1-1-1998

Gravity constraints on lithosphere flexure and the structure of the late Paleozoic Ouachita orogen in Arkansas and Oklahoma, south central North America

Dennis L. Harry

Kevin L. Mickus
Missouri State University

Recommended Citation

Harry, Dennis L., and Kevin L. Mickus. "Gravity constraints on lithosphere flexure and the structure of the late Paleozoic Ouachita Orogen in Arkansas and Oklahoma, south central North America." Tectonics 17, no. 2 (1998): 187-202.

TECTONICS, VOL. 17, NO. 2, PAGES 187-202, APRIL 1998

Gravity constraints on lithosphere flexure and the structure of the late Paleozoic Ouachita orogen in Arkansas and Oklahoma, south central North America

Dennis L. Harry
Department of Geology, University of Alabama, Tuscaloosa

Kevin L. Mickus
Department of Geosciences, Southwest Missouri State University, Springfield

Abstract. Spectral analysis of Bouguer gravity anomalies in western central Arkansas and eastern Oklahoma indicates that the thickness of the crust in the Ouachita fold and thrust belt increases from 38 km in the western Ouachitas to 44 km in the eastern Ouachitas. The change in crustal thickness occurs near the western end of the Broken Bow uplift and coincides with an abrupt decrease in the flexural rigidity of the lithosphere from 1.8×10^{24} N m in the western Ouachitas to 5.0×10^{23} N m in the eastern Ouachitas. The flexural rigidity in the western Ouachitas is similar to values determined in the Appalachian fold and thrust belt and coincides with the depth of the 450°C isotherm predicted by conductive cooling models for the thermal evolution of the early Paleozoic southern Laurentian rifted continental margin. The thick crust in the eastern Ouachitas results in lithosphere that is anomalously weak for rifted continental crust of this age. The thicker crust is attributed to an eastward transition from a rift segment to a transform segment of the Paleozoic continental margin. A layered density model derived from the gravity data shows that strata interpreted to be deformed Ouachita facies rocks are thickest in the eastern Ouachitas and are consistent with a greater amount of shortening in the central thrust belt in Arkansas as compared to Oklahoma. The opposite relationship is observed in the frontal Ouachita province, where shortening appears greater in Oklahoma. The cross-strike changes in the locus of shortening, crustal thickness, flexural rigidity, and the inferred transition from rift to transform segments of the early Paleozoic continental margin all coincide with the location of a previously hypothesized zone of diffuse right-lateral shear located at the western end of the Benton uplift. Flexural modeling indicates that the load required to produce the observed Bouguer gravity low in the Arkoma foreland basin trends parallel to the Benton and Broken Bow uplifts but is located 114 to 276 km farther south. In the western Ouachitas, the position of the load coincides with the northern edge of the Sabine uplift and is interpreted to mark the southern extent of Ouachita facies rocks that were emplaced on the Laurentian continental margin and/or attached remnant oceanic crust.

Paper number 97TC03786.
0278-7407/98/97TC-03786$12.00

1. Introduction

The Ouachita orogen is a Paleozoic fold and thrust belt that extends more than 2100 km from Alabama to northern Mexico, with approximately 80% of the orogen being buried beneath Cretaceous sedimentary rocks [*Flawn et al.*, 1961; *Viele and Thomas*, 1989] (Figure 1). The orogen developed during the Mississippian and Pennsylvanian periods as the southern margin of Laurentia changed from an early to middle Paleozoic passive continental margin dominated by thermal subsidence and carbonate shelf buildup to a late Paleozoic active margin characterized by contractional deformation, clastic sedimentation, and the formation of flexural foreland basins [*Viele and Thomas*, 1989]. Deep seismic data are scarce, and well control is limited primarily to the hydrocarbon-bearing foreland basins and immediately adjacent segments of the fold and thrust belt. As a consequence, the deep structure of the orogen is understood only in general terms, and little is known about the amount of crustal attenuation on the underlying Paleozoic rifted continental margin or the location of the Paleozoic ocean-continent transition.

The Ouachita Mountains of Arkansas and Oklahoma contain the largest exposure of strata deformed during the Ouachita orogeny. Geologic mapping, Consortium for Continental Reflection Profiling and Program for Array Seismic Studies of the Continental Lithosphere (PASSCAL) deep seismic reflection and refraction profiles, and gravity modeling provide a relatively good record of the structural and stratigraphic evolution of this portion of the orogen [*Nelson et al.*, 1982; *Lillie et al.*, 1983; *Arbenz et al.*, 1989; *Keller et al.*, 1989a; *Mickus and Keller*, 1992]. Nevertheless, questions remain concerning the amount of allocthonous material emplaced during contractional deformation, the southern extent of Ouachita facies rocks, and the structure of the underlying rifted margin. In this paper, gravity and topography data from the Ouachita Mountains and Arkoma foreland basin are used to develop a model of synorogenic flexural deformation of the lithosphere along this portion of the orogen. The flexural model constrains the total excess mass emplaced during formation of the fold and thrust belt, the distribution of the mass in the subsurface, and the strength of the underlying rifted continental margin. These parameters are determined from the analysis of nine gravity and topography profiles oriented perpendicular to the strike of the Bouguer gravity minimum associated with the Ouachita fold and thrust belt and the Ark-

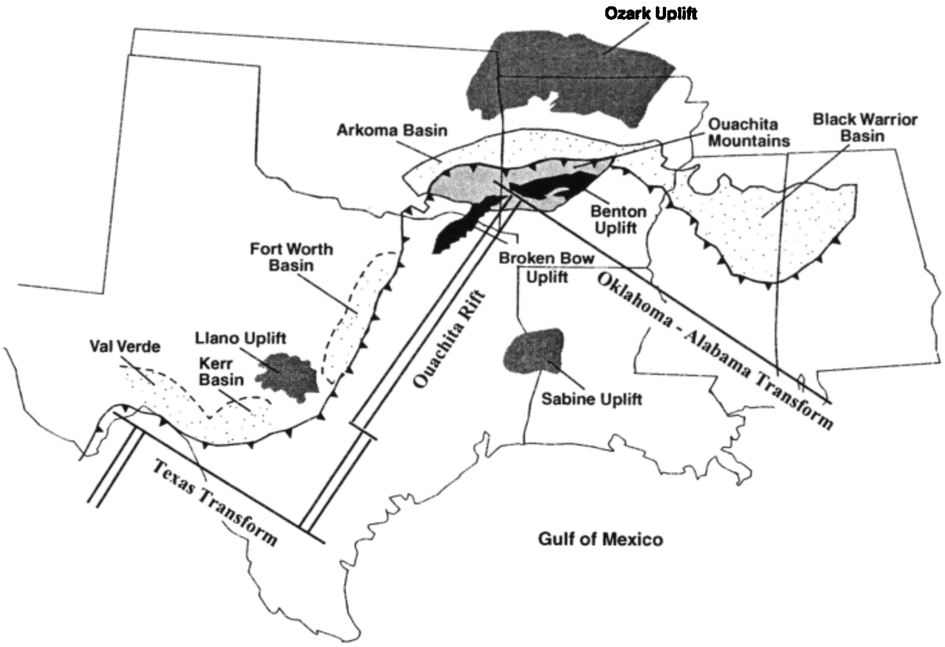

Figure 1. Tectonic setting of the Ouachita orogenic belt. Bold line indicates geometry of early Paleozoic continental margin [*Thomas*, 1993]. Solid area indicates exposed Benton and Broken Bow uplifts. Subsurface continuation of the Broken Bow uplift is indicated by dark shaded pattern. Exposed Paleozoic rocks in the Ouachita Mountains external to the Benton and Broken Bow uplifts are indicated by light shading. Medium shading indicates structural uplifts in the subsurface (Sabine uplift) and surface (Ozark and Llano uplifts). Stippled pattern indicates Paleozoic foreland sedimentary basins.

oma basin. The profiles are spaced at 25 km intervals between the eastern and western limits of the Paleozoic outcrop, traversing the Arkoma basin and the Ouachita Mountains and extending roughly 75 km south onto the coastal plain (Figure 2). Bouguer gravity power spectra from each profile are used to determine a layered density model of the crust in this region. Estimates of the Bouguer gravity coherence and admittance along each profile simultaneously constrain the flexural rigidity of the lithosphere and how the excess crustal mass emplaced during the Ouachita orogeny is partitioned between the surface and subsurface. An inverse flexural model is then used to estimate the magnitude and location of the excess mass.

2. Ouachita Fold and Thrust Belt and Arkoma Foreland Basin

The Ouachita fold and thrust belt and associated foreland basins generally follow the trend of the southern Laurentian Paleozoic passive continental margin [*Viele and Thomas*, 1989]. The geometry of the Paleozoic continental margin was established by Late Proterozoic to Early Cambrian rifting, followed by thermal subsidence and passive margin sedimentation between Late Cambrian and Early Mississippian time [*Rankin*, 1976; *Thomas*, 1976; *Viele and Thomas*, 1989]. In the Ouachita Mountains, the onset of contractional deformation is inferred from an abrupt increase in clastic sedimentation rates and the appearance of locally abundant volcanic debris in the Upper Mississippian Stanley Formation [*Niem*, 1977; *Morris*, 1989; *Loomis et al.*, 1994]. The time at which contractional deformation ended is poorly constrained, but low-amplitude, long-wavelength compressional folds in the Arkoma basin extend at least into the Late Pennsylvanian Hartshorne Formation [*Sutherland*, 1988; *Arbenz*, 1989; *Denison*, 1989; *Elmore et al.*, 1990].

Five distinct stratigraphic and structural provinces are recognized in the Ouachita Mountains region (Figure 2). From north to south, these include the Arkoma basin, the frontal imbricated thrust zone, the northern central thrust belt (central Ouachitas), the Benton and Broken Bow uplifts, and the southern central thrust belt (southern Ouachitas) [*Arbenz*, 1989; *Morris*, 1989]. The Arkoma basin contains Pennsylvanian deltaic, shallow marine, and flysch deposits that overlie carbonate rocks deposited on the former continental shelf [*Sutherland*, 1988]. The deeper southern parts of the basin contain normal faults that are inferred to have developed as a consequence of flexural subsidence of the basin [*Houseknecht*, 1986]. The leading edge of the frontal thrust zone is generally taken to be the Choctaw and Ross Creek faults, which place Lower Pennsylvanian (Jackfork Group and Johns Valley Shale) to Middle Pennsylvanian (middle Atoka Formation) strata over upper Atokan foreland basin strata [*Blythe et al.*, 1988]. The frontal thrust zone in Oklahoma is characterized by tightly spaced imbricated thrust faults involving mainly Pennsylvanian strata. Thrust faults are more widely spaced in the central and eastern Ouachitas, suggesting more shortening in the frontal thrust zone toward the west [*Arbenz*, 1989]. The northern central thrust belt consists of relatively widely spaced northwest verging thrust faults, broad synclines, and tight anticlines in mainly Pennsylvanian strata.

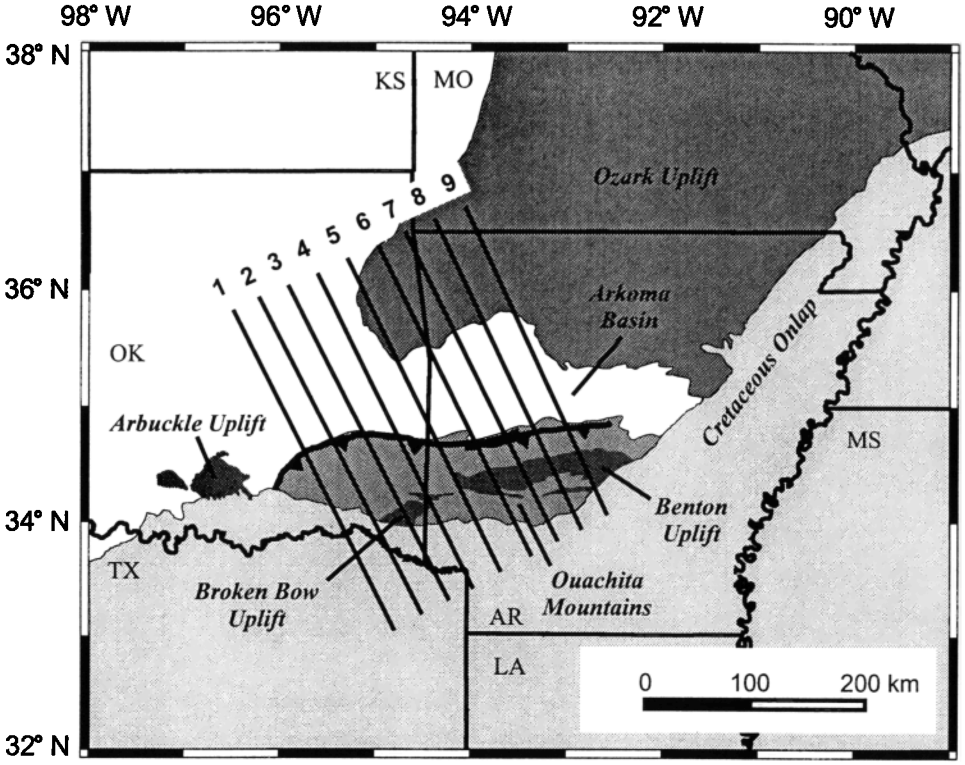

Figure 2. Simplified geologic map of Ouachita Mountains in Arkansas and Oklahoma. Bold lines indicate location of profiles analyzed in this study. The frontal imbricate zone discussed in text is restricted to the vicinity of the Choctaw (west) and Ross Creek (east) thrust faults. The Broken Bow and Benton uplifts separate the central thrust belt (medium gray) into north and south provinces.

The northern central thrust belt is structurally and stratigraphically similar to the southern central thrust belt, which is also characterized by north vergent thrust faults and broad thrust-faulted synclines in Pennsylvanian strata. The southern central thrust belt is exposed mainly in the central and eastern Ouachitas and is separated from the northern central thrust belt by the Broken Bow and Benton uplifts.

The Benton uplift in west central Arkansas and the Broken Bow uplift in southeastern Oklahoma expose the oldest strata in the Ouachita Mountains. These uplifts consist of Late Cambrian through Middle Mississippian preorogenic and synorogenic strata, termed the Ouachita facies, that are distinctly different in their degree and style of deformation from the clastic sequence found in the Arkoma basin and frontal Ouachitas [*Flawn et al.*, 1961; *Nielsen et al.*, 1989]. Both uplifts have a similar anticlinal structure and are formed by predominantly north verging listric thrust faults and attendant folds that in some places have been overturned [*Milliken*, 1988; *Arbenz*, 1989]. The thrust faults place Late Cambrian to Early Mississippian deep water sedimentary strata on top of Upper Mississippian to Early Pennsylvanian turbidites and deep water marine clastics. The Early Mississippian deep water strata are generally interpreted to have been deposited in a deep oceanic trough south of the Laurentian shelf, with the Upper Mississippian clastic sediments deposited during the early stages of closure of the ocean basin and the onset of the Ouachita orogeny [*Houseknecht*, 1986; *Viele and Thomas*, 1989]. Seismic data [*Lillie et al.*, 1983; *Milliken*, 1988; *Keller et al.*, 1989a], geological cross sections [*Blythe et al.*, 1988; *Arbenz et al.*, 1989] and gravity models [*Kruger and Keller*, 1986; *Mickus and Keller*, 1992] indicate that Ouachita facies strata are between 15 and 20 km thick beneath the Benton and Broken Bow uplifts. However, the total thickness of the thrust pile and the maximum depth of compressional deformation during contraction are unknown.

It is generally accepted that attenuated continental crust that was extended during Cambrian rifting lies beneath the Ouachita Mountains [*Keller et al.*, 1989a; *Thomas*, 1991]. However, the structure of the southern Laurentian margin and it's relationship to the deep water sedimentary rocks that comprise the Oauchita facies has been the subject of debate. *Lillie et al.* [1983] argue that the southern extent of Oauchita facies rocks lies beneath the southern Ouachitas. *Milliken* [1988], *Keller et al.* [1989a], and *Mickus and Keller* [1992] suggest that Ouachita facies strata extend as far as 300 km south of the Benton uplift. If this is the case, Ouachita facies rocks extend well south of the edge of the Laurentian continental crust interpreted by *Keller et al.* [1989a] on the basis of PASSCAL seismic data. The debate is complicated by the fact that the Ouachitas probably formed near the juncture of a northeast striking rifted segment of the Paleozoic Laurentian margin (the "Ouachita rift segment" of *Thomas* [1993]) and a southeast striking transform segment of the margin (the "Oklahoma-Alabama transform" segment) (Figure 1). The precise geometry of the Paleozoic margin is not well constrained, but this hypothesis suggests that the western Ouach-

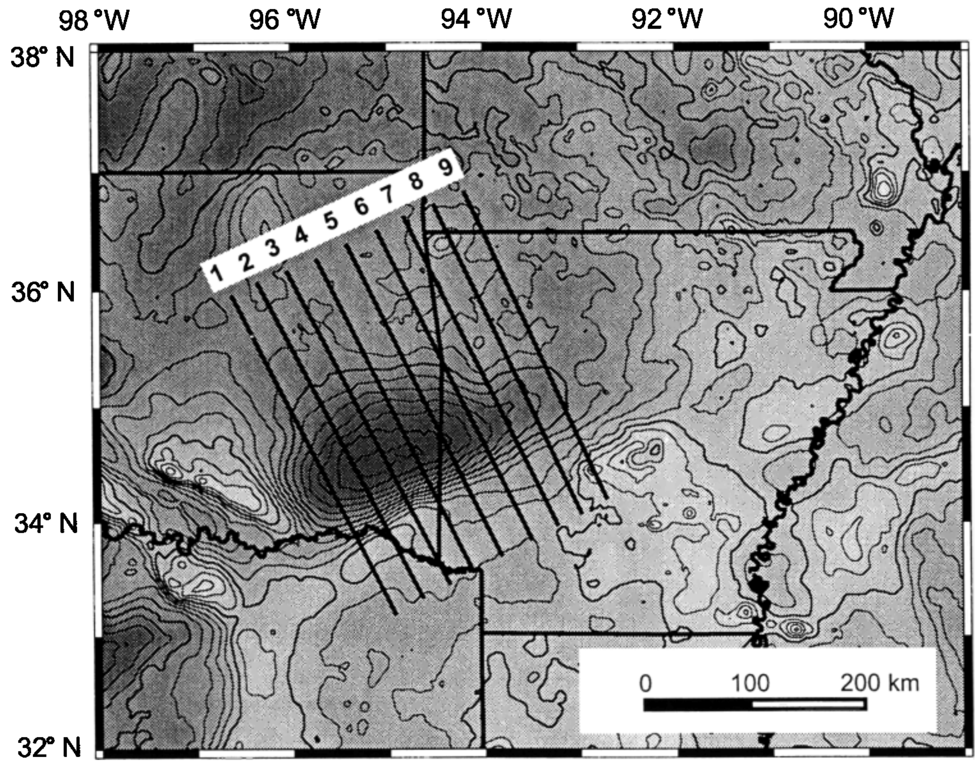

Figure 3. Bouguer gravity map of the Ouachita fold and thrust belt and the Arkoma basin region. Contour interval is 10 mGals. Gray scale ranges from -120 mGals (dark) to 30 mGals (light). Bold lines indicate location of profiles analyzed in this study.

itas overlie rifted margin crust and that the eastern Ouachitas overlie less attenuated crust on the transform margin.

3. Spectral Analysis of Gravity and Topography

The central to northern Ouachitas and Arkoma basin are associated with one of the largest-amplitude Bouguer gravity minima in the United States, reaching a minimum value of <-110 mGals in west central Arkansas and southeastern Oklahoma (Figure 3). A parallel Bouguer gravity maximum is located immediately south of the trend of the Broken Bow and Benton uplifts and reaches maximum values +10 mGals approximately 75 km south of the axis of the Bouguer gravity minimum. The Bouguer gravity maximum is best defined in Oklahoma and becomes less pronounced toward the east. This coupled Bouguer gravity maximum and minimum is characteristic of fold and thrust belts, with the gravity minimum being produced by subsidence of the foreland basin and the gravity maximum being created by a buried excess mass within the thrust belt. The mass excess may represent either thrust emplacement of deep crustal or allocthonous rocks onto less dense shelf facies or deep crustal synorogenic metamorphism [*Karner and Watts*, 1983; *Forsyth*, 1985].

Nine gravity and topography profiles in the Ouachita Mountains region were constructed by projecting 2x2 km gridded gravity and topographic data onto profiles oriented perpendicular to the trend of the regional Bouguer gravity minimum (Figure 3). Only data lying within 5 km of the projected profiles were used. The resulting profiles had a maximum sample interval of about 0.5 km. To minimize aliasing, the data were filtered to remove wavelengths shorter than 4 km and resampled to a uniform 2 km interval using spline interpolation. Each profile is 350 km long and extends from approximately 75 km south of the southern exposure of the Ouachita orogenic belt to 150-200 km beyond the thrust front. The topography and Bouguer gravity profiles are shown in Figure 4.

A layered density model of the lithosphere was determined from the slope of the Bouguer gravity power spectrum. Subsurface density interfaces result in distinct linear segments in the logarithmic Bouguer gravity power spectrum [*Banks et al.*, 1977; *Karner and Watts*, 1983]. The depth of each interface is given by one half of the slope of the appropriate segment of the power spectrum, with steep slopes corresponding to deep interfaces. The Bouguer gravity spectra for the nine profiles (Figure 5) show well-resolved linear segments that are interpreted as the density interface at the crust-mantle boundary (at depths of 38 to 44 km) and the base of the clastic section in the Arkoma basin (9.5 to 11.7 km). Two additional interfaces are indicated in the shallow crust at depths of 1.3 to 3.1 km and 5.1 to 7.7 km. Comparison with forward gravity models in the region [*Kruger and Keller*, 1986; *Mickus and Keller*, 1992] suggest that the interface between 5.1 and 7.7 km represents the base of allocthonous Ouachita facies sediments within the central Ouachitas. The interface between 1.3 and 3.1 km is only observed at wavenumbers greater than one fourth the Nyquist wavenumber ($\pi/2$ km^{-1}),

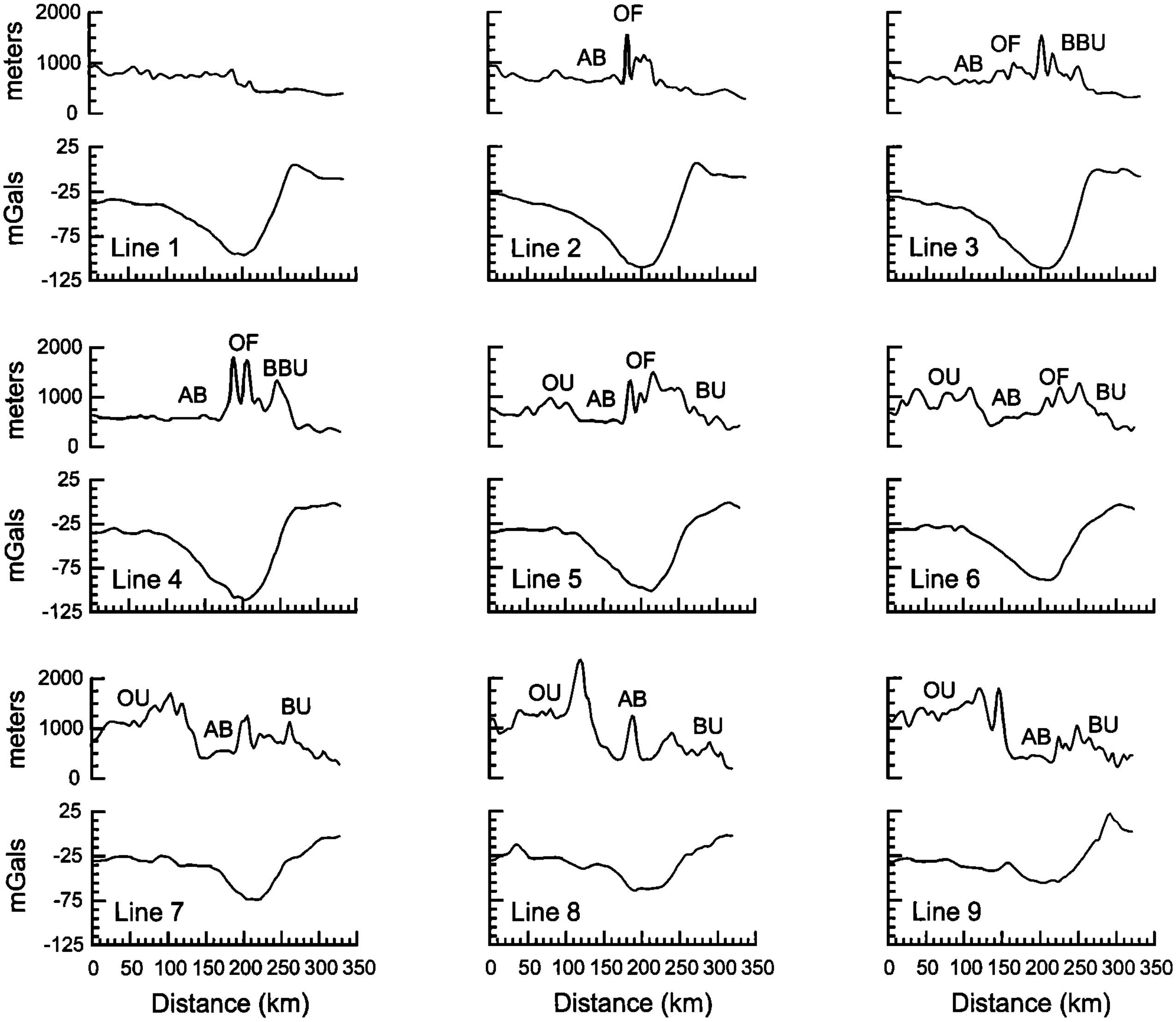

Figure 4. (top) Topography and (bottom) Bouguer gravity profiles analyzed in this study. Profile 1 is the westernmost profile; profile 9 is the easternmost. Abbreviations are as follows: AB, Arkoma basin; BBU, Broken Bow uplift; BU, Benton uplift; OF, Ouachita front; and OU, Ozark uplift.

so it questionable whether the data sampling interval is sufficient to adequately constrain the depth of this interface. If the depth estimates are correct, it may reasonably be interpreted to indicate either the base of clastic sediments within a Triassic rift basin that has been seismically imaged south of the Ouachita Mountains [*Milliken*, 1988; *Keller et al.*, 1989a] or the base of coastal plain sediments south of the Ouachita outcrop. An interface detected on line 7 at 17.9 km probably represents a midcrustal density contrast within autocthonous North American crust that has been modeled by previous workers [*Mooney and Weaver*, 1989; *Taylor*, 1989]. This interface is not clearly resolved on the other profiles.

4. Admittance and Coherence

The Bouguer gravity anomaly in the Ouachitas is produced primarily by density contrasts in the subsurface arising from flexural subsidence of the lithosphere and from emplacement of excess subsurface mass during the orogeny. These two phenomena are coupled, since flexural deformation is a response to excess mass emplacement. The mass emplaced during the orogeny may include a surface load due to topography and a subsurface load. The subsurface load is probably primarily due to emplacement of a thick wedge of deep water Ouachita facies rocks onto the edge of the Paleozoic continental shelf, but it could conceivably also involve placement of some midcrustal and deeper rocks onto less dense shallower strata and/or metamorphism of rocks in the lower crust during the orogeny. Flexural subsidence of the crust produces a broad-wavelength Bouguer gravity minimum in the foreland, whereas the buried excess mass produces an adjacent shorter-wavelength Bouguer gravity maximum. The flexural rigidity of the lithosphere and the ratio of surface to subsurface loading can be simultaneously determined by considering

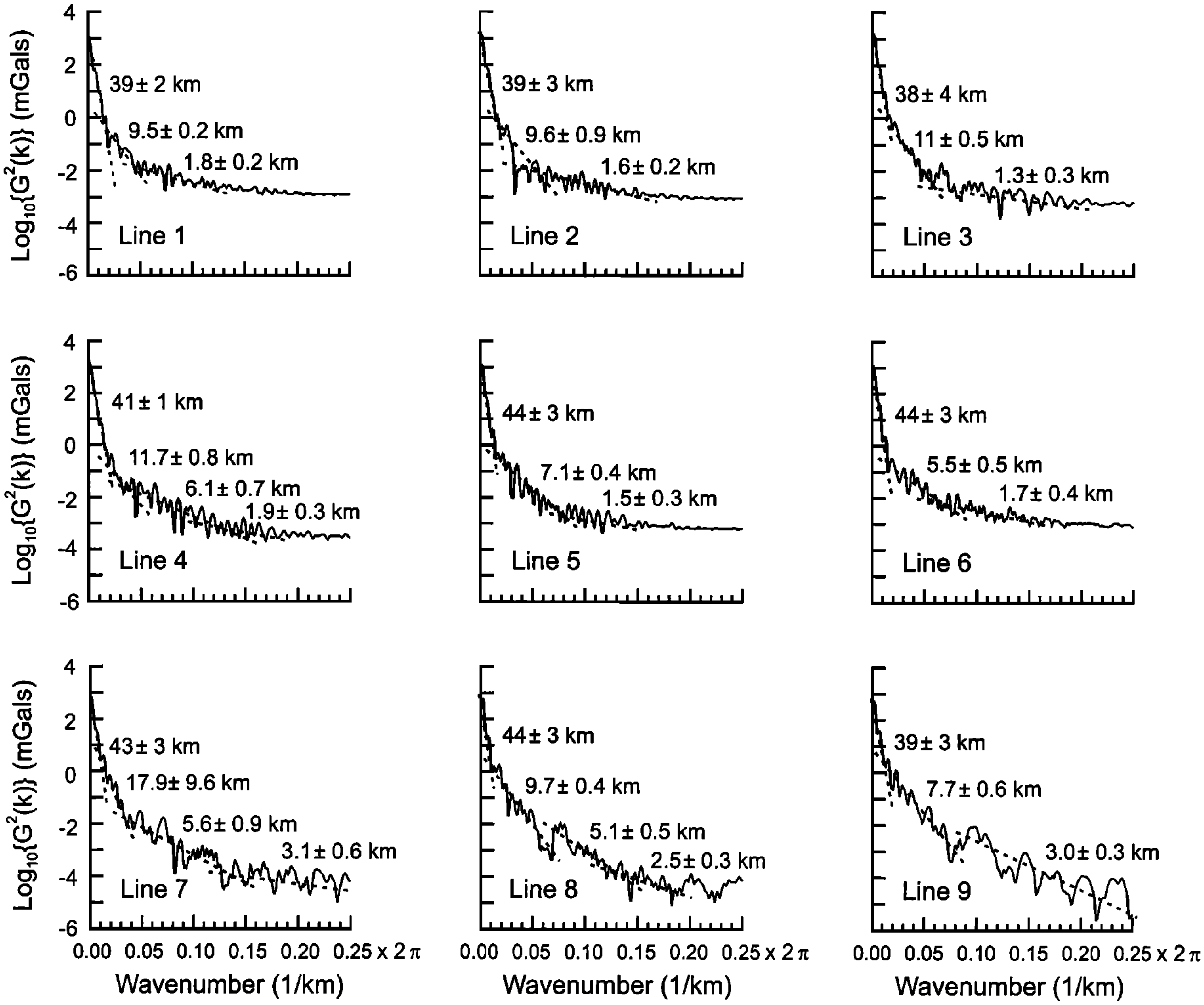

Figure 5. Bouguer gravity power spectra. Dashed lines indicate linear segments of the power spectra corresponding to density interfaces at the indicated depths.

the Bouguer gravity admittance and coherence [*Forsyth*, 1985].

Admittance is defined as the ratio of the Bouguer gravity and topography spectral estimates. The form of the admittance function depends primarily on the flexural rigidity of the lithosphere and the partitioning of the total load between the surface and subsurface [*Dorman and Lewis*, 1970; *Banks et al.*, 1977]. It is not possible to distinguish between these two factors on the basis of admittance alone, and flexural rigidity estimates obtained from admittance modeling may therefore be biased toward lower values if a subsurface load is present [*Forsyth*, 1985]. This situation can often be recognized by a pronounced peak in the admittance function at intermediate wavelengths. An independent estimate of the flexural rigidity can be obtained from the coherence between the Bouguer gravity and topography spectra [*Forsyth*, 1985]. Coherence is not very sensitive to the relative magnitudes of the Bouguer gravity and topography but is simply a measure of how well the topography and gravity correlate at various wavelengths. Coherence is less sensitive to the manner in which loading is partitioned between the surface and subsurface than the admittance as long as the surface and subsurface loads are uncorrelated and is therefore a good indicator of flexural rigidity.

Following *Forsyth* [1985], we first estimate the flexural rigidity from the coherence assuming no subsurface loading. This is accomplished by comparing the observed coherence function to theoretical curves calculated at various assumed rigidities. The estimated flexural rigidity is then used in conjunction with the admittance to estimate the ratio of surface to subsurface loading by comparing the measured admittance to theoretical curves calculated for various loading ratios. The estimated loading ratio is then used to refine the coherence model, and the procedure is repeated until the flexural rigidity and loading ratio are consistent on successive iterations. This procedure relies on the assumption that the surface and subsurface loads are uncorrelated. If the loads are correlated, as is likely in fold and thrust belts, the rigidity estimate will be

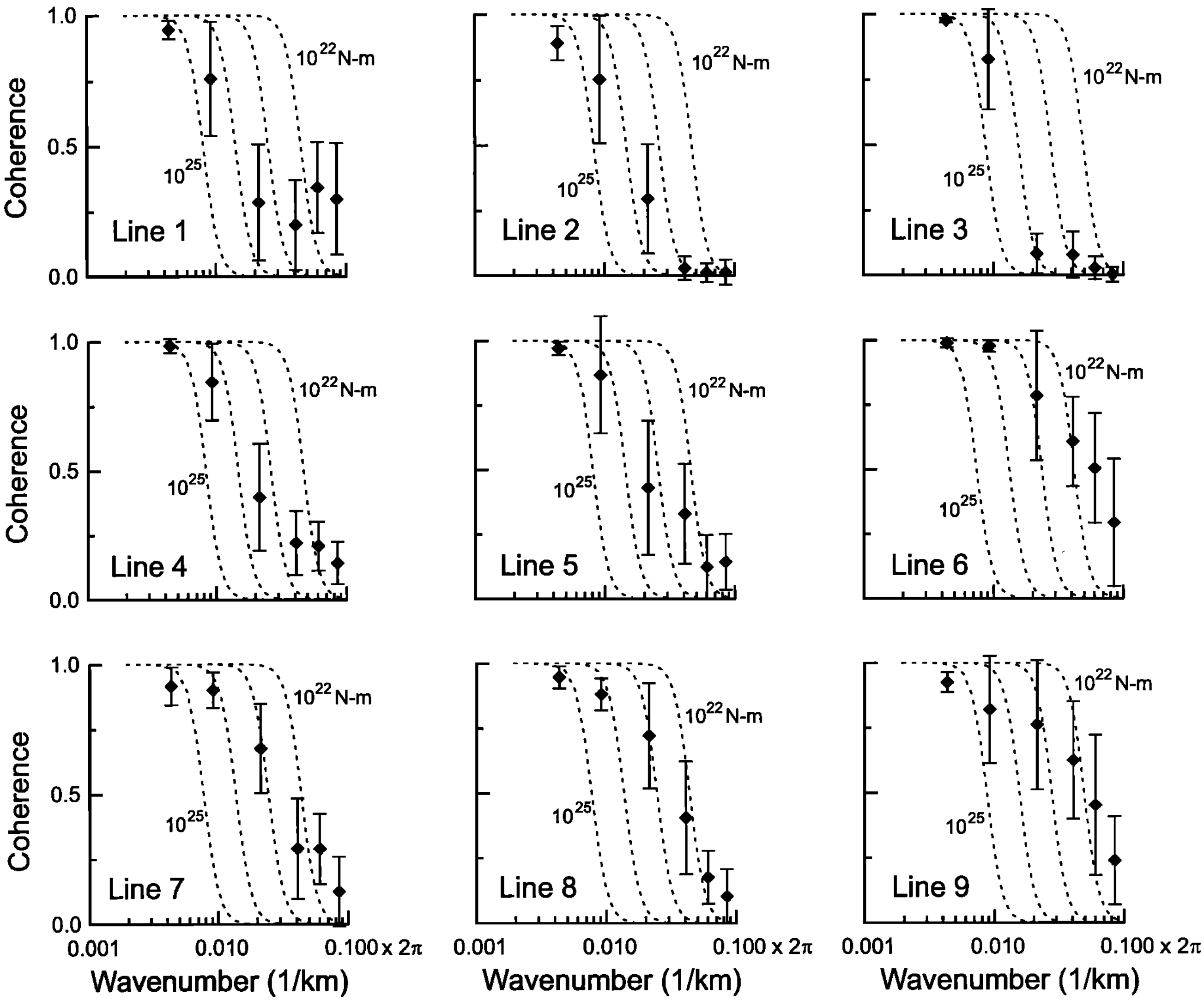

Figure 6. Coherence estimates. Diamonds indicate unbiased coherence estimates calculated using spectral averaging over a sliding window, with error bars indicating variance. Dashed lines indicate theoretical coherence for lithosphere with indicated flexural rigidities and best fitting subsurface:surface loading ratio (Table 1). See text for further discussion.

biased toward lower rigidity values [*Macario et al.,* 1995]. However, the inverse models discussed in section 5 show that the rigidity estimates obtained from the coherence allow for a flexural model that provides a good fit to the observed gravity, suggesting that the rigidity estimates are representative of the true value.

The coherence and admittance estimates are shown in Figures 6 and 7, respectively. The density interfaces used in calculating the theoretical curves were taken from the Bouguer gravity power spectra, and the density contrasts were chosen by comparison with the forward model of *Mickus and Keller* [1992]. Only well-resolved interfaces with relatively large density contrasts were used in the analysis. Experimentation with models that included all of the interfaces yielded similar results. The results are summarized in Table 1. Flexural rigidity generally decreases from west to east, from 2.0 ± 1.0 x 10^{24} N m at the location of line 1 to 5.0 ± 3.0 10^{23} N m at the location of line 9. Assuming a Young's modulus of 10^{11} Pa and Poisson's ratio of 0.25 [*Turcotte and Schubert,* 1982], these correspond to effective elastic thicknesses of 60 and 38 km, respectively. The relatively high admittance values at wavenumbers between 0.006 and 0.02 km^{-1} on profiles 2 to 4 result from the large-amplitude Bouguer gravity anomaly and subdued topography at these wavelengths. Such a pattern in the admittance function is characteristic of a relatively large subsurface component of the total load [*Forsyth,* 1985]. This is consistent with the positive Bouguer gravity anomaly seen on the southern ends of profiles 1 through 3 in the western Ouachitas (Figure 4). The positive Bouguer gravity anomaly is less pronounced in the central and eastern Ouachitas (profiles 6-8), and the admittance functions for these profiles do not display the high amplitudes at intermediate wavelengths characteristic of large subsurface loads. Thus the admittance models and gravity data both indicate a general

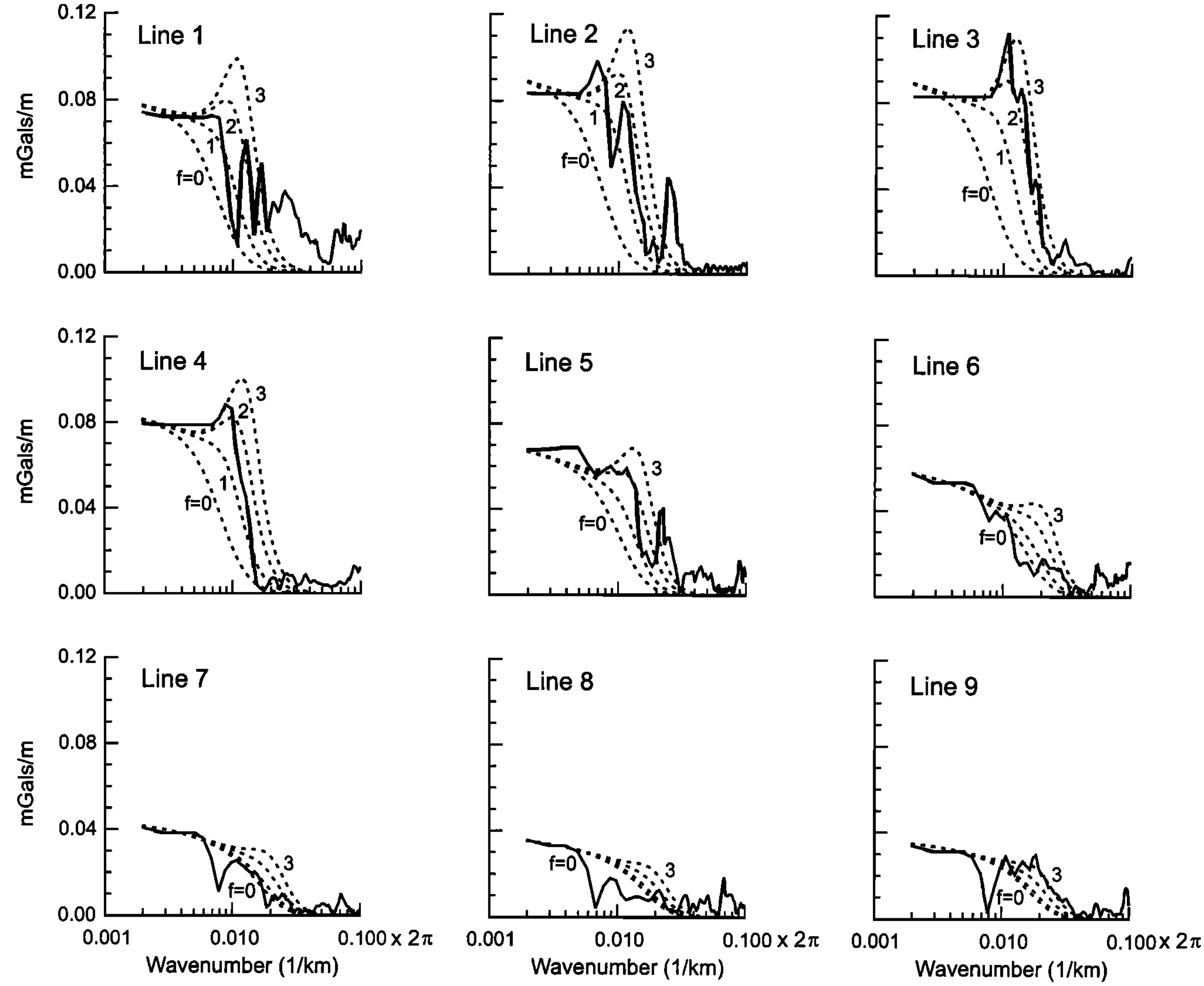

Figure 7. Admittance estimates. Solid line indicates calculated admittance using spectral averaging over discrete wavenumber bands. Dashed lines indicate theoretical admittance for indicated subsurface:surface loading ratios and best fitting flexural rigidity (Table 1). See text for discussion.

Table 1. Crustal Density Model, Rigidity, and Load Ratio

Profile	Rigidity, $\times10^{24}$ N m	Load Ratio	Base of Arkoma Basin		Base of Ouachita Facies		Base of Crust	
			Depth	Δρ, g cm^{-3}	Depth	Δρ, g cm^{-3}	Depth	Δρ, g cm^{-3}
1	2.00 ± 1.00	1.75 ± 0.25	10	0.15	-	-	39	0.23
2	1.75 ± 0.80	2.00 ± 0.25	10	0.15	-	-	39	0.23
3	1.75 ± 0.80	2.25 ± 0.15	11	0.15	-	-	38	0.23
4	1.75 ± 0.80	2.00 ± 0.25	12	0.15	6	0.06	41	0.23
5	1.25 ± 0.80	1.75 ± 0.25	-	-	7	0.06	44	0.23
6	0.50 ± 0.20	1.25 ± 0.25	-	-	6	0.06	44	0.23
7	0.50 ± 0.20	1.25 ± 0.25	-	-	6	0.06	43	0.23
8	0.50 ± 0.20	1.50 ± 0.25	10	0.15	5	0.06	44	0.23
9	0.50 ± 0.30	2.50 ± 0.50	8	0.15	-	-	39	0.23

Dash indicates interfaces that are not resolved in the gravity spectra.

eastward decrease in the ratio of subsurface to surface loading from 2.25 ± 0.15 at the location of profile 3 to 1.25 ± 0.25 at the location of profile 7 (Table 1). The loading ratio decreases slightly westward from profile 3 to 1.75 ± 0.25 at profile 1. The load ratio for profile 9 is much higher than the other profiles in the eastern Ouachitas, but the high variance in the coherence (Figure 6) leads to a large uncertainty in both the rigidity and loading ratio on this profile. Taken together, the coherence and admittance models indicate that the lithosphere is stronger beneath the western Ouachitas than beneath the eastern Ouachitas and that the subsurface load is proportionally larger.

5. Estimation of the Load

A least-squares inverse modeling method is used to estimate the magnitude and position of the point load which best approximates the excess mass emplaced in the crust during the Ouachita orogeny (see Appendix). The excess mass is indirectly constrained by the flexural deformation required to produce the observed Bouguer gravity minimum. The inverse method iteratively searches for the point load which results in a calculated flexural deformation profile that produces a Bouguer gravity anomaly similar to that which is observed. Each of the nine profiles were modeled separately, with the optimum solution for each profile being that which minimizes the squared misfit between the calculated and observed gravity data. The amount of flexural deformation is calculated from the equations governing bending of a two-dimensional elastic plate that is broken on the south end, with the load acting on the broken edge of the plate. The location of the plate's edge (and therefore the load's center of mass) and the magnitude of the point load are the unknown parameters in the inverse models. The flexural rigidity of the elastic plate is specified a priori in the inverse models and is taken from the coherence and admittance modeling (Table 1). The Bouguer gravity anomaly is calculated using *Parker*'s [1973] method, assuming a simple layered density structure for the crust. Of the five interfaces identified from the Bouguer gravity power spectra, only the interface interpreted to represent the base of the Arkoma basin clastic fill was used in the inverse modeling. It was assigned a density contrast of 0.17×10^3 kg m^{-3} by comparison with previous gravity modeling studies in the western Ouachitas [*Kruger and Keller*, 1986; *Mickus and Keller*, 1992]. On profiles 5 through 7, where this interface was not well resolved, its depth was estimated by interpolating from adjacent profiles. The density interface at the base of the crust was neglected in the inverse modeling because seismic refraction and gravity studies have shown that it is much less pronounced than the flexural subsidence of the shelf carbonate rocks that underlie the Arkoma basin. This is most likely due to Mesozoic crustal attenuation during opening of the Gulf of Mexico [*Keller et al.*, 1989b], which has resulted in subdued topography at the base of the crust along the strike of these profiles. The shallower interfaces shown in Figure 5 were interpreted to be due to Coastal Plain sediments or other features not associated with flexural deformation in the Arkoma basin and so were also neglected.

The inverse procedure attempts to fit that portion of the Bouguer gravity anomaly that is the result of flexural subsidence in the basin and uplift of the outer bulge. The positive anomaly produced by the excess subsurface mass that dominates the Bouguer gravity field south of the gravity minimum is not modeled and is not used as a constraint. Therefore only the portions of the gravity profiles lying north of the Bouguer gravity minimum (approximately the northern 200 km of the profiles) are used. However, the load that produces flexural subsidence is allowed to lie at any position along the trend of the profile and was, in fact, found to lie far south of the Bouguer gravity minimum. A spatially distributed load (rather than a point load) is obviously more realistic, but testing of synthetic models showed that the distribution of the load is poorly constrained by the gravity data in this inversion scheme. The center of mass and total mass are well constrained by the gravity data, so inverting for the best fitting point load results in a geologically meaningful solution with a stable algorithm and low estimated parameter variances. Experimentation showed that the position of the point load estimated from the inverse method does not vary greatly with realistic changes in the density contrast at the interface.

The results of the inverse models are shown in Figure 8, and the load parameters are summarized in Table 2. For all profiles, the RMS misfit between the predicted and observed gravity is less than 5 mGals. The predicted gravity systematically underestimates the observed gravity in the vicinity of the Bouguer gravity minimum and southward because the model does not include the effects of the positive anomaly produced by the excess mass. The modeled deflection of the crust reaches a maximum of about 18 km beneath the gravity minimum on profile 4 in the western Ouachitas, consistent with the thickness of clastic sedimentary rocks in the southern Arkoma basin and the frontal zone of the Ouachitas in this area [*Lillie et al.*, 1983; *Keller et al.*, 1989a; *Mickus and Keller*, 1992]. Little subsidence is predicted farther than 125 km north of this position in good agreement with the distance between the frontal zone and the northern extent of the Arkoma basin strata in Arkansas and Oklahoma. Eastward, the model predicts a monotonic decrease in the amount of deflection, reaching 4 km at profile 9 in the eastern Ouachitas. The width and depth of the modeled basin on profiles 7 through 9 are markedly less than beneath the more westerly profiles. This is reflected in the abrupt decrease in the width and amplitude of the Bouguer gravity minimum in the eastern Ouachitas (Figure 3). All of the modeled profiles have a low-amplitude outer bulge located approximately 150 to 200 km north of the gravity minimum.

6. Discussion

The Bouguer gravity power spectra indicate that the crust is thickest beneath profiles 5 through 8 in the central and eastern Ouachitas (Figure 9). Corresponding changes in the depth of the midcrustal density interface (interface 3 in Figure 9) show that some of the thickening occurs in the upper crust. This is partly due to changes in depth of the Arkoma foreland basin, which attains its maximum depth between Lines 3 and 4 (Figure 3). Changes in the thickness of the upper crust between interfaces 1 and 2 (Figure 9) are probably due to along-strike differences in synorogenic deformation of Ouachita facies rocks. Precisely how much thickening occurs in the up-

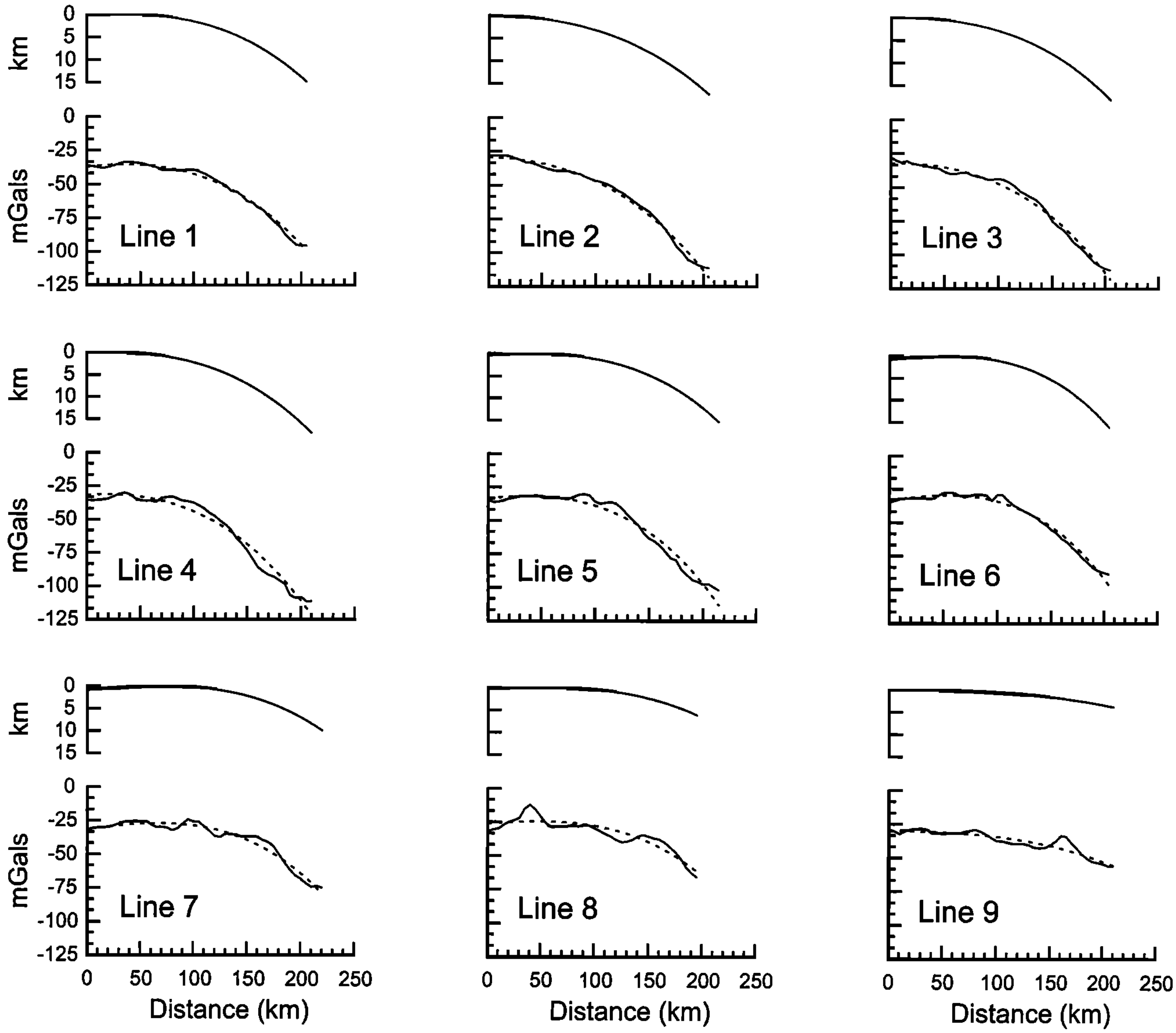

Figure 8. Results of inverse modeling. (top) Flexural deformation of the preorogenic crust and (bottom) comparison of the calculated (dashed line) and observed (solid line) gravity.

Table 2. Point Load Inversion Results

Profile	Load Magnitude, $\times 10^{13}$ N m	Load Position South of Gravity Minimum, km	RMS Error, mGals
1	12.60	276 ± 26	2
2	7.89	230 ± 22	2
3	9.68	243 ± 48	3
4	10.50	247 ± 60	5
5	6.86	207 ± 48	4
6	4.74	170 ± 47	3
7	3.67	175 ± 35	3
8	2.48	179 ± 60	5
9	0.57	114 ± 68	3

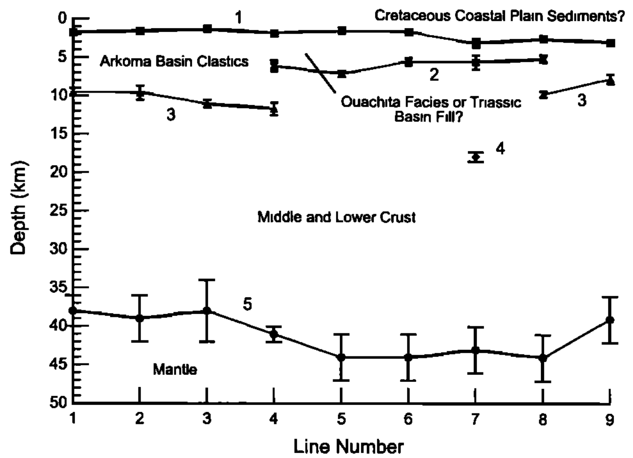

Figure 9. Depth to major density interfaces in the crust determined from Bouguer gravity spectra for each profile. Bars indicate estimated error. Interfaces are numbered corresponding to discussion in text. Depths are plotted as a function of profile number, with profile 1 on the west and profile 9 on the east. The depths represent an average interface depth across the length of each profile, so Figure 9 should not be viewed as a simple along-strike cross section. See text for discussion.

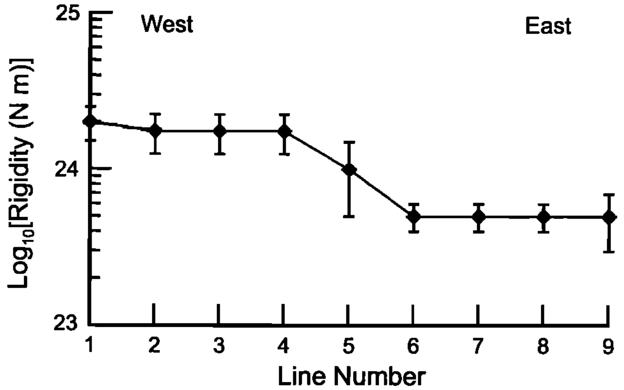

Figure 10. Flexural rigidity estimated from the coherence from each profile. Bars indicate estimated error. Profile 1 is toward the west; profile 9 is toward the east.

per crust in the central Ouachitas is unknown because interface 3 is not well resolved in the gravity data on profiles 5-7. However, on the profiles where interface 3 is resolved, it is clear that much of the variation in crustal thickness occurs at midcrustal and deeper levels. Because the depth of the Arkoma basin decreases toward the east, the thicker crust in the eastern Ouachitas cannot be attributed to a greater amount of flexural subsidence. It may instead result from along-strike variations in synorogenic deformation in the middle or lower crust, or it may simply reflect variations in the thickness of the crust that existed on the Paleozoic rifted continental margin prior to the Ouachita orogeny. Since the Ouachitas developed near the juncture of a rift segment of the Paleozoic margin and a transform segment of the margin, preexisting variations in the thickness of the crust are likely.

The eastward increase in the crust's thickness roughly correlates with changes in the flexural rigidity, which decreases abruptly at the position of profile 5 near the western end of the Benton uplift (Figure 10). The weaker lithosphere in the eastern Ouachitas is clearly indicated by the systematic shift in the coherence corner frequency toward higher wavenumbers in the more easterly profiles (Figure 6) and is also reflected in the pattern of flexural deformation in the Arkoma basin as indicated by the Bouguer gravity anomaly, which is much narrower at its eastern end (Figure 3). This eastward decrease in the strength of the lithosphere is consistent with the previously suggested structure of the Paleozoic rifted continental margin [*Viele and Thomas*, 1989; *Thomas*, 1991]. The western Ouachitas are inferred to have formed near a rift segment of the margin, with the eastern Ouachitas overlying less highly extended crust on a transform segment of the margin (Figure 1). Since the rifting episode predates the Ouachita orogeny by about 200 m.y., the lithosphere would have thermally reequilibrated. As a result, the weakest portion of the lithosphere would have been where the crust was thickest [*Kusznir and Karner*, 1985]. The mean flexural rigidity of the lithosphere in the western Ouachitas (1.8×10^{24} N m) is similar to that of the central Appalachian fold and thrust belt and is consistent with the cooling half-space model for the thermal evolution of rifted continental crust (Figure 11). Like the Ouachita orogen, the Appalachians underwent an episode

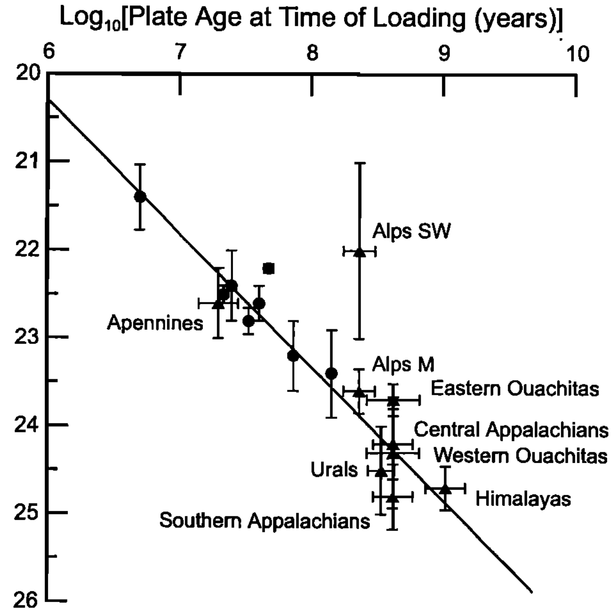

Figure 11. Flexural rigidity versus thermal age of the lithosphere at the time of loading. Triangles indicate continental compressional orogenic belts; circles indicate oceanic studies. Bars indicate estimated errors. Solid line indicates flexural rigidity if the effective elastic thickness of the lithosphere is given by the 450°C isotherm predicted by the cooling half-space model. Ouachita points are from this study. Other data are the Apennines [*Royden*, 1988], southern Applachians and Alps SW [*Stewart and Watts*, 1997], Urals [*Kruse and McNutt*, 1988], Alps M [*Macario et al.*, 1995], and central Appalachians and Himalayas [*Karner and Watts*, 1983]. Oceanic data were compiled by *Karner and Watts* [1983].

of Late Proterozoic rifting followed by a late Paleozoic collisional event, so the similarity of the flexural rigidities is not surprising. However, the rigidity of the lithosphere in the eastern Ouachitas is anomalously low for a 250 m.y. old rifted margin (Figure 11). This is attributed to the fact that the eastern Ouachitas overlie a transform continental margin rather than a rifted margin. *Stewart and Watts* [1997] make a similar argument to explain along-strike variations in the flexural rigidity of the lithosphere in the southern Appalachians, which they believe are related to the structure of the eastern North American rifted continental margin. *Stewart and Watts* [1997] argued for some form of strength recovery in rifted continental crust, noting that the lithosphere in the southern Appalachian Mountains and Ural Mountains is much stronger than that beneath orogenic belts that developed on rifted margins with a shorter elapsed time between rifting and compressional deformation. The consistency of the flexural rigidity estimates with the cooling half-space model (Figure 11) suggests that strength recovery is a result of cooling of the lithosphere following rifting, which would result in substantial strengthening of the uppermost mantle. This affect is most obvious in the Ouachita and Appalachian orogens because a relatively large amount of time elapsed between rifting and orogenesis.

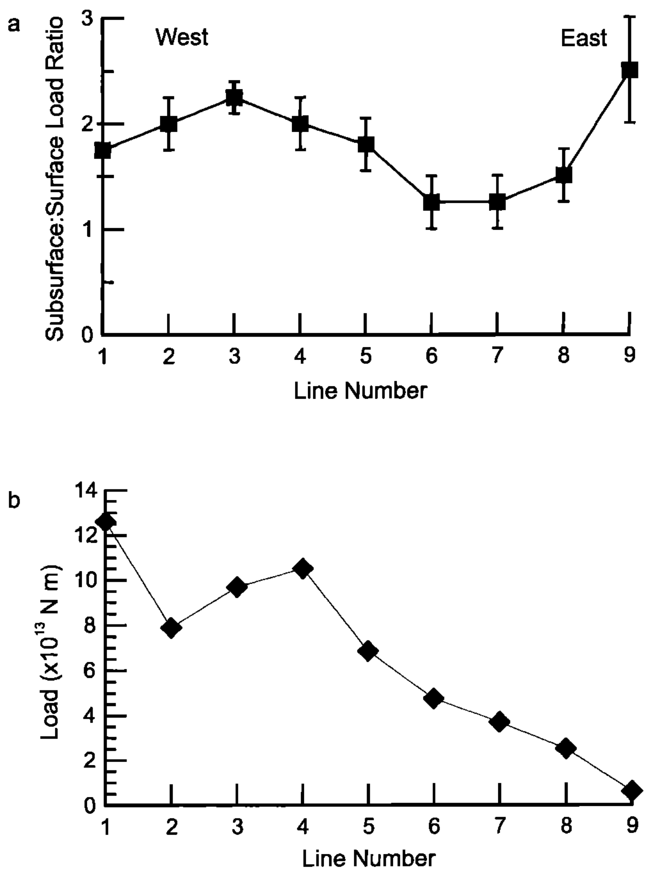

Figure 12. (a) Ratio of subsurface to surface load estimated from the admittance. Bars indicate estimated error. (b) Total uncompensated load estimated from the inverse modeling. Profile 1 is toward the west; profile 9 is towards the east.

If the flexural rigidity of the lithosphere in fold and thrust belts represents the strength of the crust just prior to thrust emplacement [*Forsyth*, 1985], the correspondence between crustal thickness and flexural rigidity changes in the Ouachitas indicates that at least some of the eastward thickening of the crust predates the Ouachita orogeny. As mentioned previously, synorogenic thickening in the upper crust, and possibly the middle to lower crust, also appears to increase toward the eastern and central Ouachitas. This may indicate more pronounced shortening in the central Ouachita province in Arkansas as compared to Oklahoma, a contention supported by the presence of overturned folds north of the Benton uplift. In contrast, shortening in the frontal Ouachita province appears greatest in the western Ouachitas. *Arbenz* [1989] suggested that the difference in the amount of apparent shortening in the frontal Ouachitas is accommodated by a zone of diffuse right-lateral shear displacement crossing the Ouachita Mountains near the western end of the Benton uplift. The Bouguer gravity maximum associated with the Broken Bow uplift abruptly terminates near this position at the Arkansas/Oklahoma border (Figure 3). The proposed shear zone may allow the intense shortening in the frontal Ouachitas of Oklahoma to be transferred to more intense shortening in the central Ouachita province in Arkansas. The location of the shear zone proposed by *Arbenz* [1989] coincides with the abrupt change in flexural rigidity at the position of profile 5 (Figure 10), which is interpreted to approximately mark the boundary between rift and transform segments of the Paleozoic continental margin. These associations suggest that preexisting weaknesses in the lithosphere arising from differences in crustal thickness served to focus contractional deformation in the eastern Ouachitas, as has been demonstrated by numerical modeling studies of contractional deformation [*Harry et al.*, 1995]. Apparently, along-strike variations in deformation over relatively short distances in fold and thrust belts may be closely related to the preexisting structure of the underlying crust.

With the exception of profile 9, the ratio of subsurface to surface loading determined from the admittance modeling is lower beneath the eastern Ouachitas than the western Ouachitas (Figure 12a). This is evidenced by the decrease in the amplitude of the peak in the admittance at intermediate wavenumbers (Figure 7). The eastward decrease in the loading ratio may be due to either a decrease in the amount of the load in the subsurface, an increase in the magnitude of the surface load, or an eastward decrease in both the surface and subsurface loads, with the decrease being most pronounced in the subsurface. The latter scenario, involving an eastward decrease in the total load, is supported by the inverse modeling (Figure 12b) and is consistent with the fact that the maximum thickness of clastic sedimentary rocks in the Arkoma basin decreases east of profile 5 in spite of the smaller flexural rigidity on the eastern side of the basin. The need for a larger buried load in the western Ouachitas is evident from the positive Bouguer gravity anomaly in the vicinity of the Broken Bow uplift (Figure 3). Furthermore, the topographic expression of the Ouachita Mountains is narrower and more subdued west of profile 4, indicating a smaller surface load in this region. The amplitudes of both the Bouguer gravity minimum and the more southerly gravity maximum decrease eastward, consistent with less flexural subsidence and a diminished subsurface load. These observations could indicate that the excess mass emplaced during the Ouachita orogeny is greatest to the west, which is in agreement with regional cross sections that show involvement of a thicker section of Ouachita facies rocks in the Broken Bow uplift than in the Benton uplift [*Arbenz et al.*, 1989]. Alternatively, the load emplaced on the North American crust during the Ouachita orogeny may have been similar along strike of the Ouachitas, with erosion having removed a larger portion of the load in the eastern Ouachitas. However, metamorphic grade and the thermal maturity of exposed rocks suggest little difference in the amount of exhumation between the eastern and western Ouachitas [*Houseknecht and Mathews*, 1985; *Keller et al.*, 1985]. A third possibility is that the eastward decrease in the load magnitude and the ratio of subsurface to surface loading simply reflects the increasing buoyancy of the thicker crust. The variations in crustal thickness discussed previously and the presence of overturned folds north of the Benton uplift support the argument for more intense deformation in the central and eastern Ouachitas, which presumably could be associated with emplacement of a greater mass of allocthonous material. Associated synorogenic thickening of the middle and lower crust might partially balance the excess mass. In this interpretation, the reduced strength of the lithosphere in the eastern Ouachitas allows for a larger degree of Airy style isostatic compensation, leaving a relatively small mass to be compensated by flexural deformation. This is not inconsistent with the admittance or inverse modeling, which are sensitive only to the uncompensated portion of the subsurface mass.

The location of the best fitting point load determined from the inverse models is located 207 to 276 km south of the Bouguer gravity minimum in the western Ouachitas and 114 to 179 km south of the gravity minimum in the eastern Ouachitas. The load position follows a trend that generally parallels the Broken Bow and Benton uplifts and the southwesterly trending positive Bouguer gravity anomaly associated with their subsurface continuation into east Texas. In the western Ouachitas, the load lies immediately north of the Sabine uplift along the Texas/Louisiana border (Figure 1), which has previously been interpreted to be formed in part by emplacement of allocthonous material near the southern edge of the Laurentian plate during the Ouachita orogeny [*Viele and Thomas*, 1989; *Mickus and Keller*, 1992]. The association of the load with the broken edge of the Laurentian plate is an assumption in the inverse modeling, but the location of the load and the edge of the Laurentian plate were not specified a priori. Therefore the coincidence of the north edge of the Sabine uplift with the estimated position of the load lends credence to the interpretation that the Sabine uplift lies near the southern edge of the Paleozoic Laurentian margin, or possibly an attached remnant of Paleozoic oceanic crust, and represents the southern limit of Ouachita facies in the subsurface [*Keller et al.*, 1989a, *Mickus and Keller*, 1992]. The southerly position of the load's center of mass indicates that the Broken Bow and Benton uplifts do not represent the surface expression of the primary load. Instead, they are probably best viewed as foreland features that involve more distal facies than those in the Arkoma basin and frontal Ouachitas.

On all of the profiles, the maximum topographic relief is less than 0.8 km and the width of the Ouachitas is less than about 150 km. Assuming a density of 2.6×10^3 kg m^{-3} for the surface rocks [*Mickus and Keller*, 1992], topography is unlikely to account for a total load greater than about 3×10^{12} N m^{-1}. The total load estimated from the inverse modeling ranges from about 1×10^{14} N m^{-1} in the western Ouachitas to 5.6×10^{12} N m^{-1} at the east end of the study area. This indicates a subsurface to surface load ratio of 30:1 or greater in the western Ouachitas, much larger than the loading ratios determined from the admittance modeling (Figure 7 and Table 2). Several factors contribute to the discrepancy. First, the admittance modeling as formulated here is based on the assumptions that the subsurface load lies at the base of the crust and that its spatial distribution is not correlated with the topography. In reality, the subsurface load may be correlated with topography (this is particularly true in fold and thrust belts), and it may be distributed at different levels throughout the crust. Second, and more importantly, the admittance modeling is only capable of estimating the loading ratio within the region traversed by the gravity profile. As the inverse models show, the major subsurface load in the western Ouachitas actually lies far south of the southern terminus of the profiles, and so the two methods are not measuring the same load. In the eastern Ouachitas, the point load is estimated to lie near the end of the profiles. The magnitude of the point load in this region is similar to that of the surface load in agreement with the low subsurface to surface load ratio in this region. Most of the load on the crust can be accounted for by surface topography in the eastern Ouachitas, with the subsurface becoming increasingly important toward the west.

The results discussed above support an interpretation in which the subsurface load is primarily attributed to emplacement of an excess thickness of Ouachita facies rocks onto highly attenuated continental or transitional crust on the southern edge of the Laurentian margin (Figure 13). This interpretation is consistent with the subsurface density model developed by *Mickus and Keller* [1992] on the basis of forward gravity modeling and with the structural interpretation of the PASSCAL seismic data reported by *Keller et al.* [1989a]. The Ouachita facies rocks are not particularly dense, so there is no significant positive gravity anomaly associated with the load. In this interpretation, the flexural shape of the Arkoma basin is viewed as a relict of Paleozoic tectonism and is not associated with a large modern subsurface load.

7. Summary

Analysis of Bouguer gravity power spectra indicates that the crustal thickness beneath the Ouachitas ranges from 38 to 44 km, with the thicker crust lying beneath the eastern Ouachitas. A major density contrast interpreted to indicate the base of Arkoma basin clastic fill lies at average depths of 7.7-11.7 km. Additional density interfaces are resolved at 5.1-7.7 km (interpreted as either the base of Ouachita facies rocks in the central Ouachitas or the base of sediments in one or more Triassic rift basins located in southern Arkansas and northern Louisiana) and 1.3-3.1 km (interpreted as either the base of Cretaceous sediments south of the Ouachitas or a pre-

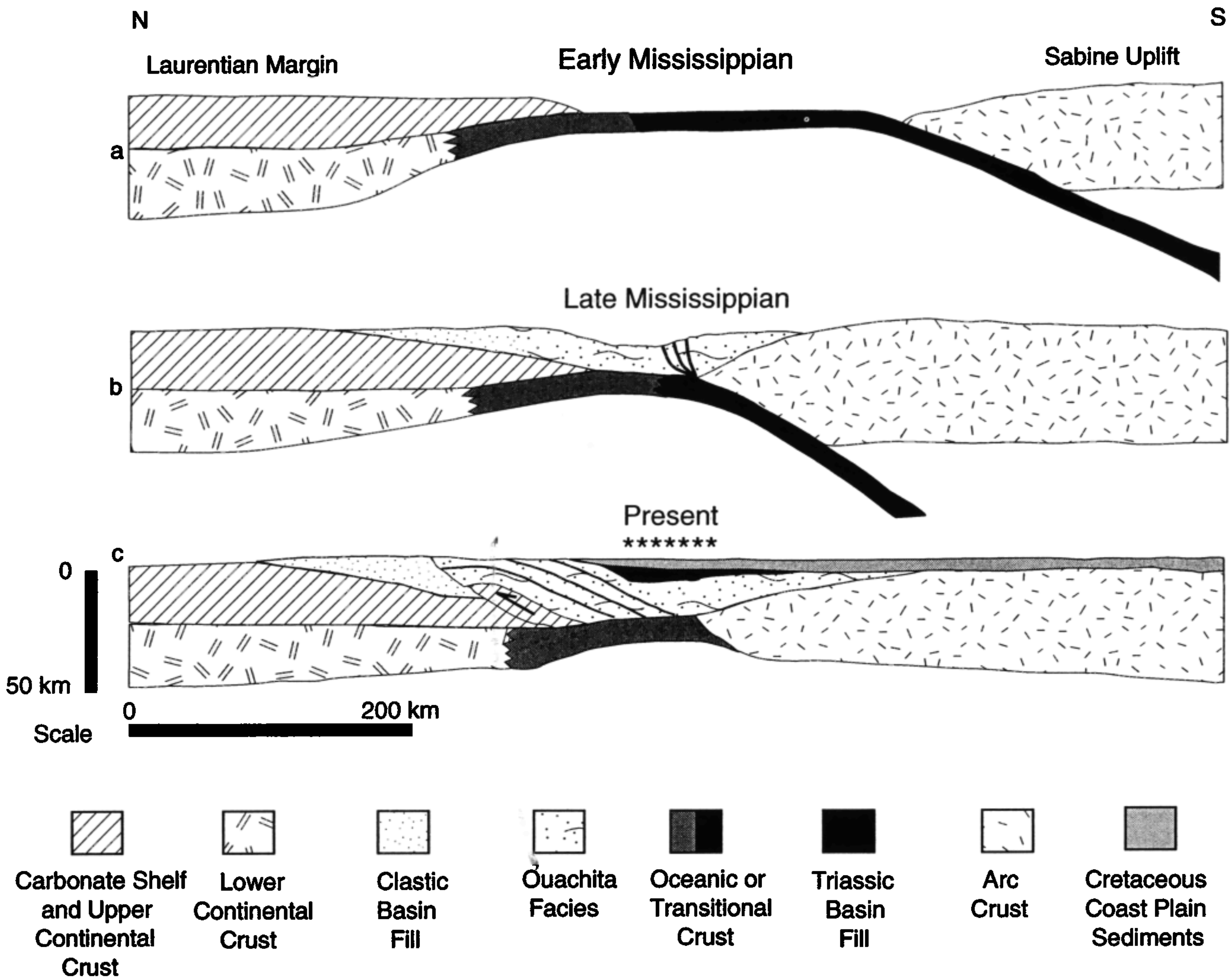

Figure 13. Schematic cross sections illustrating evolution of the Ouachita orogen in western Arkansas and northern Louisiana. (a) Early Mississippian. Deposition of deep water Ouachita facies rocks in narrow oceanic basin between Sabine Arc and southern Laurentian passive continental margin. Subduction polarity is conjectural. (b) Late Mississippian. Deposition of clastic sediments in Arkoma basin and deep water clastic Ouachita facies rocks record subsidence of southern Laurentian carbonate shelf and onset of Ouachita orogeny. (c) Present structure, modified from cross sections by *Mickus and Keller* [1992], *Roberts* [1994], and G.W. Viele and J.K. Arbenz [*Arbenz et al.*, 1989]. Asterisks indicate the approximate range of locations for the point loads estimated from the inverse modeling of each gravity profile as discussed in the text. The Sabine uplift is interpreted as the remnant of the Paleozoic island arc, with Ouachita facies strata emplaced by thrusting on the northern edge of the arc and the southern edge of the Paleozoic passive continental margin.

viously unrecognized density interface within the Arkoma basin clastic section). Bouguer gravity coherence and admittance indicates that the flexural rigidity of the lithosphere in the Ouachitas is $2.0 \pm 1.0 \times 10^{24}$ N m in the western Ouachitas and $5.0 \pm 3.0 \times 10^{23}$ N m in the eastern Ouachitas. The eastward decrease in flexural rigidity is attributed to the thicker crust in the eastern Ouachitas and probably indicates a transition from a rift segment of the early Paleozoic continental margin beneath the western Ouachitas to a transform segment of the margin beneath the eastern Ouachitas. The change in flexural rigidity occurs abruptly near the western end of the Benton uplift and is coincident with a previously proposed zone of distributed shear [*Arbenz*, 1989] that may allow for right-lateral displacement between the eastern and western Ouachitas. This accommodates differences in shortening on either side of the shear zone, with the eastern Ouachitas undergoing more shortening in the central Ouachita province and the western Ouachitas undergoing more shortening in the frontal imbricate zone. The association of the change in flexural rigidity and crustal thickness with the location of the proposed shear zone and cross-strike differences in shortening implies a close association between the structure of the early Paleozoic continental margin and synorogenic tectonism in the Ouachita Mountains.

Eastward decreases in the magnitude of the flexurally compensated load and the ratio of subsurface to surface loading within the Ouachitas may also be attributed to changes in crustal thickness, with the thicker crust in the eastern Ouachitas partially compensating the excess mass emplaced during the Ouachita orogeny. Alternatively, the mass emplaced during the orogeny may have varied significantly along strike, with a greater mass emplaced in the western Ouachitas than in the eastern Ouachitas. Flexural modeling indicates that the primary load responsible for subsidence in the Arkoma basin lies 207 to 276 km south of the Broken Bow uplift in the western Ouachitas and 114 to 179 km south of the Benton uplift in the eastern Ouachitas. The position of the load in the western Ouachitas coincides with the northern edge of the Sabine uplift and is inferred to mark the southern extent of autocthonous North American crust and allocthonous Ouachita facies emplaced during the orogeny.

Appendix

The deflection of a broken elastic plate subjected to a vertical end load at position x_0 is

$$w(x - x_0) = w_0 e^{-(x-x_0)/\alpha} \cos[(x - x_0)/\alpha] \qquad \text{(A1)}$$

where

$$w_0 = \frac{P_0 \alpha^3}{2D} \qquad \alpha = \left[\frac{4D}{(\rho_m - \rho_f)g}\right]^{1/4}.$$

Deflection is positive downward, P_0 is the force exerted by a point load at the broken edge of the plate, D is the flexural rigidity, g is the acceleration of gravity, ρ_m is the mantle density and ρ_f is the density of the sediment filling the basin [*Turcotte and Schubert*, 1982]. The Bouguer gravity spectrum generated by a density contrast $\Delta\rho$ at depth z within the flexurally deformed lithosphere is determined using the method of *Parker* [1973]:

$$G(k) = 2\pi\gamma\Delta\rho e^{-kz} \sum_{n=1}^{4} \frac{k^{n-1}}{n!} W^n(k), \qquad \text{(A2)}$$

where $W(k)$ is the deflection spectrum and k is the wavenumber. Multiple density interfaces within the flexed lithosphere are dealt with individually, and their spectra are summed to get the total Bouguer gravity spectra. The Bouguer gravity anomaly in the spatial domain is then calculated by performing an inverse Fourier transform on the net Bouguer gravity spectrum. We use (A1) and (A2) to calculate the Bouguer gravity anomaly for a flexed plate with flexural parameter α subjected to a vertical load P_0 at x_0, which marks the broken edge of the plate.

The forward-modeling method described above was adapted for use with a linear least squares inverse method to determine the magnitude and position of the point load that produces flexural deformation that best fits the observed Bouguer gravity anomaly. An attempt was made to simultaneously estimate the flexural parameter α, but the inverse algorithm was found to be unstable when this third unknown was included. Therefore the flexural parameter was specified a priori using the rigidity estimates obtained from the coherence modeling. The inverse algorithm is based on the general linear inverse method for overdetermined problems. Given a vector containing initial estimates of the model parameters $\mathbf{b}=[P_0^{est}, x_0^{est}]$ and a vector of Bouguer gravity observations at discrete points $\mathbf{d}=[g_1, g_2, g_3, \ldots, g_N]^T$, a perturbation parameter vector $\delta\mathbf{b}=[\delta P_0, \delta x_0]^T$ can be calculated to determine how the parameter estimates must be revised in order to minimize the square error between the calculated and observed Bouguer gravity anomalies. The solution is obtained from the linear system of equations

$$\delta\mathbf{b} = [\mathbf{A}^T\mathbf{W}\mathbf{A}]^{-1}\mathbf{A}^T\mathbf{W}\mathbf{d}, \qquad \text{(A3)}$$

where $\mathbf{W}$ is a diagonal weighting matrix whose elements are the variance of the data and $\mathbf{A}$ is the matrix of partial derivatives describing the dependence of the gravity misfit on the parameter perturbation vector,

$$\mathbf{d}\text{-}\mathbf{d}^{est} = \mathbf{A}\delta\mathbf{b}. \qquad \text{(A4)}$$

The revised estimate of the model parameters is given by $\mathbf{b}=\mathbf{b}+\delta\mathbf{b}$, and the procedure is iterated until the modeled and observed gravity agree to within a specified error or the change in parameter estimates is less than a specified value. The vector of parameter variances is given by

$$\mathbf{S}_b = \mathbf{H}^T\mathbf{H}\mathbf{W}, \qquad \text{(A5)}$$

where $\mathbf{H}$ is the generalized inverse of $\mathbf{A}$.

Experimentation showed that stable and accurate solutions for synthetic gravity models were obtained if the elements of the weighting matrix $\mathbf{W}$ are specified by the variance of the gravity observations. This also has the effect of nondimensionalizing the data vector. The parameters should also be nondimensionalized by weighting them by an estimate of their final values ($1\text{x}10^{12}$ N m and 1 km were used as weights for the load magnitude and position, respectively). The parameter variance (A5) was then multiplied by the parameter weights to get the variance in the proper dimensions. The components of $\mathbf{A}$ were determined numerically by solving the forward problem with two parameter choices: P_0^{est} and $P_0^{est}/2$ and x_0^{est} and $x_0^{est}/2$. The derivatives were then approximated by the differences $\delta g_b/\delta P_0$ and $\delta g_b/\delta x_0$, where $\delta P_0 = P_0^{est}/2$, $\delta x_0 = x_0^{est}/2$, and δg_b is the difference in the modeled gravity at the two choices of parameter values. Comparison with synthetic models showed that this simple method was robust and stable for a wide range of flexural parameters, loading scenarios, and initial parameter estimates.

Acknowledgments. The gravity and topography data were provided by G.R. Keller of the University of Texas at El Paso. This research was partially supported by grants to D.L.H. from The University of Alabama School of Mines and Energy Development and the American Chemical Society Petroleum Research Fund (grant 30919-G2). We thank Sarah Kruse, Bob Lillie, Carolyn Ruppel, and Bill Thomas for insightful discussions and advice.

References

Arbenz, J.K., Ouachita thrust belt and Arkoma basin, in *The Geology of North America*, vol. F-2, *The Appalachian-Ouachita Orogen in the United States*, edited by R.D. Hatcher, W.A. Thomas, and G.W. Viele, pp. 621-634, Geol. Soc. of Am., Boulder, Colo., 1989.

Arbenz, J.K., W.R. Muehlberger, R.L. Nicholas, P.R. Tauvers, and G.W. Viele, Ouachita system cross-sections, in *The Geology of North America*, vol. F-2, *The Appalachian-Ouachita Orogen in the United States*, edited by R.D. Hatcher, W.A. Thomas, and G.W. Viele, plate 11, Geol. Soc. of Am., Boulder, Colo., 1989.

Banks, R.J., R.L. Parker, and S.P. Huestis, Isostatic compensation on a continental scale: Local versus regional mechanisms, *Geophys. J. R. Astron. Soc.*, *51*, 431-452, 1977.

Blythe, A.N., A. Sugar, and S.P. Phipps, Structural profiles of Ouachita mountains, western Arkansas, *Am. Assoc. Pet. Geol. Bull.*, *72*, 810-819, 1988.

Denison, R.E., Foreland structure adjacent to the Ouachita foldbelt, in *The Geology of North America*, vol. F-2, *The Appalachian-Ouachita Orogen in the United States*, edited by R.D. Hatcher, W.A. Thomas, and G.W. Viele, pp. 681-688, Geol. Soc. of Am., Boulder, Colo., 1989.

Dorman, L.M., and B.T.R. Lewis, Experimental isostasy, 1, Theory of the determination of the Earth's isostatic response to a concentrated load, *J. Geophys. Res.*, *75*, 3357-3365, 1970.

Elmore, R., P.K. Sutherland, and P. White, Middle Pennsylvanian recurrent uplift of the Ouachita fold belt and basin subsidence in the Arkoma basin, Oklahoma, *Geology*, *18*, 906-909, 1990.

Flawn, P.T., A.J. Goldstein, P.B. King, and C.E. Weaver, The Ouachita system, *Publ. 6120*, 401 pp., Bur. Econ. Geol., Univ. Tex. at Austin, 1961.

Forsyth, D.W., Subsurface loading and estimates of the flexural rigidity of continental lithosphere, *J. Geophys. Res.*, *90*, 12,623-12,632, 1985.

Harry, D.L., J.S. Oldow, and D.S. Sawyer, Stress and strain partitioning in continental margin orogenic belts: Implications of dynamic modeling studies, *Geol. Soc. Am. Bull.*, *107*, 1411-1426, 1995.

Houseknecht, D.W., Evolution from passive margin to foreland basin: The Atoka formation of the Arkoma basin, south-central U.S.A., in *Foreland Basins*, edited by P.A. Allen and P. Homewood, *Spec. Publ. Int Assoc. Sedimentol.*, *8*, 327-345, 1986.

Houseknecht, D.W., and S.M. Matthews, Thermal maturity of Carboniferous strata, Ouachita Mountains, *AAPG Bull.*, *69*, 335-345, 1985.

Karner, G.D., and A.B Watts, Gravity anomalies and flexure of the lithosphere at mountain ranges, *J. Geophys Res.*, *88*, 2923-2948, 1983.

Keller, G.R., L.W. Braile, G.A. McMechan, W.A. Thomas, S.H. Harder, W.-F. Chang, and W.G. Jardine, Paleozoic continent-ocean transition in the Ouachita mountains imaged from PASSCAL wide-angle seismic reflection-refraction data, *Geology*, *17*, 119-122, 1989a.

Keller, G.R., J.M. Kruger, K.J. Smith, and W.M. Voight, The Ouachita system: A geophysical overview, in *The Geology of North America*, vol. F-2, *The Appalachian-Ouachita Orogen in the United States*, edited by R.D. Hatcher, W.A. Thomas, and G.W. Viele, pp. 689-724, Geol. Soc. of Am., Boulder, Colo., 1989b.

Keller, W.D., C.G. Stone, and A.L. Hoersch, Textures of Paleozoic chert and novaculite in the Ouachita Mountains of Arkansas and Oklahoma and their geologic significance, *Geol. Soc. Am. Bull.*, *96*, 1353-1363, 1985.

Kruger, J.M., and G.R. Keller, Interpretation of regional gravity anomalies in the Ouachita mountains area and adjacent gulf Coastal Plain, *AAPG Bull.*, *70*, 667-689, 1986.

Kruse, S., and M.K. McNutt, Compensation of Paleozoic orogens: A comparison of the Urals to the Appalachians, *Tectonophysics*, *154*, 1-17, 1988.

Kusznir, N., and G. Karner, Dependence of the flexural rigidity of the continental lithosphere on rheology and temperature, *Nature*, *316*, 138-142, 1985.

Lillie, R.J., K.D. Nelson, B. deVoogd, J.A. Brewer, J.E. Oliver, L.D. Brown, S. Kaufman, and G.W. Viele, Crustal structure of Ouachita mountains, Arkansas: A model based on integration of COCORP reflection profiles and regional geophysical data, *AAPG Bull.*, *67*, 907-931, 1983.

Loomis, J., B. Weaver, and H. Blatt, Geochemistry of Mississippian tuffs from the Ouachita Mountains and implications for the tectonics of the Ouachita orogen, Oklahoma and Arkansas, *Geol. Soc. Am. Bull.*, *106*, 1158-1171, 1994.

Macario, A., A. Malinverno, and W F. Haxby, On the robustness of elastic thickness estimates obtained using the coherence method, *J. Geophys. Res.*, *100*, 15,163-15,172, 1995.

Mickus, K.L., and G.R. Keller, Lithospheric structure of the south-central U.S., *Geology*, *20*, 335-338, 1992.

Milliken, J.V., Late Paleozoic and early Mesozoic geologic evolution of the Arklatex area, M.S. thesis, 223 pp., Rice Univ., Houston, Tex., 1988.

Mooney, W., and C. Weaver, Regional crustal structure and tectonics of the Pacific coastal states: California, Oregon and Washington, in *Geophysical Framework of the Continental United States*, edited by L. Pakiser and W. Mooney, *Mem., Geol. Soc. Am.*, *172*, 129-162, 1989.

Morris, R.C., Stratigraphy and sedimentary history of post-Arkansas Novaculite Carboniferous rocks of the Ouachita Mountains, in *The Geology of North America*, vol. F-2, *The Appalachian-Ouachita Orogen in the United States*, edited by R.D. Hatcher, W.A Thomas, and G.W. Viele, pp. 591-602, Geol. Soc. of Am., Boulder, Colo., 1989.

Nelson, K.D., R.J. Lillie, B. de Voogd, J.A. Brewer, J.E. Oliver, S. Kaufman, and L. Brown, COCORP seismic reflection profiling in the Ouachita mountains of western Arkansas: Geometry and geologic interpretations, *Tectonics*, *1*, 413-430, 1982.

Nielsen, K.C., G.W. Viele, and J. Zimmerman, Structural setting of the Benton-Broken Bow uplifts, in *The Geology of North America*, vol. F-2, *The Appalachian-Ouachita Orogen in the United States*, edited by R.D. Hatcher, W.A. Thomas, and G.W. Viele, pp. 635-660, Geol. Soc. of Am., Boulder, Colo., 1989.

Niem, A.R., Mississippian pyroclastic flow and ash-fall deposits in the deep-marine Ouachita flysch basin, Oklahoma and Arkansas, *Geol. Soc. Am. Bull.*, *88*, 49-61, 1977.

Parker, R.L., The rapid calculation of potential anomalies, *Geophys. J. R. Astron. Soc.*, *31*, 347-355, 1973.

Rankin, D.W., Appalachian salients and recesses: Late Precambrian continental breakup and opening of the Iapetus Ocean, *J. Geophys. Res.*, *81*, 5605-5619, 1976.

Roberts, M.T., Geologic relations along a regional cross section from Spavinaw to Broken Bow, eastern Oklahoma, in *Geology and Resources of the Eastern Ouachita Mountains Frontal Belt and Southeastern Arkoma Basin, Oklahoma*, edited by N.H. Suneson and L.A. Hemish, Guideb., Okla. Geol. Surv., *29*, 137-160, 1994.

Royden, L., Flexural behavior of the continental lithosphere in Italy: Constraints imposed by gravity and deflection data, *J. Geophys. Res.*, *93*, 7747-7766, 1988.

Stewart, J., and A.B. Watts, Gravity anomalies and spatial variations of flexural rigidity at mountain ranges, *J. Geophys. Res.*, *102*, 5327-5352, 1997.

Sutherland, P.K., Late Mississippian and Pennsylvanian depositional history in the Arkoma basin area, Oklahoma and Arkansas, *Geol. Soc. Am. Bull*, *100*, 1787-1802, 1988.

Taylor, S., Geophysical framework of the Appalachians and adjacent Grenville province, in *Geophysical Framework of the Continental United States*, edited by L. Pakiser and W. Mooney, *Mem., Geol. Soc. Am.*, *172*, 317-348, 1989.

Thomas, W.A., Evolution of the Ouachita-Appalachian continental margin, *J. Geol.*, *84*, 323-342, 1976.

Thomas, W.A., The Appalachian-Ouachita rifted margin of southeastern North America, *Geol. Soc. Am. Bull.*, *103*, 415-431, 1991.

Thomas, W.A., Low-angle detachment geometry of the late Precambrian-Cambrian Appalachian-Ouachita rifted margin of southeastern North America, *Geology*, *24*, 921-924, 1993.

Turcotte, D.L., and G. Schubert, *Geodynamics: Application of Continuum Physics to Geological Problems*, 255 pp., John Wiley, New York, 1982.

Viele, G.W., and W.A. Thomas, Tectonic synthesis of the Ouachita orogenic belt, in *The Geology of North America*, vol. F-2, *The Appalachian-Ouachita Orogen in the United States*, edited by R.D. Hatcher, W.A. Thomas, and G.W. Viele, pp. 695-728, Geol. Soc. of Am., Boulder, Colo., 1989.

D.L. Harry, Department of Geology, University of Alabama, Box 870338, Tuscaloosa, AL, 35487-0338. (email: dharry@wgs.geo.ua.edu)

K.L. Mickus, Department of Geosciences, Southwest Missouri State University, Springfield, MO 65804. (email: mickus@cart1.smsu.edu)

(Received May 13, 1997;
revised December 12, 1997;
accepted December 31, 1997.)