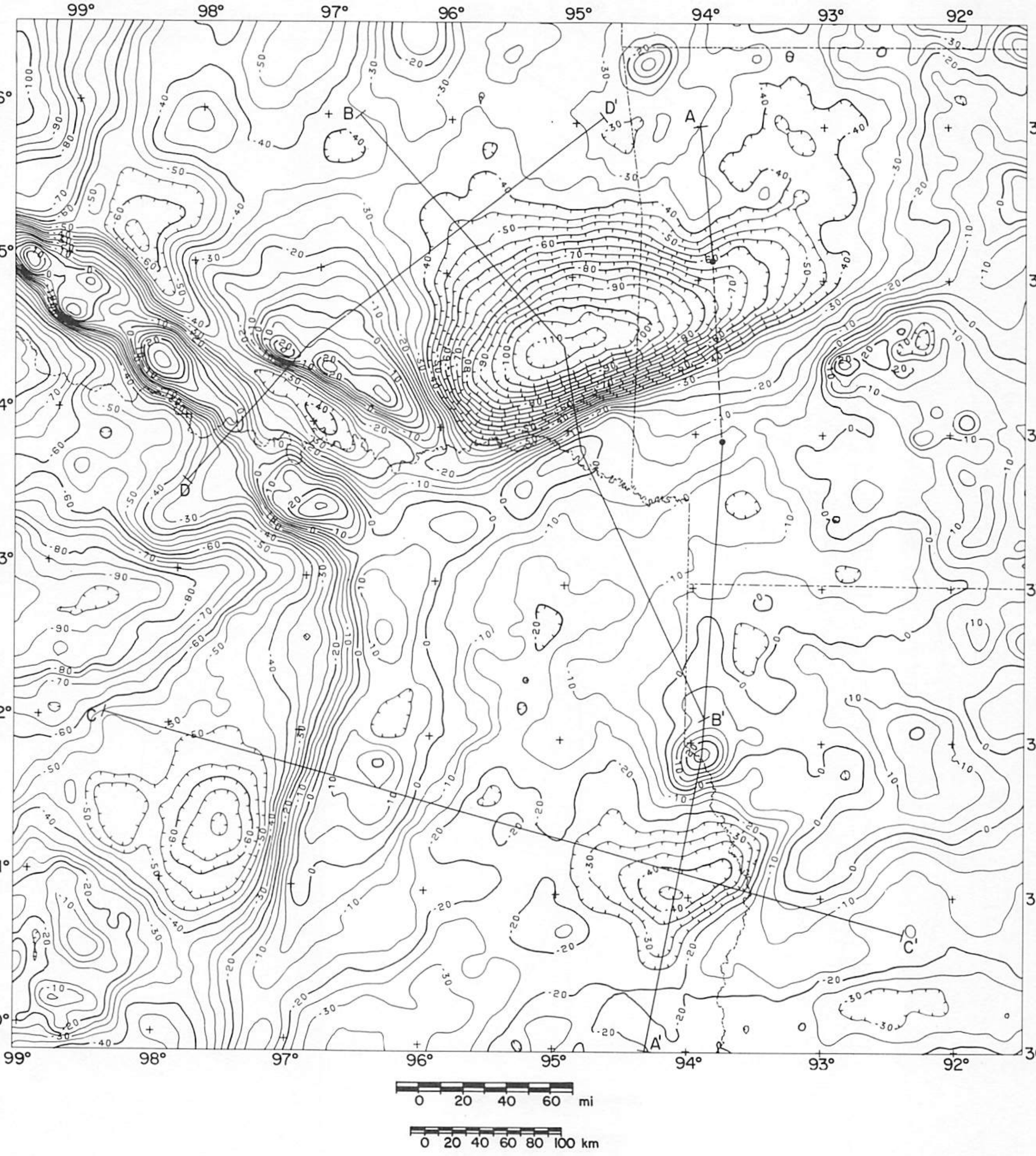

Figure 3—Bouguer gravity map of study area. C.I. = 5 mgal. Sea level datum. Reduction density = 2.67 g/cm^3. Locations of gravity profiles are also shown (AA′, BB′, CC′, DD′). Dashed line indicates approximate line on which COCORP line gravity stations are projected.

We modeled profiles AA′ through DD′ using the two-dimensional technique of Talwani et al (1959), and we used data from 96 wells (Figure 9) as constraints in the modeling process. The name, location, and total depth of each well, along with tops used to construct the starting models, are available in Kruger (1983). In profile AA′, the starting model for the upper crust, from distances of 98 to 229 km (61 to 143 mi) south of A, was based on results from the COCORP seismic profile in Arkansas (Nelson et al, 1982; Lillie et al, 1983). In profile DD′, the starting model for the upper part of the upper crust, from distances of 52 to 134 km (33 to 84 mi) northeast of D, was based on a cross section by Denison (1982). All models were constructed to extend from sea level to 60 km (37 mi) below sea level. The

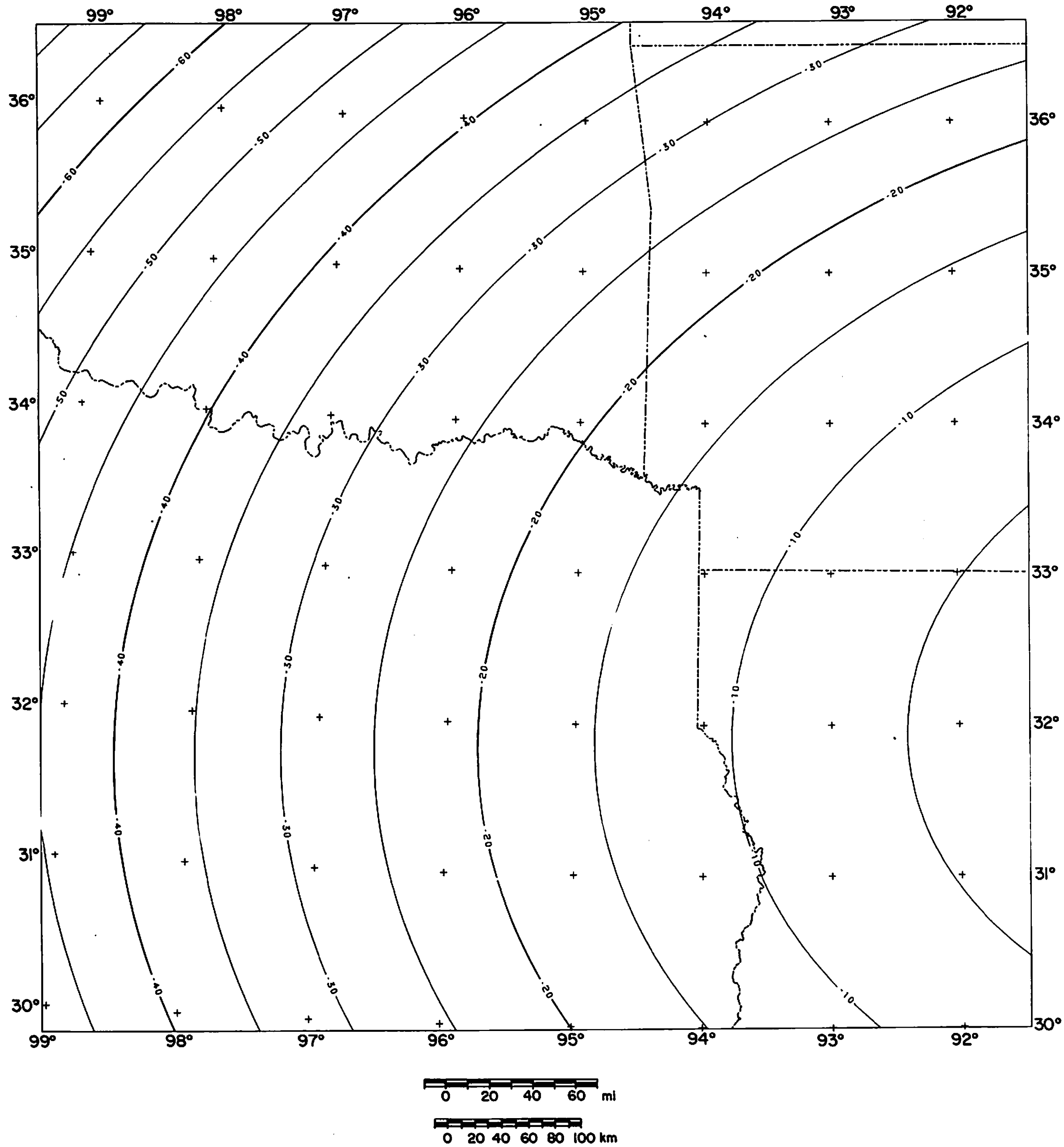

Figure 4—Contour map of second-order polynomial surface. C.I. = 5 mgal.

models derived for each profile, with the observed gravity values and theoretical values calculated from the models, are shown in Figures 10-13. Figure 14 contains the legend for Figures 10-13. For display purposes on these figures, we added a constant of 120 mgal to the observed gravity values of each station. This constant did not affect the results because relative anomaly values were actually modeled. In some cases, these models contain more geologic detail than required to satisfy the gravity data alone. However, our goal was to construct crustal-scale cross sections that reflected the integrated analysis of all available data. Therefore, these models represent geologic cross sections in which gravity data were used to extrapolate beneath and between points where seismic and drilling data were available.

The models of profiles AA′ through DD′ were constructed using seismic measurements within the craton in Missouri, Arkansas, and Oklahoma, to determine starting crustal structure models (McCamy and Meyer, 1966; Ste-

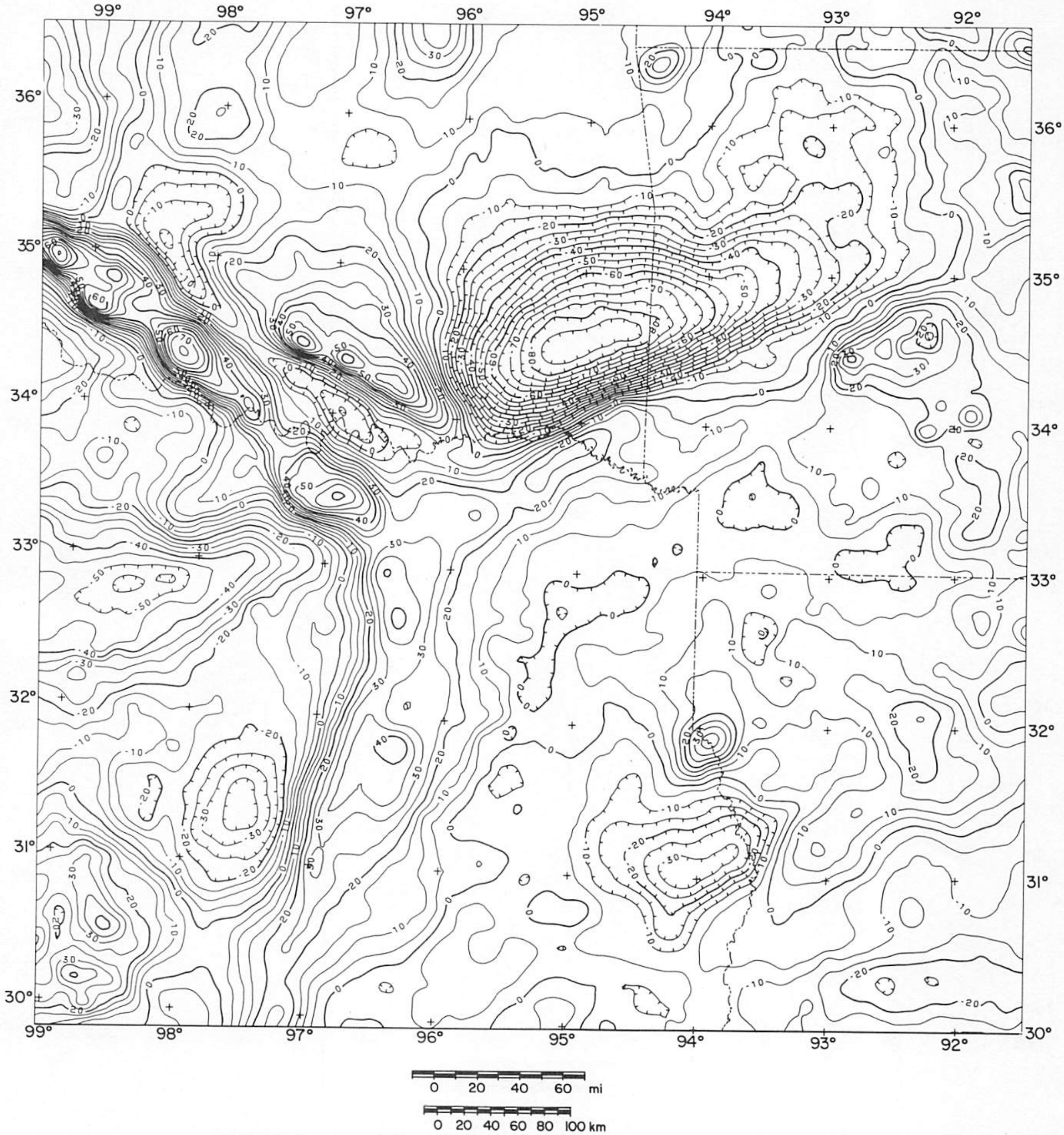

Figure 5—Contour map of residual gravity values with respect to surface shown in Figure 4 (i.e., values in Figure 5 = values in Figure 3 − values in Figure 4). C.I. = 5 mgal.

wart, 1968; Mitchell and Landisman, 1970). Because crustal thickness in these models ranged from 40 to 50 km (25 to 31 mi) and upper crustal thickness (P-wave velocities up to 6.3 km/sec) ranged from 15 to 30 km (9 to 19 mi), an average crustal thickness of 43 km (27 mi) and an average upper crustal thickness of 22 km (14 mi) were chosen (for modeling purposes) for the craton around the margins of the Ozark uplift (northern parts of profiles AA′ and DD′). These thicknesses were chosen in the starting models of all four profiles, but were modified in all but the northern parts of profiles AA′ and DD′.

Densities of 3.3 g/cm^3 and 3.0 g/cm^3 were used for the mantle and lower crust, respectively. A density of 2.7 g/cm^3 was used for the major part of the upper crust. For the cratonic parts of the profiles, rocks of the dense lower Paleozoic section (Ellenburger or Arbuckle formations and lower) were included in the upper crust. For the part of the profiles within and south of the Ouachita system, the upper

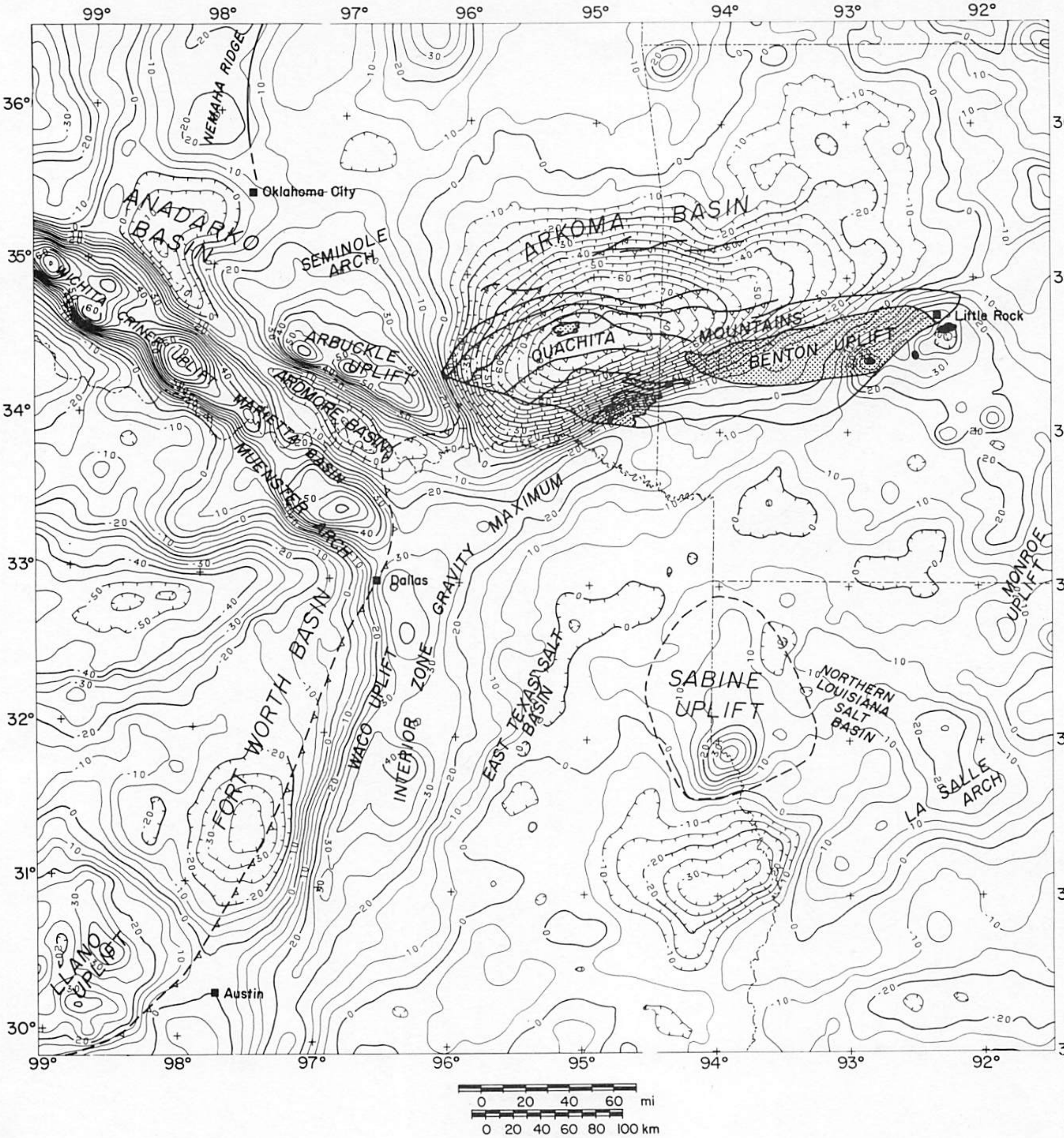

Figure 6—Residual gravity map of Figure 5 with major tectonic features labeled.

crust includes lower Mississippian and stratigraphically lower metasedimentary rocks. Without additional seismic and drilling control, determining the thickness of these units was beyond the resolution of the data. In the earth models derived from profiles AA′ through CC′, the rocks assigned a density of 2.58 g/cm^3 are comprised primarily of Mississippian and higher Paleozoic rocks (flysch and molasse facies). These rocks include a relatively thin sequence of post-Arbuckle and post-Ellenburger to Mississippian sedimentary rocks (including the Potato Hills uplift strata in BB′, Figure 11). This 2.58-g/cm^3 density is based on a measured density of 2.56 g/cm^3 for surface exposures of the Atoka and other Pennsylvanian formations in the Ouachita Mountains (Dobrin, 1976) and on the fact that density generally increases with depth due to compaction. However, this average density is probably a minimum for the sedimentary rocks in the Ouachita Mountains, Arkoma basin, and Fort Worth basin; therefore, the thicknesses of these rocks shown in Figures 10-12 are probably minimum estimates, as well.

In profile DD′, post-Arbuckle rocks are divided into a middle Paleozoic layer with a 2.6-g/cm^3 density, and an

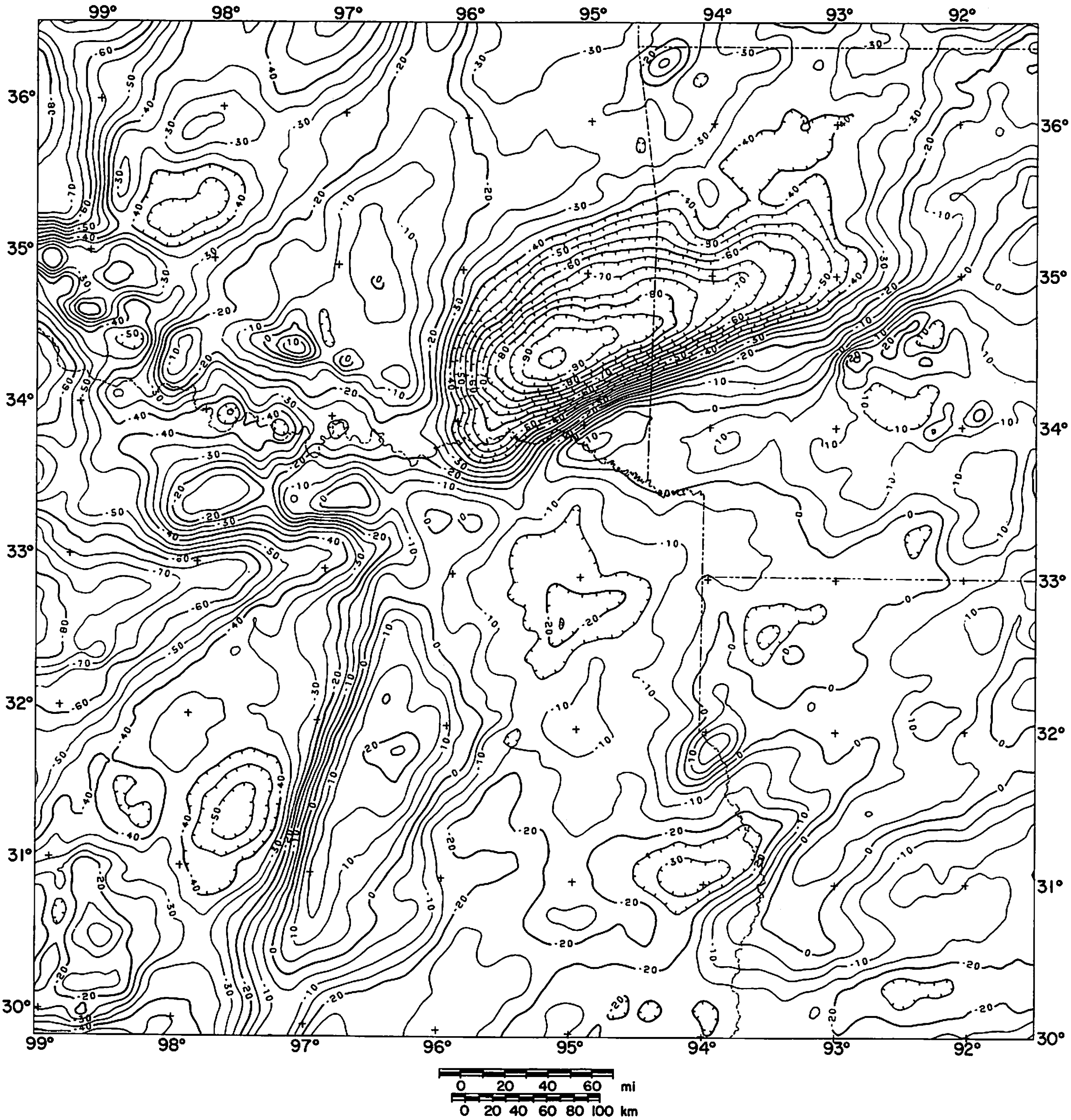

Figure 7—Strike-filtered gravity map in which trend of southern Oklahoma aulacogen (N28°W to N68°W) has been rejected. C.I. = 5 mgal.

upper Paleozoic and higher layer with a 2.5-g/cm^3 density. These density estimates are based on values reported by Pruatt (1975). In earth models shown as Figures 10-12, rocks with a 2.65-g/cm^3 density consist of Paleozoic metasedimentary and sedimentary rocks and, in places, Smackover and lower Mesozoic rocks. Rocks with a 2.5-g/cm^3 density consist of Mesozoic and Cenozoic sedimentary rocks. These density estimates are based on a depth-density function determined by Crosby (1971). Although the near-surface densities of rocks in the Gulf coastal plain are much less than 2.5 g/cm^3, compaction of the rocks with depth should increase the density to give an approximate average density of 2.5 g/cm^3 for the entire Mesozoic-Cenozoic section.

INTERPRETATION OF GRAVITY ANOMALIES

Regional Anomalies in the Coastal Plain

Average regional gravity values within the Gulf coastal plain (Figure 3), south and southeast of the Ouachita system, are greater than average regional gravity values in the

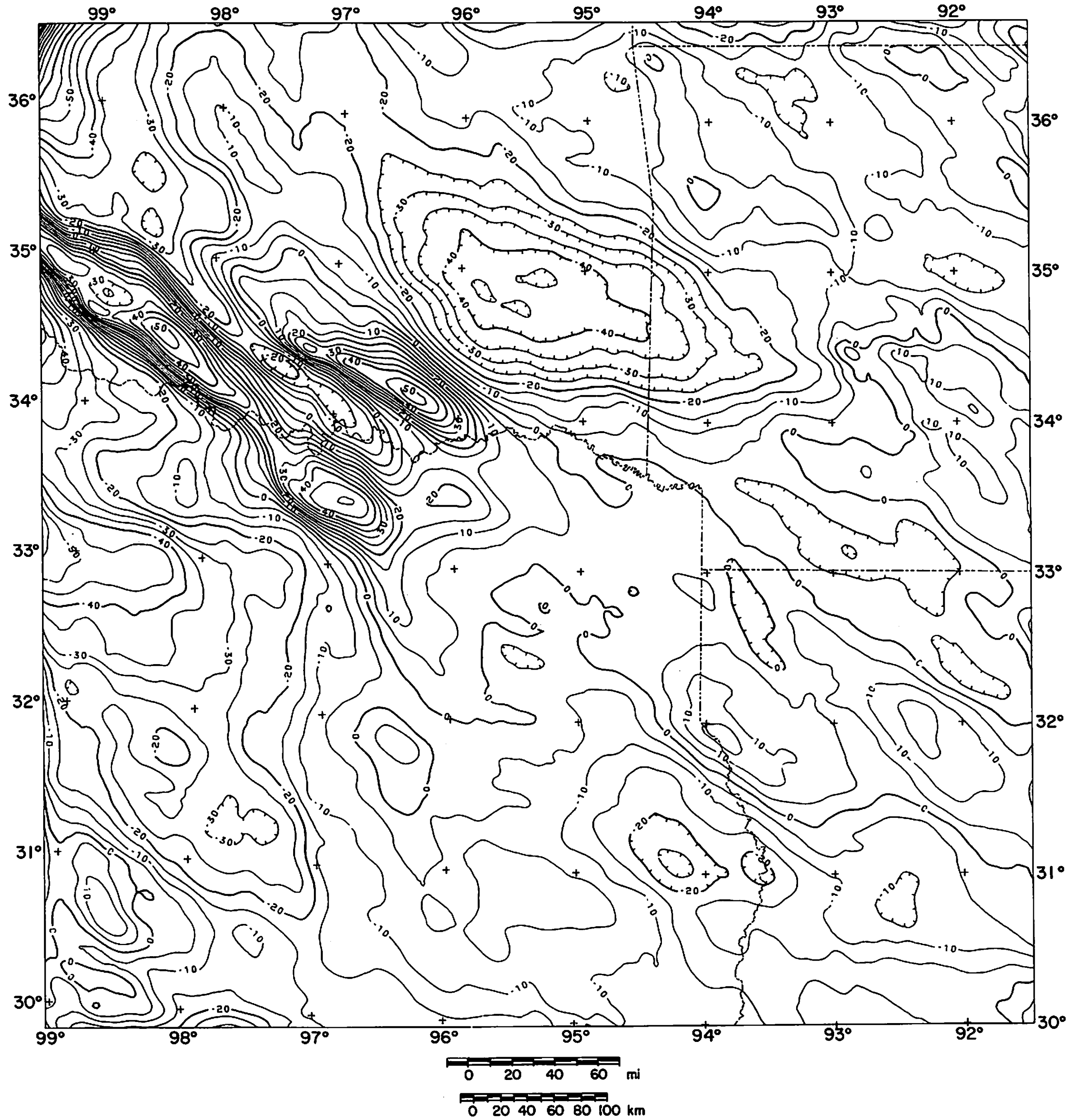

Figure 8—Strike-filtered gravity map in which trends associated with Ouachita system (N0° to N85°E) have been rejected. C.I. = 5 mgal.

craton, north and northwest of the Ouachita system. This variation also can be seen on profiles AA′ through CC′ (Figures 10-12). After the long wavelength (lithospheric?) effects of this gradient are removed by subtracting the second-order polynomial surface (Figure 4) from the Bouguer anomaly values, many anomalies of interest in this study are better defined (Figures 5, 6). Previous seismic measurements (Cram, 1962; Antoine and Ewing, 1963; Qualls, 1965; McCamy and Meyer, 1966; Warren et al, 1966; Stewart, 1968; Hales et al, 1970; Mitchell and Landisman, 1970; Dorman et al, 1972; Hales, 1973; Kurita, 1973; Keller and Shurbet, 1975) and gravity studies (Watkins, 1961; Fish, 1970; Keller and Cebull, 1973; Worzel and Watkins, 1973; Lillie et al, 1983; Ando et al, 1984) indicate that regional variation in gravity values can be attributed to a generally thinner and/or denser crust in the Gulf coastal plain relative to the craton.

In the models shown in Figures 10-12, the higher average

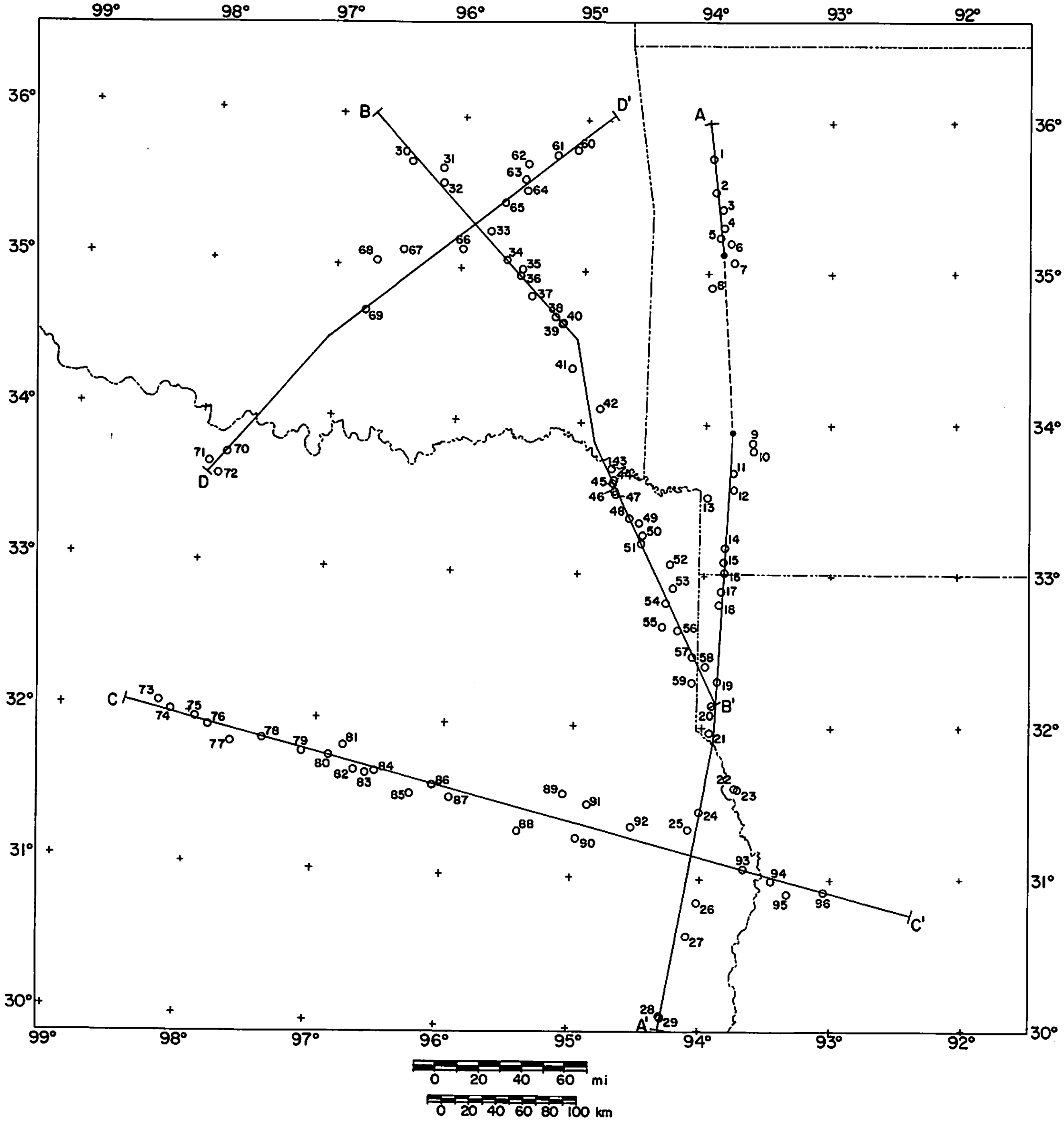

Figure 9—Location map for gravity profiles and wells (open circles with numbers) used in computer modeling. Dashed line is the approximate line on which COCORP line gravity stations were projected. Numbers correspond to well data available in Kruger (1983).

Bouguer values in the Gulf coastal plain have been modeled by thinning both the lower and the upper crust. However, due to the nonuniqueness of gravity modeling, these basic results could be achieved by thinning either the upper crust or the lower crust alone. We chose to thin both the upper and lower crusts in accordance with the limited crustal structure data in the area, results from other continental margins, and recent models of lithospheric extension (e.g., McKenzie, 1978). Models for crustal extension and passive margin subsidence, similar to that which occurred in the Gulf of Mexico and Gulf coastal plain, have been discussed by many authors (e.g., Steckler and Watts, 1978; Bott, 1979; Buffler et al, 1979; Grow et al, 1979; Sclater and Christie, 1980; Pitman, 1983).

Seismic and gravity studies of the central and western Texas Gulf coastal plain area (Cram, 1962; Dorman et al,

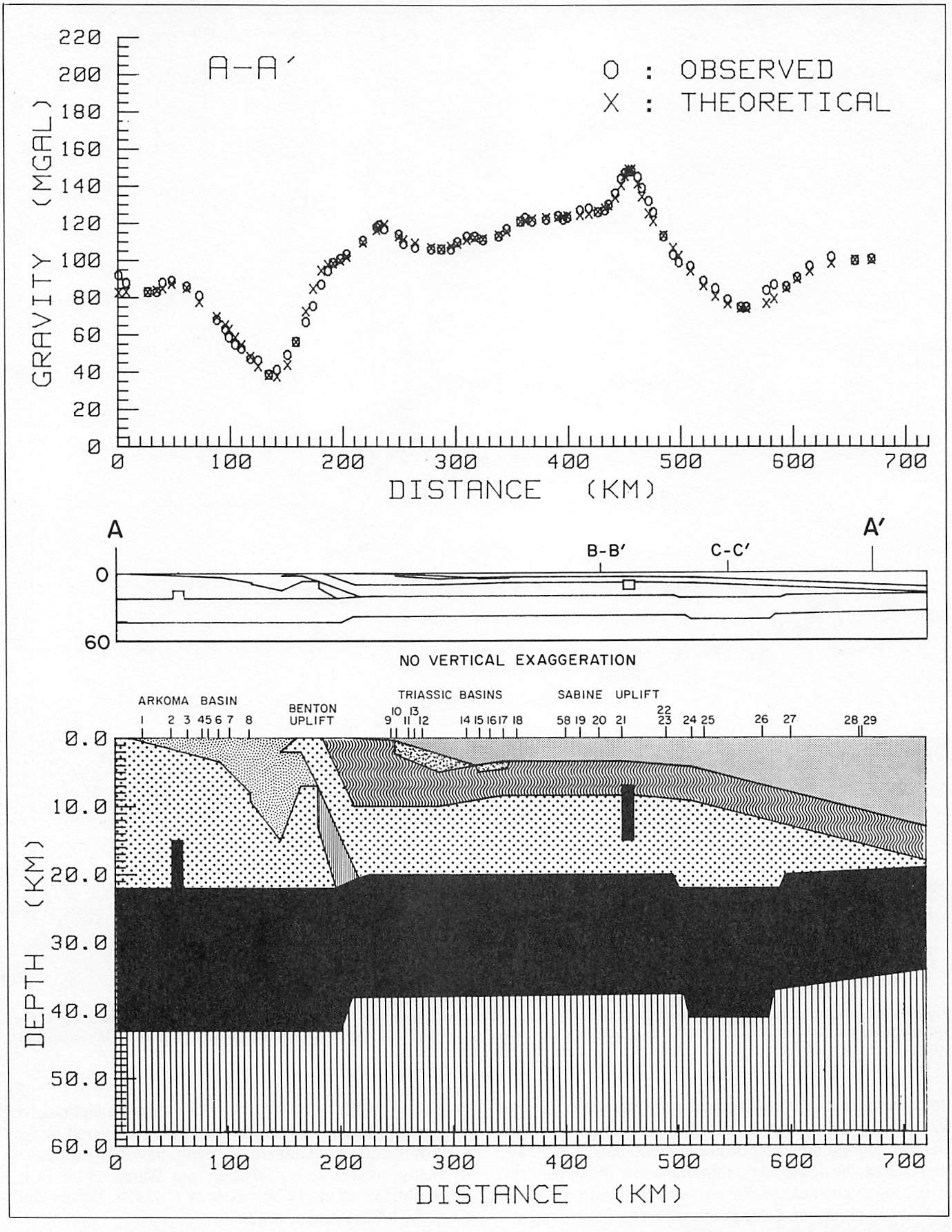

Figure 10—Gravity values and computer model derived for profile AA′. Locations of well control used are numbered in accordance with Figure 9. See Figure 14 for legend.

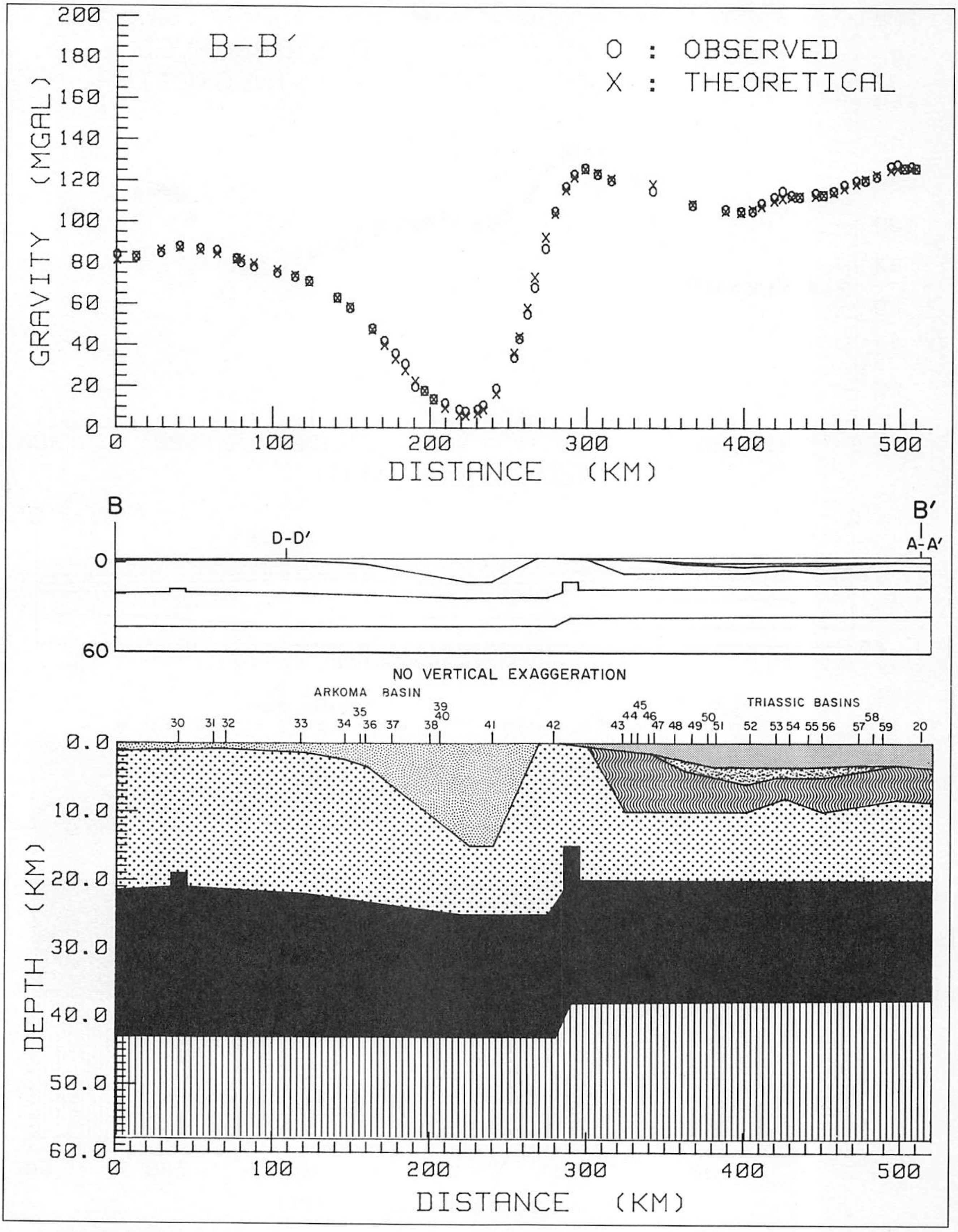

Figure 11—Gravity values and computer model derived for profile BB′. Locations of well control used are numbered in accordance with Figure 9. See Figure 14 for legend.

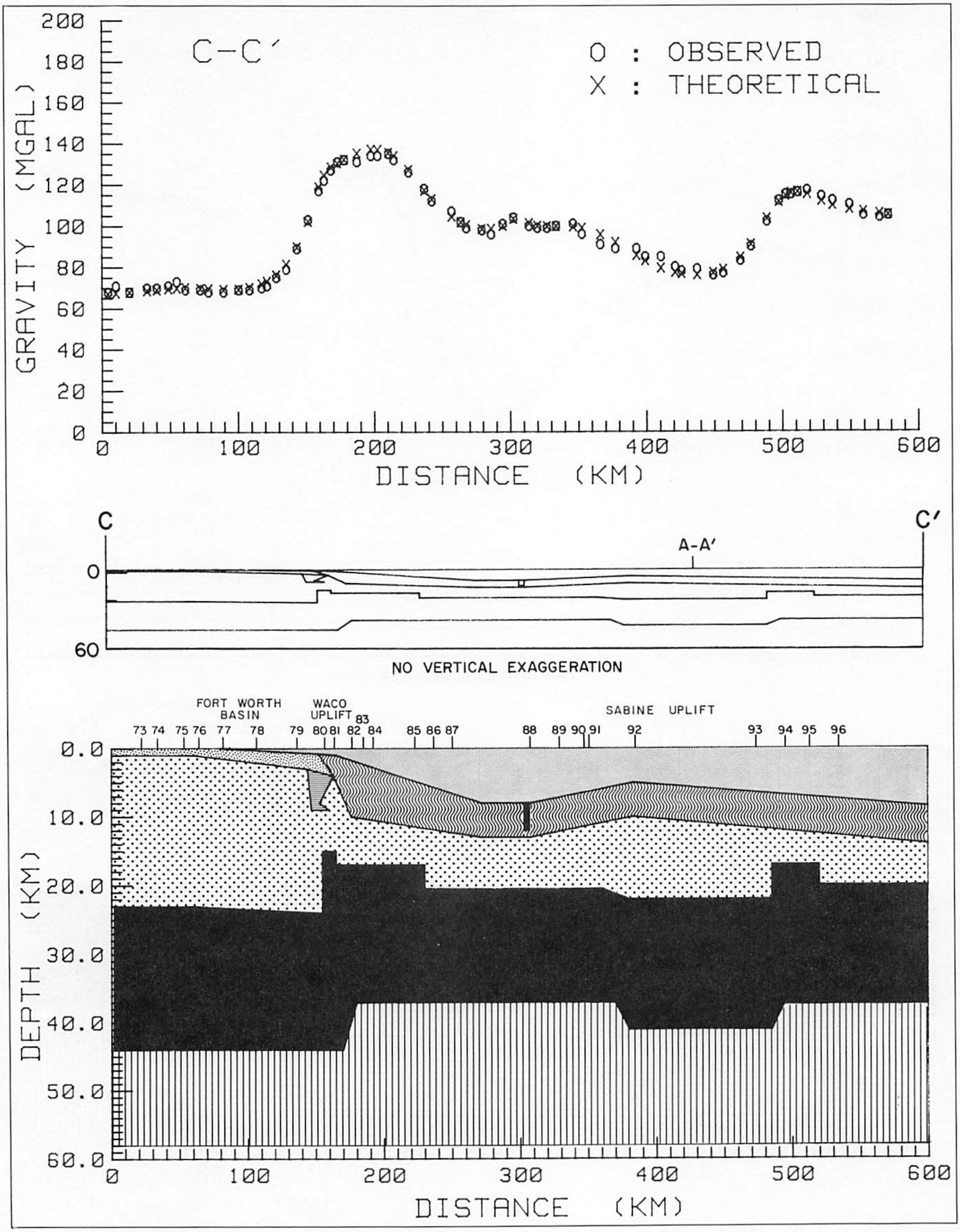

Figure 12—Gravity values and computer model derived for profile CC′. Locations of well control used are numbered in accordance with Figure 9. See Figure 14 for legend.

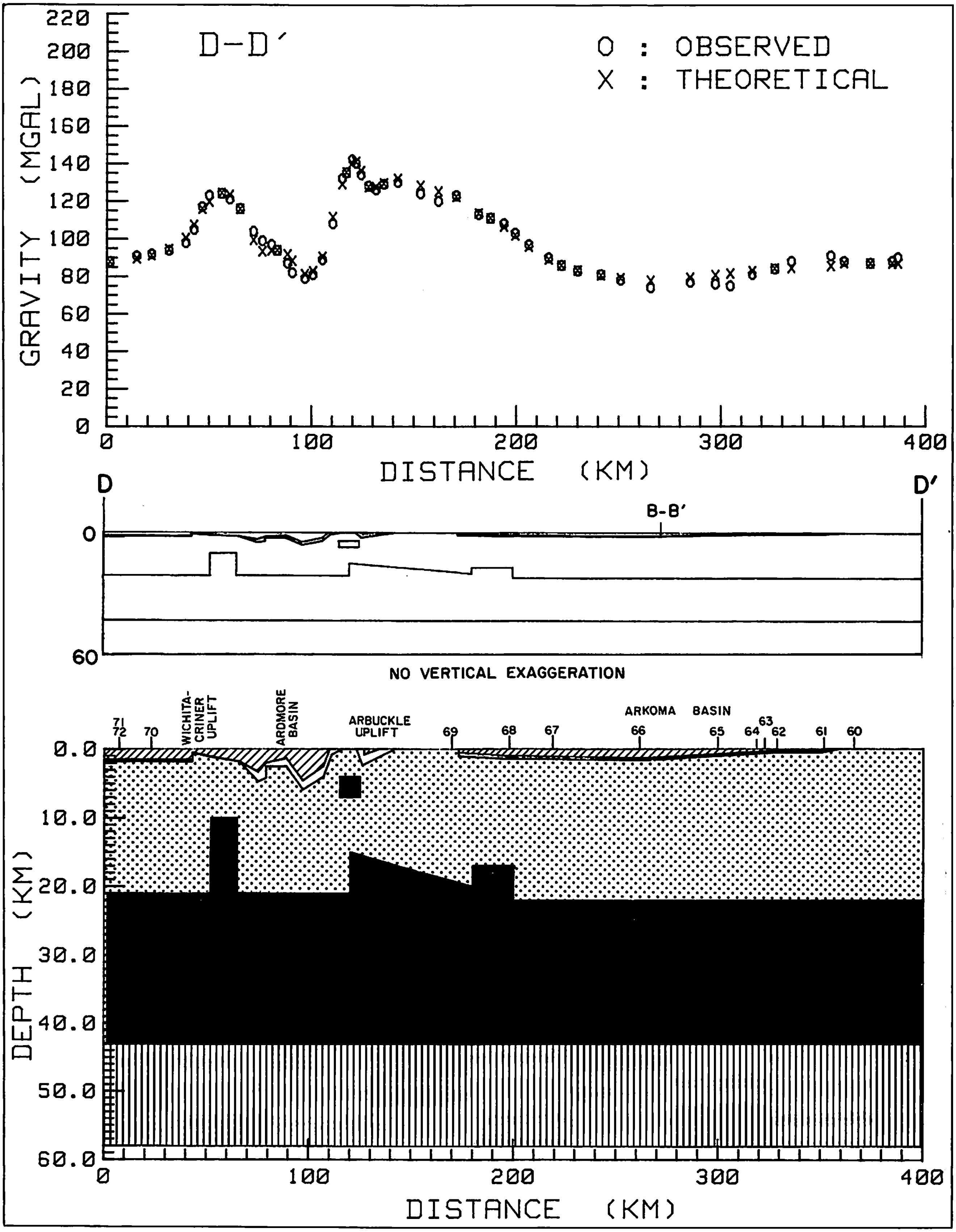

Figure 13—Gravity values and computer model derived for profile DD′. Locations of well control used are numbered in accordance with Figure 9. See Figure 14 for legend.

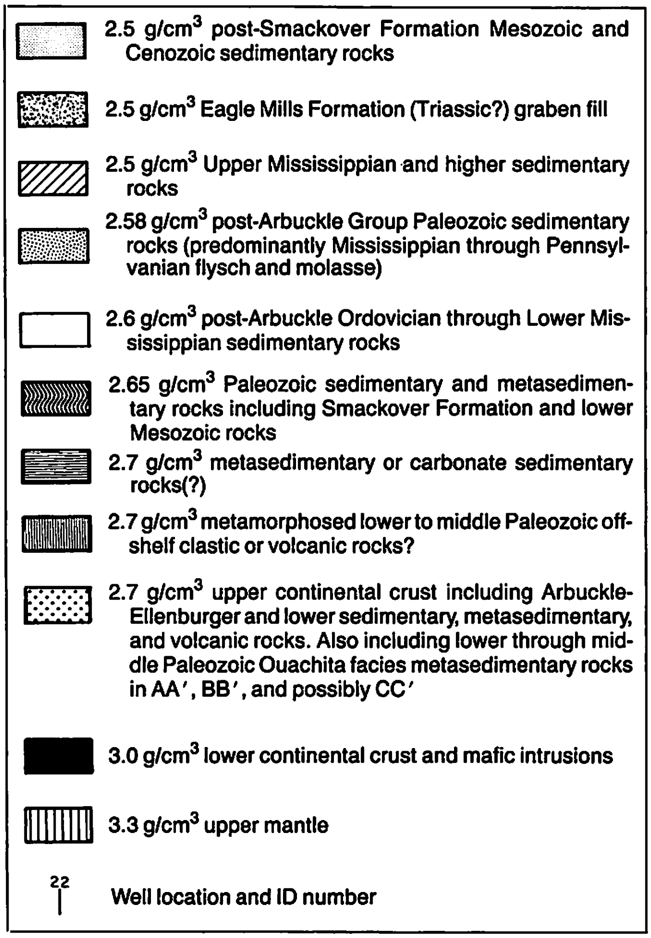

Figure 14—Legend for Figures 10-13.

1972; Keller and Cebull, 1973; Keller and Shurbet, 1975) suggest little if any upper continental crustal layer (6.0-km/sec P-wave velocity; 2.7-g/cm^3 density) is present. A similar situation of purely oceanic crust with a thick cover of sediments (Lillie et al, 1983; Lillie, 1985) seems unlikely in the study area because, if the 3.0-g/cm^3 rocks in the models are interpreted as oceanic crust, then it would be a minimum of 15 km thick (Figures 10-12). Typical oceanic crust is approximately 5 km thick (Menard, 1967; Hales, 1973); therefore, we would have to find a method of tripling the thickness of typical oceanic crust (or at least doubling it if the relatively high Bouguer anomalies are due to mantle upwarping only) to obtain the observed thicknesses. According to profiles AA′ through CC′, if the 3.0-g/cm^3 rocks beneath the Gulf coastal plain are oceanic crust, we would have to account for 10 to 15 km (6 to 9 mi) of Paleozoic sedimentary and metasedimentary rocks above the crust. To achieve an average thickness of 5 km for the 3.0-g/cm^3 rocks, we would have to thicken the overlying units further or decrease their density, which seems unreasonable due to the depths involved.

In some ways, the Ouachita system in the Ouachita Mountains area is analogous to the Carpathian system. Both systems are arcuate and occur within reentrants into cratons. Structures in both systems are predominantly convergent and verge toward the cratons. Both contain a thick, folded, and thrust-faulted flysch sequence above cratonic basement and probably result from continent-continent or continent-microcontinent collision, with the plate containing the craton being subducted. In both cases, the overriding plates were probably highly deformed as they encountered an irregular continental margin, whereas basement within the cratonic areas received little internal deformation as a result of the collision. Volcanism and metamorphism also occurred in the overriding plates of both systems, but the thick cover of sediments in the Gulf coastal plain precludes further checking of this analogy. Although the Gulf coastal plain may be analogous to the overriding plate in the Carpathian system, the latter exposes basement as well as other rocks deformed during the collision, whereas the former does not. However, large sedimentary basins occur within the overriding plate of the Carpathian system and may be analogous to late or post-orogenic basins within the Gulf coastal plain. Exposures of compressionally deformed rocks do not occur in the Gulf coastal plain as they do in the Carpathian system, probably because of regional postcollisional extension, subsidence, and deposition in the Gulf coastal plain as opposed to localized basin development within the overriding plate of the Carpathian system. See Burchfiel (1976, 1980) and Royden et al (1983a, b) for a discussion of the geology and tectonic evolution of the Carpathian area.

Arkoma Basin Minimum

A large gravity minimum underlies the Ouachita Mountains and Arkoma basin of Arkansas and Oklahoma (Figure 6) and has been modeled in profiles AA′ through CC′ (Figures 10-12). Bouguer values within this trough are the lowest in the study area (less than −110 mgal) and suggest a great thickness of sediments and/or thickening of the crust.

Figures 6-8 show this minimum is elongated in two directions approximately perpendicular to one another. The primary axis of elongation is east-northeast, approximately parallel to the northeast trend of structures in the westernmost Ouachita Mountains and Arkoma basin of Oklahoma. However, it diverges from the more east-west-trending structures of the Ouachita Mountains and Arkoma basin in Arkansas. This east-northeast trend is particularly evident in Figure 7 where northwest-trending anomalies were removed by strike filtering. The secondary axis (Figure 6) trends northwest from the corner of the minimum (lat. 35°N, long. 96°W). This anomaly trend is particularly clear on Figure 8 where anomalies associated with the Ouachita system have been attenuated by strike filtering. One prominent feature of the northwest arm of the Arkoma basin minimum is its parallelism to the trend of the southern Oklahoma aulacogen (Figures 6, 8). Because of this parallelism, this anomaly may delineate a Cambrian basin which formed during the rifting episode that initiated development of the southern Oklahoma aulacogen. It also may have been formed, or deepened (profile DD′, Figure 13), by stresses that deformed the aulacogen during the late Paleozoic. These stresses may have downwarped the upper or lower crusts, or both.

The gravity model of profile AA′ (Figure 10) suggests that the east-northeast-trending gravity minimum near the profile may be due solely to a thick sequence of upper Paleozoic sediments (flysch and molasse), assuming that an average density of 2.58 g/cm^3 for the rocks is reasonable and that the structural configuration of the basin as determined by COCORP (Nelson et al, 1982; Lillie et al, 1983; Ando et al, 1984; Lillie, 1984) and well information is valid. This thick sedimentary section agrees with the findings of Lillie et al (1983) but cannot preclude the possibility of a thinner or denser sedimentary section in conjunction with crustal thickening. However, to model profile BB′, the minimum in the area of this profile must result not only from a thick sedimentary section, but also from a downwarp of the upper crust into the lower crust, and/or the lower crust into the mantle (Figure 11).

Prior to arriving at the model shown in Figure 11, theoretical gravity values were calculated for a model in which a 2.58-g/cm^3 density for the upper Paleozoic sediments of the flysch basin was maintained to a depth of 20 km (12 mi), 2 km (1 mi) above the lower crust. In this preliminary model (not shown), no downwarp of the upper or lower crust was used. The gravity values calculated from this model were still at least 20 mgal greater than the observed anomalies over the deepest part of the basin. Using a triangular-shaped basin, the calculated gradient on the north side of the minimum was steeper than the observed gradient. Therefore, we concluded that for areas around profile BB′, and possibly for areas around profile AA′, an upper crustal and/or lower crustal downwarp (Figure 11) is needed in conjunction with the upper Paleozoic sedimentary basin to explain the Arkoma basin minimum. If the assumed 2.58-g/cm^3 average density of the upper Paleozoic sediments is too low, then upper and/or lower crustal downwarping is even more likely. This downwarping could reflect the effects of lithospheric flexure as discussed by Karner and Watts (1983).

A final observation concerning anomalies in the Arkoma basin area is that the divergence between near-surface geologic trends and gravity trends associated with gross crustal structure attests to the allochthonous nature of the Ouachita system.

Southern Fort Worth Basin Minimum

Like the Arkoma basin minimum, the minimum beneath the southernmost part of the Fort Worth basin (less than −35 mgal, Figure 6) and the smaller minimum northwest of the Lampasas arch (Figure 1) probably result from a late Paleozoic sedimentary (flysch and molasse) basin separated by an arch, and upper and/or lower crustal downwarping of the craton (Figure 6). Profile CC′ was chosen to follow the seismic profile and geologic cross section of Rozendal and Erskine (1971) and Nicholas and Rozendal (1975). Thus, it extends across the extreme northern edge of the Fort Worth basin minimum, but the model still includes a late Paleozoic sedimentary basin that thickens eastward toward the edge of the craton. This basin probably deepens southwestward, toward the center of the minimum. The upper crustal downwarp, as modeled in profile CC′ (Figure 12), may also increase southward.

The part of the basin between wells 79 and 80 (Figure 12), which is between the Waco uplift and the normal faulted Ellenberger shelf (Rozendal and Erskine, 1971; Nicholas and Rozendal, 1975) has been modeled with a density of 2.7 g/cm^3 because the 2.58-g/cm^3 density assumed for the remaining basin sediments produced a gravity minimum that was not observed. Thus, we concluded that the rocks between the Waco uplift and the normal faulted shelf to the northwest are either metamorphosed Paleozoic Ouachita facies rocks, carbonates, "granitic" basement, or a combination of these units.

Ouachita System Interior Zone Maximum

An arcuate, elongated, gravity maximum, composed of several interconnected maxima, lies gulfward of and parallel to the Ouachita system gradient (Figure 6). This maximum has been interpreted to result from various phenomena, such as metamorphic effects (densification) in the Ouachita system interior zone, basement uplifts (e.g., Broken Bow and Waco), mafic intrusions, the orogenic core of the Ouachita system, and a major crustal structure transition (e.g., Flawn et al, 1961; Watkins, 1961; Fish, 1970; Keller and Cebull, 1973; Nicholas and Rozendal, 1975; Lillie et al, 1983; Lillie, 1984). These interpretations may be valid locally, but a single interpretation probably is not valid along the entire length of this anomaly (Arkansas to Mexico).

In parts of the study area (profile AA′), crustal structure transition, basement uplifts, an interior zone of metamorphic rocks, and a Mesozoic basin south of the maximum explain the observed gravity anomalies. However, these features alone are not enough to model the maximum in other parts of the orogene (profiles BB′ and CC′). Thus, in general, dense (3.0 g/cm^3) rock in the upper crust may be needed to explain the discrepancy. Such rock could represent an upwarp of the lower continental crust, mafic intrusions, accreted island-arc material, obducted oceanic crust, or relatively undeformed oceanic crust. Because of the location of this rock (290 km, Figure 11; 160-230 km, Figure 12) with respect to the major crustal transition between the craton and the coastal plain, and because of its location within a possible zone of weakness (old rift zone), we suggest that this dense rock (not the entire anomaly) could result from igneous intrusions that formed during Eocambrian extension in the area. Intrusions associated with aborted Mesozoic rifting may have followed this zone of weakness and may be partly the source of this body. Also, a series of positive magnetic anomalies occurs along this trend (Hinze and Zietz, 1985).

Minima Gulfward of Interior Zone Maximum

A series of arcuate, elongated minima lie gulfward of the interior zone maximum (Figure 6). These anomalies extend from the Monroe uplift around the northern and western parts of the Sabine uplift through the East Texas Salt basin and northwestern arm of the Gulf Coast salt dome province into the southeast Texas Gulf coastal plain region.

As shown in Figures 10-12, these anomalies have been modeled as a series of Triassic and younger basins. We believe the crust was attenuated (thinned) during Triassic and Jurassic(?) rifting. This deformation resulted in horst and graben development. After rifting ceased, cooling of the attenuated crust caused subsidence and formation of the deep elongated basins filled by Mesozoic and Cenozoic sedimentary rocks. This hypothesis is supported by the presence of Triassic grabens and red beds (Eagle Mills Formation), Triassic(?) intrusions and flows (Vernon, 1971), and Jurassic flows and ash-fall tuffs (Jackson and Seni, 1983) in an area within and on the margins of the gravity minima. Jackson and Seni (1983) proposed a similar evolutionary scheme for the East Texas Salt basin (Figure 1).

Sabine Uplift Maximum

The gravity maximum in the northern and central parts of the Sabine uplift consists of an isolated circular high superimposed on a more regional maximum (Figure 6). Also superimposed on the regional high is a series of smaller wavelength minima and maxima. The crust underlying the regional maximum is probably attenuated continental crust; thus, this high is the result of a crustal uplift and shallowing of the Moho and/or upper crust-lower crust boundary toward its center (Figures 10, 11). The uplift may have occurred as an isostatic or thermal response to cooling and regional subsidence of the crust (Jackson and Seni, 1983). The shorter wavelength anomalies superimposed on the northern margins of this maximum probably result from Triassic horsts and grabens but could be due to intrusions. The isolated, roughly circular maximum in the southern Sabine uplift area (Figure 6) probably results from an igneous intrusion (450 km, Figure 10). This interpretation is supported by similar circular features such as the Jackson dome, and those within the Sharkey platform area (eastern Monroe uplift, Figure 1) that are known to be underlain by igneous rocks (Harrelson and Bicker, 1979). The location of this intrusion may have been influenced by a zone of weakness in the upper crust adjacent to a crustal thickness transition occurring south of the intrusion (Figure 10).

Minimum South of Sabine Uplift

South of the isolated high on the Sabine uplift, a regional minimum occurs with relatively steep gradients and residual Bouguer values less than −30 mgal (Figure 6). This anomaly has been interpreted (Figures 10, 12) to be associated with an area of relatively thick, unattenuated continental crust. A similar anomaly is associated with the Wiggins arch in southern Mississippi (Figure 1), where a 40-km (25-mi) thick crust is reported adjacent to the 28-km (17-km) thick crust under the Mississippi Salt basin to the north (Warren et al, 1966; Worzel and Watkins, 1973). We suggest that, during the Mesozoic spreading episode in the gulf, these areas remained relatively unextended whereas the surrounding continental crust was thinned by extension. The shape of the minima and, hence, the geometries of these relatively unattenuated crustal fragments probably result from the stress orientations during extension and the trends of inherent weaknesses in the crust.

La Salle Arch Area Maximum

The gravity maximum in the La Salle arch area has two primary trends (Figure 6). One trend is oriented northwest, parallel to the trend of the southeastward extension of the Ouachita orogenic belt (Figure 8), and it encompasses the northern La Salle arch and the North Louisiana Salt basin. The other trend is oriented northeast (Figure 7), and includes the southern La Salle arch and an area to the southwest.

The northeast-trending part of this maximum has been modeled on profile CC′ (500 km, Figure 12) as a mafic intrusion or lower crustal upwarp similar to that associated with the Ouachita interior zone maximum. A mantle upwarp could also give the same result. We can assume that the northwest-trending part of the maximum is due to a similar feature. This inferred magmatic activity probably occurred during Mesozoic rifting and crustal attenuation.

Isolated Maxima Along Southeastern Extension of Ouachita System

Isolated, roughly circular gravity maxima extend from the Benton uplift toward the Monroe uplift (Figure 6). They overlie and occur adjacent to the intrusions exposed around Little Rock, Benton, Magnet Cove, and Hot Springs, Arkansas, as well as intrusions (e.g., Jackson dome) buried beneath the sedimentary cover (Harrelson and Bicker, 1979). These isolated maxima probably result from Cretaceous(?) intrusions in the upper crust (Flawn et al, 1961). The occurrence, location, and trend of these intrusions are probably due to a zone of weakness (transform fault?) (Cebull et al, 1976; Thomas, 1976), which formed during Eocambrian rifting. Thus, these intrusions may approximately delineate the southwestern edge of the proto–North American continent.

Wichita Uplift–Muenster Arch Maximum

The Wichita-Criner uplift and the Muenster arch are part of the southern Oklahoma aulacogen (e.g., Hoffman et al, 1974) and lie on a prominent feature, consisting of a northwest-trending linear gravity maximum and numerous, superimposed, roughly circular isolated maxima (Figures 5, 6, 8), that ends at the Ouachita system maximum. Although the uplifts and arches contribute to the linear gravity maximum, they are totally inadequate as a cause of these anomalies. Thus, a mafic intrusion (similar to a dike) or lower crustal upwarp was used in modeling profile DD′ (Figure 13) to account for the observed anomalies. Profile DD′ extends over part of the maximum that has relatively low simple Bouguer values when compared to the rest of the maximum; therefore, the dimensions of the mafic intrusion shown in Figure 13 are probably minimum values. This intrusion probably extends the length of the maximum (Papesh, 1983).

The isolated, roughly circular maxima along this trend (Figure 6) are probably due to shallow mafic intrusions. These features also correlate well with prominent positive magnetic anomalies (Hinze and Zietz, 1985). The mafic intrusions creating these anomalies were probably intruded into the upper crust during the Eocambrian rifting episode that created the southern Oklahoma aulacogen. If, according to Denison (1982), the southern Oklahoma area underlain by Cambrian rhyolites represents the regional central rift graben, and areas underlain by Precambrian granitic and metamorphic rocks represent the margins of the central graben, then these mafic intrusions were probably injected into fractures along the southwestern boundary fault zone of the central graben. These intrusions also may have affected uplift of the crust during late Paleozoic deformation.

Arbuckle Uplift–Seminole Arch Maximum

The Arbuckle uplift–Seminole arch maximum is triangular (Figure 6). Isolated maxima along the Arbuckle uplift probably resulted from intrusions at or near the surface (110 km, Figures 7, 13). As in the Wichita uplift–Muenster arch, the basement uplift observed is insufficient to produce the observed anomalies. A wedge-shaped intrusion or lower crustal upwarp is required (Figure 13) to satisfy the observations.

The most straightforward interpretation would be that the mafic intrusions comprising the Arbuckle uplift–Seminole arch maximum formed during the same Eocambrian rifting event that caused the linear maximum to the south. However, the deep wedge-shaped body at least generally reflects the basement geometry in the Arbuckle uplift region, suggesting that the uplift may have involved the entire crust. This suggestion is not unreasonable considering the large-scale effects of Eocambrian rifting and the Ouachita orogeny. Nevertheless, the Arbuckle uplift lies either on or northeast of the northeastern flank of the Eocambrian central graben. This position, like the linear maximum to the southwest, may have been conducive to the formation of intrusions along rift boundary faults. A similar interpretation has been proposed by Hildenbrand et al (1982) for intrusions along boundary faults of the Mississippi Valley graben (Figure 1).

Ardmore Basin–Anadarko Basin Minima

The Ardmore and Anadarko basin minima are both northwest-trending, elongated minima corresponding to Paleozoic basins (Figures 5, 6, 8). The Anadarko basin minimum, which is separated from the Ardmore basin minimum by a small saddle, is less elongated than the Ardmore basin minimum. However, gravity anomalies in the Anadarko basin area are complicated by deep crustal structure (Papesh, 1983).

The Ardmore basin minimum, modeled in Figure 13, is partly due to the thick sequence of sedimentary rocks and partly due to its position between two large maxima. Unlike the Ardmore basin, which probably developed initially as a regional central graben, the Anadarko basin has the gravity expression of an asymmetrical basin with the major fault-bounded side adjacent to the Amarillo-Wichita uplift gravity maximum.

DISCUSSION AND CONCLUSIONS

The results of this study are a modified interpretation of the tectonic history of the Ouachita Mountains area and of the deep-seated structural relationships present (Figure 15). These interpretations are neither totally new nor unique. However, we have strived to produce a synthesis that is consistent with available geophysical, geological, and drilling data. We hope that the ideas presented will foster additional data collection efforts, particularly deep seismic studies.

The evolution of the Ouachita system and southern Oklahoma aulacogen probably began with Eocambrian rifting that formed the proto–North American continent, proto–Gulf of Mexico, and proto–Atlantic Ocean (Keller and Cebull, 1973). The southern margin of this Paleozoic proto–North American continent probably coincided with the trend of the Ouachita system interior zone gravity maximum or Ouachita system gravity gradient (Nicholas and Rozendal, 1975; Ando et al, 1984; Lillie, 1984), including the northwest-southeast trend of isolated gravity maxima southwest of the Mississippi Embayment. During this rifting episode, failed rifts such as the southern Oklahoma aulacogen (Hoffman et al, 1974) and Reelfoot rift–Mississippi Valley graben (Ervin and McGinnis, 1975) formed at triple junctions (Burke and Dewey, 1973). Mafic intrusions probably occurred along these rifts, thus creating linear gravity maxima, and more isolated, superimposed circular maxima were intruded closer to the surface. As suggested by Lowe (1985), the successful rift (i.e., ocean basin) in the Ouachita system may have been farther south than generally assumed.

Parts of the Ouachita system interior zone gravity maximum also probably result from mafic intrusions that remained as rift remnants when the other two arms of the triple junction developed into an ocean basin. The location of the interior zone gravity maximum relative to the proposed edge of the proto–North American continent seems analogous to the position of the East Coast gravity anomaly relative to the edge of the present-day North American continental crust (e.g., Rabinowitz, 1982). Another conse-

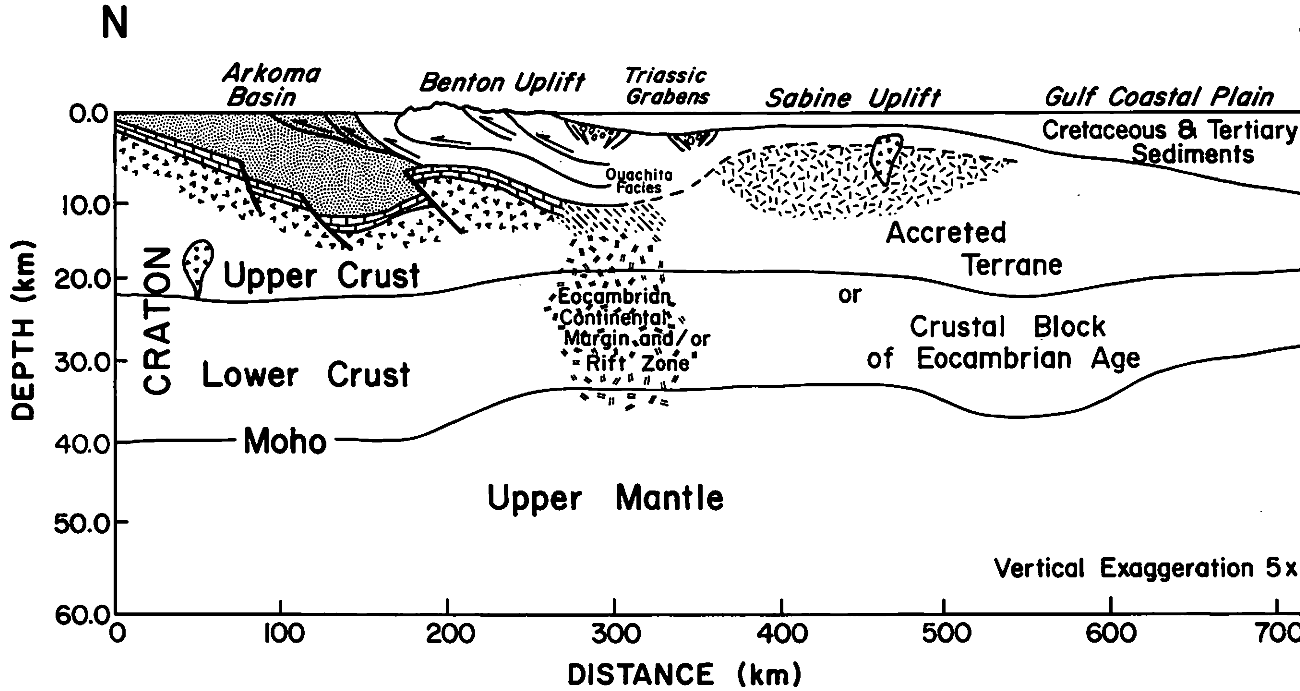

Figure 15—Interpretive cross section approximately along profile AA′.

quence of this rifting episode was that thinning of the proto-North American crust toward the south may have produced, in total or in part, the Ouachita system gravity gradient (Lillie et al, 1983; Ando et al, 1984; Lillie, 1984, 1985). A similar explanation has been proposed for the Appalachian system gravity gradient (Cook, 1983; Ando et al, 1984; Lillie, 1984).

During the Eocambrian rifting event, the Oklahoma basin—the precursor to the Anadarko and Ardmore basins—began to develop. This development may have included an intracratonic basin parallel to, but northeast of the Arbuckle uplift, thus explaining the northwest extension of the Arkoma basin gravity minimum.

After continental rifting ceased and an ocean had developed, Paleozoic passive margin marine sedimentation began along most of the Ouachita orogenic belt and continued until the late Paleozoic Ouachita orogeny (Keller and Cebull, 1973; Morris, 1974; Thomas, 1976). Cooling of the crust caused subsidence of the failed rift zone in southern Oklahoma until the Arbuckle-Wichita orogeny resulted in a wider intracratonic basin than had developed during rifting (Denison, 1982).

During the middle or late Paleozoic, subduction of the proto-North American plate beneath the proto-South American (Yucatan?) plate or a microcontinent began (e.g., Wood and Walper, 1974; Briggs and Roeder, 1975; Graham et al, 1975, 1976; Dickinson, 1981; Pindell and Dewey, 1982; Pindell, 1985). The resulting collision resulted in obduction of the accretionary wedge of the southern landmass onto the proto-North American craton—downwarping the North American craton southward—and development of deep flysch basins, such as the Arkoma and Fort Worth basins, on the margins of the proto-North American craton (Walper, 1977, 1982; Lillie et al, 1983; Lillie, 1984). The crustal downwarping and deep sedimentary basins resulted in foreland gravity minima (Lyons, 1967) as well as accentuating the gravity gradient on the gulfward margin of the basins.

During the final stages of collision, deformation of the flysch basins and accretionary wedges probably occurred with development of the Benton, Broken Bow, and Waco uplifts (Lillie et al, 1983). These uplifts, which are probably underlain by relatively thin basement of the proto-North American craton (Rozendal and Erskine, 1971; Nicholas and Rozendal, 1975; Lillie et al, 1983), contribute to the Ouachita system gravity gradient and interior zone gravity maximum. Obducted or relatively undeformed remnant Paleozoic oceanic crust (Lillie et al, 1983; Lillie, 1984, 1985) may contribute to the gravity maximum in some areas. In addition, Paleozoic metasedimentary rocks of the interior zone contribute to the gravity maximum (Watkins, 1961).

Gravity data indicate that the thrusting during the Ouachita orogeny was extensive. As shown in Figure 6, the apex of the gravity minimum associated with the Arkoma basin actually lies beneath the Ouachita Mountains northwest of the Broken Bow uplift. Assuming the maximum negative gravity anomaly correlates with the maximum thickness of sedimentary rocks, the extent of overthrusting is at least a few tens of kilometers. Clearly, many subthrust exploration targets are possible in this area.

During the Carboniferous, the southern Oklahoma aulacogen region was deformed (Hoffman et al, 1974; Denison, 1982), creating uplifts and basins that added to the gravity expression of the aulacogen. Whether this deformation is the result of primarily northwest-trending wrench-fault movements (Dickinson, 1981; Kluth and Coney, 1981) or southwest-trending compressional forces (Denison, 1982) is still controversial. However, both the timing and the structural relations imply a link to the Ouachita orogeny.

After late Paleozoic deformation ceased, a relatively quiescent period occurred (Flawn et al, 1961) before Mesozoic extensional stresses began to deform part of the upper plate. Triassic rifting created grabens that were filled with clastic sediments (Vernon, 1971; Woods and Addington, 1973; Jackson and Seni, 1983). Many of the deeper rift basins show up as smaller scale linear gravity minima. Diabase dikes, as well as tuffs and basalt flows, formed at this time (Vernon, 1971; Woods and Addington, 1973; Harrelson and Bicker, 1979; Jackson and Seni, 1983). On a regional scale, extension during the Mesozoic probably caused variable attenuation of the crust in part of the upper plate (Jackson and Seni, 1983).

Although no attempt to resolve the structural details of individual grabens was made for this study, these features extend across a large area (Figure 6). The Triassic grabens of the Atlantic Coast have been the target of recent exploration activity, and there is potential for increased emphasis on these features in the Gulf Coast region.

The Ouachita system interior zone gravity maximum may partly result from mafic rock that intruded the felsic upper crust during the Mesozoic. If this is the case, the interior zone maximum, or at least parts of it, may be the reactivation of a failed rift zone. One point of interest is that the Mexia-Talco and Arkansas fault zones (Figure 1) roughly parallel the interior zone maximum (Martin and Case, 1975), suggesting a major crustal zone of weakness along this trend. Triassic(?) diabase dikes also have been reported near the gravity maximum (Vernon, 1971).

Linear gravity maxima in the Gulf coastal plain, such as along the La Salle arch and North Louisiana Salt basin, may be due to mafic intrusions or highly attenuated continental crust along failed rifts. Although the North Louisiana Salt basin contains a thick, low-density sedimentary section, it is probably not thick enough to offset the gravity maximum created by the thin crust and/or mafic intrusions. However, regional gravity minima north and west of the Sabine uplift appear to result from low-density sedimentary rocks in Mesozoic depositional basins such as the East Texas Salt basin (Jackson and Seni, 1983). Locally, these low-density rocks offset the gravitational effect of crustal thinning beneath the basin.

The gravity minimum south of the Sabine uplift may be a result of relatively thick, unattenuated crust, surrounded by thinner, relatively attenuated crust. This minimum appears to be analogous to the minimum that roughly coincides with the Wiggins arch (Figure 1). Although a deep sedimentary basin may also cause the minimum south of the Sabine uplift, a post-Paleozoic basin of the scale required to cause the anomaly is not supported by well control. A Paleozoic basin could possibly cause the anomaly, but it would have to be very deep and/or contain rocks with densities unlikely for the depths involved. Gravity models across the anomaly, as well as a gravity model and refraction seismic measurements across the Wiggins arch (Warren et al, 1966; Worzel and Watkins, 1973) support thick crust as the explanation for this gravity low.

The Sabine uplift is a complex feature that may contain Triassic grabens beneath the Jurassic and younger sediments. It is also associated with undeformed upper Paleozoic carbonates, clastics, and volcanic rocks (e.g., Nicolas and Waddell, 1982). This uplift, and features such as the La Salle arch, may be due to isostatic or thermal equilibrium adjustments (Jackson and Seni, 1983; Nunn et al, 1984).

During the Cretaceous, an episode of magmatism involving mafic intrusions occurred along a trend roughly coinciding with the interior zone of the Ouachita system (Flawn et al, 1961). Many of these intrusions (some actually formed volcanoes) can be found in Texas adjacent to the Llano uplift (e.g., Pilot Knob near Austin). The more prominent features form a trend of gravity anomalies extending between the Ouachita Mountains and the Appalachian system and include Hot Springs, Magnet Cove, and Jackson dome intrusions. These intrusions seem to be localized along the zone of weakness formed during the early Paleozoic continental breakup. Isolated gravity anomalies of a similar nature, such as on the south flank of the Sabine uplift, occur south of the Ouachita trend and may be of similar origin. However, they may be Paleozoic in age and a result of subduction-related magmatism.

Regional subsidence and episodes of transgressions and regressions influenced the thick Cenozoic sedimentary section of the Gulf coastal plain. Southward thickening of these sedimentary rocks suggests that a corresponding thinning of the crust in this direction is needed to maintain isostatic equilibrium and explain the observed Bouguer gravity anomalies (Worzel and Watkins, 1973).

These events resulted in the crustal-scale, interpretive cross section shown in Figure 15. However, Figure 15 represents an integrated synthesis of available data that has many implications for exploration efforts, and it provides a model that can be tested by future studies.

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Lower Jurassic Navajo-Aztec-Equivalent Sandstones in Southern Arizona and Their Paleogeographic Significance[1]

WILLIAM L. BILODEAU[2] and STANLEY B. KEITH[3]

ABSTRACT

Thick sequences of Lower Jurassic rhyolitic and andesitic volcanic rocks in several mountain ranges of southern Arizona contain interbedded quartzarenites. Locally up to 250 m thick, these sandstone lenses, composed of well-sorted and well-rounded quartz grains, commonly contain large-scale cross-stratification and are considered to be eolian sand deposits. The eolian sands were blown up against the continental side of the Early Jurassic volcanic arc that trended northwest-southeast across the southwestern margin of the North American continent and/or plate at that time. Paleocurrent data suggest southerly eolian transport of the sands from the Colorado Plateau area. Correlation of these sandstones with the Lower Jurassic Navajo and Aztec Sandstones is indicated by the paleocurrent data as well as radiometric dating of the interbedded volcanics. Eolian sand transport southward across central Arizona in the Early Jurassic indicates that the Mogollon highlands either did not then exist, or were merely low, discontinuous inselbergs on a broad back-arc ramp, more appropriately called the Mogollon slope.

INTRODUCTION

The Early Jurassic paleogeography of southern Arizona was dominated by a northwest-southeast-trending continental margin magmatic arc of "Andean" type (Figure 1). This magmatic arc extended northwestward into California, southward into Mexico, and was directly related to eastward subduction of an oceanic plate beneath the southwestern edge of North America (Coney, 1978; Dickinson, 1981). The Lower Jurassic volcanic, plutonic, and associated sedimentary rocks of the arc terrane are widely scattered in the isolated mountain ranges of the southern Basin and Range province of southeastern California (Armstrong and Suppe, 1973; Marzolf, 1980), southern Arizona (Hayes et al, 1965; Hayes and Drewes, 1978; Haxel et al, 1980b, 1984), and northern Mexico (Anderson and Silver, 1978; Rangin, 1978). The arc terrane was separated from thick Triassic-Jurassic nonmarine sedimentary sequences to the northeast on the Colorado Plateau (Harshbarger et al, 1957) by a northwest-trending belt about 250 km wide, where virtually no rocks of comparable age are known. This belt, which trends across central Arizona and New Mexico, lies just south of the southern margin of the Colorado Plateau or Mogollon Rim and contains Precambrian basement exposed over a large area. This belt occupied a position of great paleotectonic and paleogeographic significance through geologic time, and has been given several different names: Texas lineament (Albritton and Smith, 1956; Moody and Hill, 1956), Deming axis (Turner, 1962), and Mogollon highlands (Harshbarger et al, 1957; Cooley and Davidson, 1963; Stewart, 1969; Stewart et al, 1972).

The relationship of the Lower Jurassic clastic sedimentary strata of the Colorado Plateau to the very different rocks of the volcanic-arc terrane to the southwest, across this belt of Precambrian exposure, is of significant paleogeographic importance. Coney (1978) suggested that the ancestral Mogollon highlands stood as a topographic barrier between the alluvial systems of the Colorado Plateau region and the arc terrane to the southwest. Most workers dealing with the Triassic-Jurassic strata of the Colorado Plateau consider the ancestral Mogollon highlands to have been the source area for the sediments (Harshbarger et al, 1957; Cooley and Davidson, 1963; Stewart, 1969; Stewart et al, 1972; Blakey and Gubitosa, 1984). The presence of eolian Navajo-Aztec-equivalent sandstones intercalated with Lower Jurassic arc volcanics along the rear or continent side of the arc and definitely south of the inferred northwest-trending ancestral Mogollon highlands in central Arizona suggests that a modification of prior paleogeographic reconstructions is needed (Bilodeau and Keith, 1979, 1984; Bilodeau, 1985).

Thick sequences of mature quartz sandstones intercalated with rhyolitic to andesitic volcanics of Early Jurassic age are located in several southern Arizona mountain ranges. Based on similarities in age, lithology, petrography, and paleowind directions deduced from the orientation of large-scale cross-stratification, we correlated these rocks with the Aztec Sandstone exposed 600 km to the northwest in the Mojave Desert (Miller and Carr, 1978; Marzolf, 1980, 1982), and the Navajo Sandstone 400 km to the north on the Colorado Plateau (Peterson and Pipiringos, 1979). The widespread southern Arizona quartzarenites of Early Jurassic age are best exposed in five widely separated southern Arizona localities: (1) within the Mount Wrightson For-

[1]Manuscript received, July 1, 1985; accepted, February 27, 1986.

[2]Department of Geology, University of Colorado at Denver, 1100 14th Street, Denver, Colorado 80206.
[3]MagmaChem Exploration, Inc., 10827 South 51st Street, Suite 202, Phoenix, Arizona 85044.

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