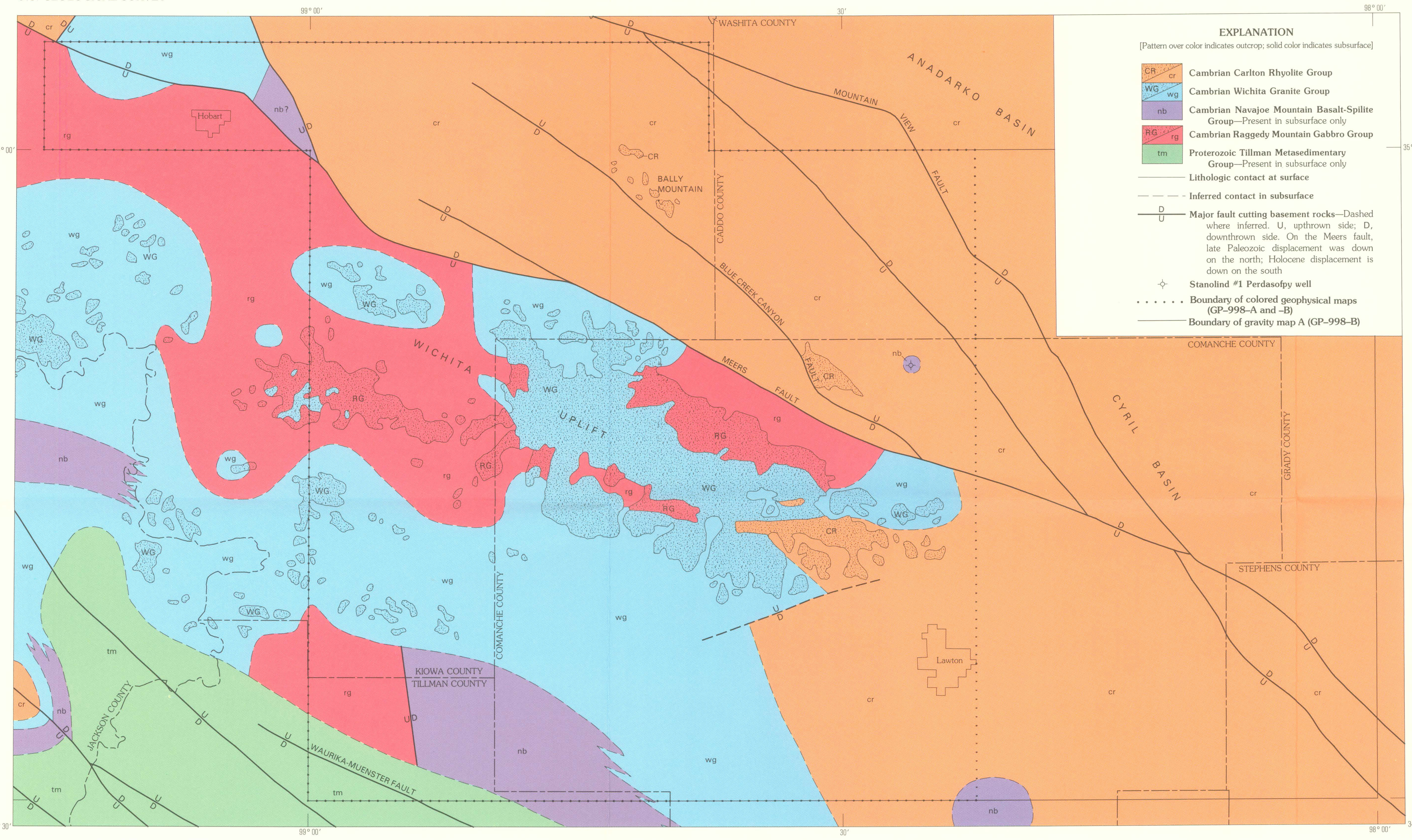

Figure 1. Map showing basement geology and areas covered by the aeromagnetic maps (GP–998–A) and the gravity maps (GP–998–B) relative to the surface and subsurface rocks interpreted from well data (after Ham and others, 1964). Location of the Stanolind #1 Perdasofpy well is shown with rocks of the Navajoe Mountain Basalt-Spilite Group indicating presence of this group below the Carlton Rhyolite Group. ESRI's ARC/INFO was used to digitize parts of the geology and faults from plate 1 of Ham and others (1964). The digitized data were fitted using ARC/INFO's "rubber sheeting" to a Universal Transverse Mercator projection and enlarged to a scale of 1:250,000. The detail and accuracy of the data are less than mapping at 1:250,000 would produce.

INTRODUCTION

These maps are the first of two sets (GP–998–A and GP–998–B) that display available aeromagnetic and gravity data for the region around the eastern Wichita Mountains and the "Frontal Wichita fault system" (a network of faults that separates the igneous rocks of the Wichita uplift from the sedimentary rocks of the Anadarko basin). GP–998–A includes a map showing igneous rocks and locations of the aeromagnetic maps as well as the gravity maps of GP–998–B (fig. 1), a map showing localities of susceptibility measurements (fig. 2), a table showing magnetic susceptibility measurements by rock type and locality (table 1), an aeromagnetic map mathematically reduced to the North Pole (map A), and a map showing horizontal gradients derived from the aeromagnetic map (map B). The area of maps A and B includes most of the 1:100,000, 30 × 60-minute Lawton quadrangle, the southwestern part of the adjacent Anadarko quadrangle, and the southeastern part of the adjacent Elk City quadrangle. GP–998–B, the companion group of maps, displays the gravity field anomalies.

GEOLOGIC SETTING

From south to north, the study area encompasses a small part of the Burch-Waurika-Muenster fault system (which separates the Hollis-Hardeman basin, located west-southwest of the area shown in figure 1, from the Wichita uplift), the eastern Wichita Mountains (composed of a bimodal suite of igneous rocks within the Wichita uplift), and a part of the more than 300-km-long (186-mi) Frontal Wichita fault system (fig. 1). The Hollis-Hardeman basin contains approximately 3 km (1.9 mi) of Paleozoic sedimentary rocks overlying 7–10 km (4–6 mi) of Proterozoic clastic sedimentary and felsic volcanic rocks interpreted from deep seismic reflection profiling (Brewer and others, 1981). The Anadarko basin, the deepest basin in the continental United States, contains at least 12 km (7 mi) of Paleozoic sedimentary rocks (Ham and Wilson, 1967).

Many authors have related the structural evolution of the Wichita Mountains and the adjacent Anadarko basin to the development of the southern Oklahoma aulacogen (Hoffman and others, 1974; Burke, 1977; Wickham, 1978; and Gilbert, 1983, to mention some of the earlier authors). Deposition and extrusion of rocks in the Hollis-Hardeman basin (perhaps beginning with deposition of the Tillman Metasedimentary Group (Hoffman and others, 1974)) may have been the earliest evidence of the crustal extension initiating the southern Oklahoma aulacogen (Brewer and others, 1981; Perry, 1989). Farther north, in the area of the present-day Wichita uplift, basic igneous rocks were emplaced during Early to Middle Cambrian rifting followed by acidic igneous extrusion and intrusion. The exact location of the rift boundaries is unclear, though Ham and others (1964) suggested that the northern edge coincides with the Washita valley fault in the vicinity of the Arbuckle uplift, east of the eastern Wichita Mountains, and implied that it coincides with the Meers fault in the area of the Wichita Mountains. Subsidence caused by cooling of the previously rifted crust resulted in the deposition of a thick section of Upper Cambrian through Lower Mississippian sedimentary rocks. Perry (1989) distinguished between the southern Oklahoma aulacogen and the Anadarko basin. In his view, continental collision caused most of the later subsidence of the basin; only the southernmost and deepest part of the Anadarko basin was formerly part of the southern Oklahoma aulacogen. Reactivation of old fault trends, uplift of the Wichita Mountains, growth of the Anadarko basin, and development of the ancestral Rocky Mountains occurred in Late Mississippian through Pennsylvanian time, apparently due to continental collision associated with the Ouachita orogeny.

The amount of strike-slip displacement on the northwest-trending faults of the aulacogen that accompanied the late Paleozoic collision is controversial. For example, Budnik (1987) proposed as much as 120 km (75 mi) left-lateral displacement in Pennsylvanian time along a megashear that may extend to the northwest as far as Utah and encompasses the Frontal Wichita fault system. Tanner (1967) proposed 64 km (40 mi) of left-lateral offset along the Washita valley fault, but Perry (1989) proposed approximately 5 km (3 mi) of left-lateral offset. McConnell (1989) used palinspastic restoration of isopachs across the Mountain View fault, within the Frontal Wichita fault system in the area of the present report, to estimate the ratio of reverse to left-lateral slip, and obtained between 1.1:1 for a west-northwest-trending segment and 4.3:1 for a northwest-trending segment.

Holocene reactivation of at least 26 km (16 mi) of the Meers fault (Ramelli and Slemmons, 1986) is the most recent element known in the structural history of the area. Holocene movement on the Meers fault consisted of both left-lateral strike-slip and normal dip-slip movement, the north side up (Crone and Luza, 1990). The Meers fault, trending approximately N. 60° W., is particularly noticeable on the aeromagnetic maps as one of the most prominent, steep, linear gradients in the map area.

The igneous rocks of the Wichita uplift and the Frontal Wichita fault system are the primary source rocks for the magnetic and gravity anomalies shown on the maps in this series. A description of their ages and history follows as an aid to interpreting the maps. Gabbro bodies of the Raggedy Mountain Gabbro Group were intruded into the upper crust during rifting of the southern Oklahoma aulacogen along the southern margin of the North American craton beginning around 577±165 Ma (Lambert and others, 1988). The Glen Mountains Layered Complex, which is the lowest formation within the Raggedy Mountain Gabbro Group, is petrologically similar to parts of the Bushveld, Skaergaard, and Stillwater Complexes, all gabbroic layered intrusions (Powell, 1986). The Roosevelt Gabbros, a younger formation within the Raggedy Mountain Gabbro Group, intruded the Glen Mountains Layered Complex. The Roosevelt Gabbros were probably derived from parent magma different from that of the layered complex (Powell, 1986). The Navajoe Mountain Basalt-Spilite Group, thought to be an extrusive equivalent of the Raggedy Mountain Gabbro Group, has been identified in drill cores from areas north and south of the outcropping Wichita Mountains (Ham and others, 1964). A brief period of erosion followed the emplacement of basic igneous rocks. The Carlton Rhyolite Group was extruded at approximately 525±25 Ma (Ham and others, 1964) over much of the area. Sills of the Wichita Granite Group were intruded along the contact between the Carlton Rhyolite Group and the eroded surface on the Raggedy Mountain Gabbro Group. In the final rift-related igneous activity, diabase dikes intruded all preceding igneous units. The diabase may be a small, shallowly intruded part of a much larger underlying mafic body associated with rifting (Coffman and others, 1986; Gilbert and Hughes, 1986).

GEOLOGIC BASE MAP

The igneous units and faults indicated in figure 1 and on maps A and B were digitized from the approximately 1:360,000 basement rock map of Ham and others (1964). The Navajoe Mountain Basalt-Spilite Group is shown around the Stanolind #1 Perdasofpy well (approximately 8 km (5 mi) north of the junction of the Blue Creek Canyon and Meers faults), where more than 305 m (1,000 ft) of spilite and basalt were found underlying Carlton Rhyolite Group and a granite sill (Ham and others, 1964). The digitized map units and faults were enlarged to fit the 1:250,000 map in figure 1 and the 1:100,000 physiographic and demographic data shown on maps A and B by means of "rubber sheeting" within the Geographic Information System (GIS) program ARC/INFO, developed by ESRI, Inc. Because the original basement rock map of Ham and others (1964) was generalized at less than one-third the scale of maps A and B, interpretation based on exact location of boundaries of rock units or locations of faults would be invalid. For example, the width of the fault lines drawn on the Ham and others (1964) map introduces as much as 200 m (656 ft) variability in fault location on the 1:100,000 map.

TOTAL-FIELD AEROMAGNETIC MAP, REDUCED TO THE POLE

The data for the aeromagnetic map were originally gathered by a U.S. Geological Survey 1954 total-magnetic-field airborne survey, flown 152 m (500 ft) above the ground along east-west flight lines 0.40 km (0.25 mi) apart. North of lat 35° N., the flight-line spacing increased to as much as 1.6 km (1 mi). These data were published as a contour map (U.S. Geological Survey, 1975), which was digitized and is reprinted here as a colored map. A small part of the original data for the area west of long 99° 15′ W. was omitted due to inaccuracy in the published contour map (presence of mismatched contours). By means of a Summagraphics digitizer and an unpublished digitizing program (VTDIGIT, by R.H. Godson), the intersection of the contours and flight lines were digitized and converted into latitude, longitude, and magnetic-intensity values. These values were gridded as 0.13411- by 0.13411-km (0.08336- by 0.08336-mi) squares, using a minimum-curvature contouring program (MINC, Webring, 1981). Because the inclination of the present geomagnetic field in the study area is 64°, highs, lows, and gradients of anomalies generally do not directly overlie their source bodies: the magnetic anomaly generated by an equidimensional rock body magnetized in the direction of the present-day field consists of a magnetic high and a low of lesser magnitude immediately to the north (see, for example, fig. 55, Nettleton, 1971). By mathematically translating the data to the magnetic North Pole (inclination of 90°), anomalies are shifted to lie more directly over their sources. The program FFTFIL (Hildenbrand, 1983), which incorporates a discrete Fourier transform algorithm, was used to reduce the data to the North Pole. The transformed data were then plotted using Scitex Corporation color software and equipment.

Reducing the magnetic data to the magnetic North Pole can simplify anomalies if the rock magnetization is parallel to the present geomagnetic field at the site. This is true if the magnetization is caused solely by induced magnetization. However, paleomagnetic studies have shown that components of both primary and secondary remanent magnetization, which need not point in the direction of the present field, are present in many of the igneous rocks of the Wichita Mountains (Ku and others, 1967; Spall, 1968, 1970; Vincenz and others, 1975; and Roggenthen and others, 1981). The Koenigsberger ratio (Q), which is the ratio of the intensity of the remanent magnetization to the magnetic susceptibility times the intensity of the Earth's field, gives an indication of the relative contribution of the remanent magnetization to the observed total-field magnetization. Ku and others (1967) measured *Q* for 18 igneous samples from the Wichita Mountains, including gabbro, granite, and rhyolite. Many samples showed *Q* values greater than 1.0, the highest value being 226. The directions of natural remanent magnetization (NRM) for these samples and for samples from other studies show a great deal of scatter (see discussion in Vincenz and others, 1975) and, as a result, cannot be used to remove an NRM direction for the entire map area during reduction to the North Pole. Thus, although the asymmetric highs and lows over simple source geometries and the displacement of gradients caused by induced magnetization have been removed in map A, some of the anomalies may still be shifted relative to their source bodies or show paired high and low anomalies due to remanent magnetization or geometric factors.

MAP SHOWING MAXIMUM HORIZONTAL GRADIENT DERIVED FROM AEROMAGNETIC DATA

The boundaries between adjacent bodies of contrasting magnetization can produce large horizontal gradients in components of the magnetic field, if favorably oriented relative to the magnetic field direction. Within the Frontal Wichita fault system many bodies having different magnetizations have been structurally juxtaposed. By calculating the horizontal gradient, the location of faults and intrusive boundaries can be inferred in favorable circumstances. Map B was generated using the technique of Cordell and Grauch (1985) and program BOUNDARY described in Blakely and Simpson (1986). The technique requires transformation of the digital total-field aeromagnetic data into pseudogravity data, calculation of the horizontal gradient from the pseudogravity data, and location of linear ridges along the maxima of the gradient. The pseudogravity transformation, calculated by means of Fourier transforms, converts the magnetic field into the gravity field that would be observed if magnetic material within the source rocks were replaced by the same proportion of dense material. To emphasize that the transformation is based on the assumption, rather than the observation, of exact proportionality between susceptibility and density, the density units are labelled "pseudo-milligals" on map B.

In using the program BOUNDARY, several options had to be chosen. As was the case in reducing the magnetic data to the North Pole, only induced magnetization was assumed. BOUNDARY calculates the maxima of horizontal gradients by comparing a grid value to all adjacent grid values. A grid value must exceed adjacent grid values in two, three, or four directions to be flagged as being close to a maximum value. The actual value and location of the maximum are then determined by interpolation and plotted by computer (Blakely and Simpson, 1986). Lines of contiguous maxima at least 2.0 km (1.2 mi) in length were selected from these plots and are shown on map B as black lines superimposed on the horizontal gradients.

Areas of high horizontal gradients, shown in warmer colors on map B, help to define boundaries between bodies of contrasting pseudo-density, and therefore of contrasting magnetization. Because the horizontal gradient is greater for abrupt changes in field, shallow features are most enhanced by this method. Known faults, such as the Waurika-Muenster, the Meers, and the Blue Creek Canyon faults, are in regions of high horizontal gradient. The map also shows many other structural features or intrusive boundaries that are not mapped from surface geology. The most striking of these features are considered in the discussion section of this report.

DISCUSSION

This set of aeromagnetic maps gives a detailed view of a part of the southern Oklahoma aulacogen, where its magnetic and gravity signature changes in trend and width. Regional geophysical maps that cover the area of this map set include the "Magnetic Anomaly Map of North America" (Committee for the Magnetic Anomaly Map of North America, 1987), "Gravity Anomaly Map of North America" (Gravity Anomaly Map Committee, 1987), "Gravity Anomaly Map of the United States" (Society of Exploration Geophysicists, 1982), "Complete Bouguer and Isostatic Residual Gravity Maps of the Anadarko Basin, Wichita Mountains, and Surrounding Areas, Oklahoma, Kansas, Texas, and Colorado" (Robbins and Keller, 1992), "Vertical-Intensity Magnetic Map of Oklahoma" (Jones and Lyons, 1964), "Bouguer Gravity-Anomaly Map of Oklahoma" (Lyons, 1964), and the total-field aeromagnetic surveys by the U.S. Department of Energy (1976a–e) conducted as part of the National Uranium Resource Evaluation (NURE). Please note the errors near Clinton, Okla., pointed out by Purucker (1986) on the NURE maps. As shown on all of these regional maps in the area southeast of the eastern Wichita Mountains, the magnetic and gravity high associated with the southern Oklahoma aulacogen trends in a northwesterly direction. In the vicinity of the eastern Wichita Mountains, the high widens and separates into two parallel highs. Farther to the northwest, the two parallel highs merge, narrow, and change trend to a more westerly direction. The 0.40-km (0.25-mi) flight-line spacing of the aeromagnetic data, color enhancement, reduction to the North Pole, and calculation of horizontal gradients shown on maps A and B provide detail not visible on the regional maps and (or) difficult to see on the original contoured map (U.S. Geological Survey, 1975). For this report, discussion will be limited to a qualitative overview of the study area.

The total-field aeromagnetic map (map A) reflects major geologic features within the map area. From southwest to northeast, these features include a small part of the Burch-Waurika-Muenster fault system, the eastern Wichita Mountains, and the Frontal Wichita fault system. Specific anomalies within the map area, labelled "A"– "I" on maps A and B, are discussed in greater detail below. The steepest gradient within the map area (map B, southwest corner) is in the approximate location of the Waurika-Muenster fault, one of the northern boundary faults of the Hollis-Hardeman basin. This steep gradient is probably caused by juxtaposition of rocks of contrasting susceptibility due to faulting and (or) intrusion. Farther to the northeast, the area of the eastern Wichita Mountains is generally characterized by high magnetic values apparently having near-surface sources as indicated by the short-wavelength features showing a marked northwest trend. The Frontal Wichita fault system, between the Wichita Mountains and the Anadarko basin, is bounded on the south by the almost-linear Meers fault and contains broad, elongate magnetic highs, showing virtually none of the short-wavelength features characteristic of the Wichita uplift. The exposed rocks within the Frontal Wichita fault zone are primarily Cambrian and Ordovician carbonates with minor outcrops of Carlton Rhyolite Group northeast of the Blue Creek Canyon fault zone. The lack of short-wavelength features similar to those seen in the Wichita uplift may indicate buried source rocks or source rocks of different character. The presence of magnetic source rocks at depth within the Frontal Wichita fault zone must be accounted for in any interpretation of the fault zone. Finally, approaching the Anadarko basin in the vicinity of the Mountain View fault and in the area of what Harlton (1972) termed the Cyril basin, the magnetic character becomes one of a very broad low.

In order to interpret specific anomalies, it is necessary to consider magnetic susceptibility, remanent magnetization, and geometry of the source rocks. Previous paleomagnetic studies have shown a great variability among measurements of remanence from the igneous rocks within the Wichita uplift, and consequently no single correction can be made for remanent magnetization. Subsequently, remanent magnetization would need to be determined and taken into account for each anomaly. Geometry of source bodies can be surmised only in a general fashion without modeling. Because potential field models are not unique, even they provide only constrained best guesses without further supporting data. However, many of the highly magnetic rocks associated with the initial rifting stages of the aulacogen are exposed in the Wichita Mountains.

Table 1 shows the results of direct measurements of the magnetic susceptibility of some of the source rocks. The Roosevelt Gabbros and the undifferentiated gabbros in the eastern part of the Glen Mountains Layered Complex have the highest magnetic susceptibility values of exposed rocks sampled in the eastern Wichita Mountains. Their median values (table 1) are within the range of susceptibilities of basic igneous rocks shown in figure 14–14 of Dobrin (1976), the values from localities 4, 5, and 6 being near the upper end of the range. The magnetic susceptibilities of zones L, M, and N of the Glen Mountains Layered Complex are approximately one order of magnitude lower and in the lower end of the range of values for basic igneous rocks in figure 14–14 of Dobrin (1976). The average susceptibility of the Carlton Rhyolite Group is less magnetic than the average susceptibility of zones L–N of the Glen Mountains Layered Complex and slightly less than the average susceptibility value for acid igneous rocks (Dobrin, 1976). The Wichita Granite Group has the greatest variability of susceptibility values, but overall, the lowest susceptibilities of the measured igneous rocks, more than an order of magnitude lower than the average value for acid igneous rocks reported by Dobrin (1976). The coarser grained Quanah Granite has the lowest susceptibility values of the localities measured within the Wichita Granite Group (table 1). Finally, the Cambrian-Ordovician sedimentary rocks and the overlying Permian sedimentary rocks are near or below the threshold of the susceptibility meter and can be considered nonmagnetic.

Though, in general, the magnetic susceptibilities of table 1 reflect the magnetic highs and lows shown on map A, there are exceptions. The high-susceptibility, unassigned Glen Mountains Layered Complex of site 10 (fig. 2) is in a magnetically low area. In this case, magnetic remanence probably dominates the total magnetization and is of different orientation than the Earth's magnetic field. Paleomagnetic measurements on rocks near site 10 show an NRM of reversed polarity, easterly declination, and shallow inclination (Roggenthen and others, 1981). Elsewhere, the Raggedy Mountain Gabbro Group does not correspond with magnetic highs as expected from susceptibility measurements; NRM directions within the Raggedy Mountain Gabbro Group include normal and mixed polarities of southwest declination and shallow inclination (Roggenthen and others, 1981) that probably dominate the total magnetization of these rocks. Another inconsistency between table 1 and map A occurs in the area of fine-grained granitic rocks (fig. 2). Magnetic susceptibility sites 2 and 10 have relatively high magnetic susceptibilities. However, the remaining sites in fine-grained granitic rock would not predict the magnetic highs associated with these rocks in map A. This could be caused by inhomogeneities within the granitic rocks or surface weathering that produced low magnetic susceptibility readings. In addition, residual topographic effects contribute to the association of granitic rocks with short-wavelength magnetic highs. Topographic highs typically occur within the granitic rocks, and, although the magnetic survey flight lines were draped 152 m (500 ft) above the ground, the total-field measurements will include the magnetization of adjacent topographic highs. This effect is not seen in the less magnetic coarse-grained granitic rocks.

Magnetic high A is immediately north of the Waurika-Muenster fault. Wells drilled in this area encountered gabbros of the Raggedy Mountain Gabbro Group (Ham and others, 1964) below the exposed Permian sediments. In comparison to areas where the Raggedy Mountain Gabbro Group is exposed, this anomaly is both larger and of greater magnitude. This difference may be due to greater concentration of magnetic minerals within this particular section of the gabbro group, a different remanent magnetization, or a larger and (or) thicker intrusive body. Comparison of magnetic high A with gravity high A on maps B and C in GP–998–B shows that the magnetic signature of the buried body is displaced to the south relative to the gravity high. This displacement may be caused by a large component of remanent magnetization; the low immediately north of the magnetic high suggests the pairing of high and low anomalies commonly arising from remanent magnetization that is not parallel to the Earth's magnetic field. Other reasons for the displacement of the magnetic anomaly relative to the gravity anomaly could be that the body is inhomogeneous in concentration of magnetic minerals or in density distribution, or that there is an underlying dense body and (or) feeder dike (for example, see model in Coffman and others, 1986). Significantly wider spacing of gravity stations relative to magnetic flight-line spacing results in a more generalized anomaly shape. Data from more closely spaced gravity stations might lessen the displacement of the gravity anomaly relative to the magnetic anomaly. Ham and others (1964) described two wells immediately east of the gravity and magnetic high that penetrated basaltic lavas of the Navajoe Mountain Basalt-Spilite Group. One of these two wells also penetrated anorthosite of the Raggedy Mountain Gabbro Group at a depth of 386 m (1,265 ft). Ham and others (1964) suggested that a north-trending fault separates the gabbro bodies from the basaltic lava (fault shown on figure 1 and on geologic base map for maps A and B). The eastern boundary of anomaly A is quite sharp and is near this proposed fault. However, the orientation of the magnetic gradient suggests that the fault (or possible intrusive boundary) trends northeasterly. The northern boundary of anomaly A is a very steep gradient identified as a ridge of high gradient values by program BOUNDARY and shown as a solid black line on map B. Either an intrusive boundary between bodies of contrasting magnetization or a fault could be the cause of this steep gradient.

Magnetic anomaly B is similar in several respects to anomaly A: surface exposures are of a rock type of lesser magnetic susceptibility (the Wichita Granite Group) relative to other rock types in the map area; wells in the area of anomaly B penetrated gabbros of the Raggedy Mountain Gabbro Group (Ham and others, 1964); the anomaly is larger and of higher magnitude than areas where the gabbro group is exposed, including exposures in the Glen Mountains immediately south of this anomaly; and it is displaced relative to the gravity high shown on the maps in GP–998–B. The displacement of the magnetic anomaly relative to the gravity anomaly is more pronounced than in the case of anomaly A. A reverse component of remanent magnetization could contribute to the relative displacement. Again, the displacement might be somewhat lessened by denser spacing of gravity stations. Inhomogeneities within, and possible asymmetry of, the source body could explain the relative displacement. An alternate cause could be a separate body at a depth that is the source of the gravity anomaly but has lower magnetic mineral content or is deep enough to produce little or no magnetic signature. An aeromagnetic survey by Cooper and Crebs (1986), flown 122 m (400 ft) above ground level at 0.4- to 0.8-km (0.25- to 0.5-mi) spacing between flight lines oriented N. 30° E., covers a small part of the areas of maps A and B, in the area of anomaly B and farther to the southeast. They proposed several possible explanations for anomaly B: a significant magnetic mineral component within a large gabbro body, a part of the Glen Mountains Layered Complex, an ultramafic body, a funnel-shaped intrusion, placer deposits derived from the Glen Mountains Layered Complex, or a unit within the Roosevelt Gabbros. A gravity high in the area implies a source that is dense as well as highly magnetic and suggests a large, thick source body rather than a source such as a placer deposit.

Anomalies C and D are lows within the Wichita uplift. Anomaly C is relatively featureless compared with anomaly D (map A) and covers a broader area. The Permian Post Oak conglomerate is the primary rock type exposed over anomaly C. Outcrops of the Quanah Granite, a coarser grained formation within the Wichita Granite Group having low magnetic susceptibility, are limited to the northern edge of anomaly C (see table 1 and fig. 2). As a low-susceptibility body (0.00054 SI), it could contribute to anomaly C if it extends south of the outcrop area in the subsurface. Absence, thinning, or greater depth to the gabbros having higher magnetic susceptibility and greater density all could be possible causes of both low anomalies. Small outcrops of finer grained formations within the Wichita Granite Group occur in the lower part of anomaly D, probably causing some of the short-wavelength components. In addition, there is an arcuate magnetic high generally trending northwest to southeast that extends through anomaly D and passes approximately 2 km (1.2 mi) north of the point indicated by "D" on map A. Cooper and Crebs (1986) interpret this arcuate high to be a steeply dipping intrusion of the Cold Springs Breccia or Cold Springs Granite. The zoned Glen Mountains Layered Complex, having the lowest measured susceptibility of the Raggedy Mountain Gabbro Group, crops out between this arcuate magnetic high and the high of anomaly B. South of the point indicated by "D" there is a west-northwest linear alignment of highs that appears to continue in a much more subdued form into anomaly C. South of the linear alignment, the magnetic field is higher in magnitude. The gravity field, however, shows anomaly D to be relatively featureless and higher in magnitude than the gravity low of anomaly C. Some of the smoothness of the gravity anomaly is due to wider spacing of stations. As an illustration of the effect of flight-line spacing on magnetic maps, U.S. Department of Energy (1976a–e) NURE mapping of the same area at a 4.8-km (3-mi) flight-line spacing shows anomaly D as a featureless low compared to the detail visible on map A. However, the gravity field may also reflect deeper sources and (or) sources more homogeneous in density than in distribution of magnetic minerals.

The line defined by E–E′ is a trend of anomalies delineating a boundary. Local highs occur along this boundary, but more notable is the change in character across the boundary (map A). Both large magnetic highs (anomalies A and B) are on the west side of the boundary. Both highs correspond with highs having subdued expression on the east side of the boundary. The low of anomaly D is reflected as a stronger low east of the boundary (anomaly C). The change in magnitude of the highs and lows across this boundary could be caused by downdropping of the eastern source rocks relative to the west. The gravity maps of GP–998–B do not show this boundary in the same fashion as the magnetic maps. In particular, anomaly B is at its widest in the north-south dimension along the line defined by E–E′ and is elongated along this line to the south. Similar to the decrease in magnitude of the magnetic expression of these anomalies east of the boundary, the extension of the gravity highs defined by anomalies A and B are of lower magnitude east of the boundary. If it is a fault boundary, there is either intrusion of a dense body along the fault to produce the gravity high or the fault boundary dips to the east. An easterly dipping fault could result in a magnetic high that reflects the near-surface location of the source body and a gravity high that reflects the deeper, more easterly location of the deeper part of the source body. Previously discussed linear and arcuate magnetic features that cross line E–E′ show no apparent offset across the boundary, implying that if E–E′ is a fault boundary it has not had significant strike-slip movement or that the linear and arcuate features post-date any strike-slip movement. The boundary defined by line E–E′ is directly in line with, and immediately south of, the proposed southern extension of the midcontinent gravity high (Robbins and Keller, 1992). If the midcontinent gravity high extends this far south, it could have provided a zone of weakness along E–E′ that could have been reactivated during Early Paleozoic rifting or Late Paleozoic structural inversion of the southern Oklahoma aulacogen. In addition, Coffman and others (1986) suggested a major fault along the northern part of this boundary trending in a slightly more northwesterly direction.

Anomaly F–F′, along with lesser magnetic highs labelled G, H, and I on maps A and B of GP–998–A are within the Frontal Wichita fault system. The magnetic susceptibility of exposed rhyolite is too low to reasonably be the source of the observed highs. Below about 792 m (2,600 ft) of rhyolite, the Stanolind #1 Perdasofpy well (fig. 1) penetrated a granite sill and more than 305 m (1,000 ft) of spilite and basalt of the Navajoe Mountain Basalt-Spilite Group (Ham and others, 1964). The basalt and spilite or their intrusive equivalents may be the source of these observed anomalies. Jones-Cecil and Crone (1989) modeled the buried source as a broadly folded and faulted body of susceptibility 0.029 SI (within the range of the Raggedy Mountain Gabbro Group samples in table 1). Anomalies G, H, and I suggest anticlinal fold axes, where the magnetic rocks would be closer to the surface. However, the model of Jones-Cecil and Crone (1989) was able to produce a theoretical field across anomaly I that was simply a marked decrease in dip of the magnetic source body. More modeling and information from drill holes is necessary to resolve the nature of these lesser anomalies. Gravity stations contributing data for gravity maps of GP–998–B are not close enough together to define anomalies F–F′, G, H, and I in the detail seen on the aeromagnetic maps. There is, however, an apparent widening of the gravity gradient along this part of the Frontal Wichita fault zone that could indicate a source of denser rocks at depth than are exposed at the surface. The location of the Blue Creek Canyon fault, taken from Ham and others (1964) and shown in figure 1 and the geologic base map for maps A and B, deviates from the location of the maximum magnetic gradient between Bally Mountain and the juncture of the Blue Creek Canyon and Meers faults; the aeromagnetic maps suggest a more complex geometry for the fault in this area.

As pointed out previously, the Meers fault and its extension to the northwest form one of the most striking linear features shown on these maps. The program BOUNDARY identified much of this gradient as continuous segments of a narrow high (map B). Without extensive modeling and additional evidence, the orientation of the Meers fault at depth is uncertain. Interpretation of a deep seismic-reflection profile west of the Holocene scarp of the Meers fault and immediately west of the boundary labelled E–E′ shows the Meers fault dipping moderately to the southwest (Brewer and others, 1983). The linearity of the fault trace in the vicinity of the Holocene offset and simple models in Jones-Cecil and Crone (1989) across and immediately adjacent to the Holocene offset, though not unique, suggest a nearly vertical dip of the Meers fault. A possible barrier limiting the length of Holocene rupture of the Meers fault is the apparent splaying of the fault northwest of the northwest end of the Holocene fault scarp, approximately 10 km (6 mi) east of the point indicated by "E" on maps A and B. The apparent splaying is particularly noticeable on the map showing maximum horizontal gradient (map B) and appears to have carried the highly magnetic and dense material northward of the straight trace of the Meers fault during oblique slip within the Frontal Wichita fault system. Along part of the reactivated Meers fault, the aeromagnetic data suggest highly magnetic dikelike bodies along the Wichita Mountain edge of the fault (Purucker, 1986). Ground magnetic profiles (Jones-Cecil, 1990) and two detailed gravity profiles across the Holocene fault scarp (Robbins and others, 1989) also suggest a dikelike body along the northwestern half of the Holocene rupture of the Meers fault. Near the junction of the Blue Creek Canyon fault and the Meers fault, the maximum horizontal gradients of the pseudogravity enhance both sides of one of the dikelike bodies.

The Oklahoma Geological Survey drilled five holes adjacent to the Meers fault, one of which was 200 ft (61 m) south of the Meers fault in the vicinity of the dikelike magnetic anomaly (see fig. 2). They encountered sheared and altered gabbroic and dioritic rock below 200 ft (61 m) depth (Kenneth Luza, oral commun., 1990; Collins, 1992). The median magnetic susceptibility (0.073 in SI units, Bradley and Jones-Cecil, 1991) of this highly magnetic rock is within the range of the magnetic susceptibilities of the Mount Sheridan Gabbro, the most magnetic rocks listed in table 1. No similar, highly magnetic rocks were encountered in drill holes 200 ft (61 m) south or 400 ft (122 m) north of this well. The sheared and altered gabbroic and dioritic rock is probably the source for the dikelike magnetic anomaly.

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CONVERSION FACTORS

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inches (in)	2.54	centimeters (cm)
feet (ft)	0.3048	meters (m)
miles (mi)	1.609	kilometers (km)

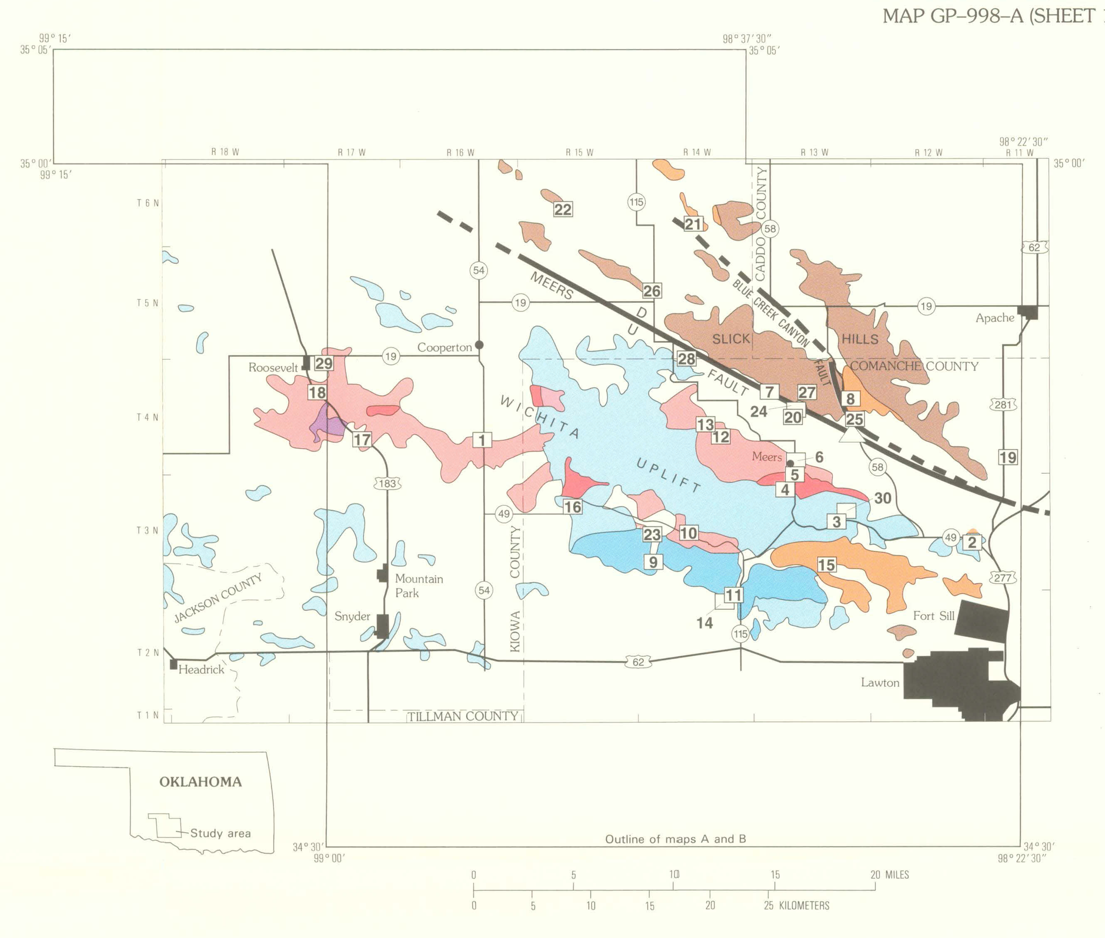
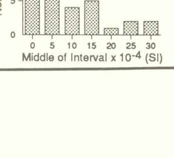
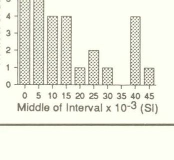
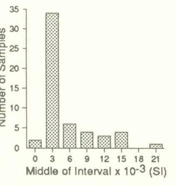
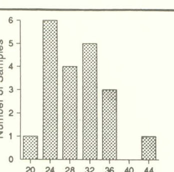
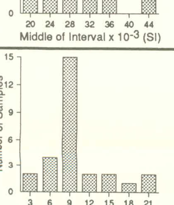
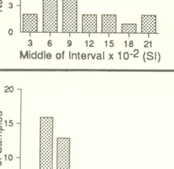
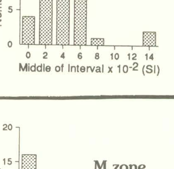


Figure 2. Localities for magnetic susceptibility measurements (after Bradley and Jones-Cecil, 1991). Geology, showing minor differences from figure 1, was adapted from figure 1 of Powell and others (1980) and shows surface exposures of igneous rocks in the area. Numbers show locations where susceptibility was measured with portable Geoinstruments susceptibility meter JH-8 having an accuracy of 2×10^{-5} SI. See table 1 for summary of the measured values.

Table 1. *Summary of magnetic susceptibility measurements*

[Summary for susceptibility localities shown in figure 2. NA, not applicable; "unassigned" means that rock sample is not assigned to a particular zone of unit. See Bradley and Jones-Cecil (1991) for detail regarding localities, contributors, and methods of measurement]

Age	Group	Formation	Member	Map locality (number of measurements in parentheses)	Mean magnetic susceptibility and standard deviation (SI units)	Median (SI units)	Histograms (X axis is magnetic susceptibility)
Permian	Hennessey[1]	NA	NA	7(14),19(5),20(1), 24(2)	< =0.00003	NA	
Ordovician and Cambrian	Simpson (Middle Ordovician), Arbuckle (Early Ordovician and Late Cambrian), Timbered Hills (Late Cambrian)	NA	NA	8(3),22(2),25(12), 26(2),27(5)	< =0.000075	NA	
[2]525±25 m.y.	Wichita Granite	Quanah Granite	NA	9(15) 11(14) 14(14) ------------------ 9+11+14(43)	0.00023±0.00025 0.0011±0.00065 0.0013±0.0012 ---------------- 0.00085±0.00089	0.0001 0.0011 0.00055 -------- 0.00054	
		Saddle Mountain Granite	NA	28(3)	0.0079±0.008	0.002	
		Mount Scott Granite	NA	2(10) 3(12) 23(5) 30(5) ------------------ 2+3+23+30(32)	0.024±0.014 0.0043±0.004 0.006±0.0042 0.026±0.016 ---------------- 0.014±0.014	0.018 0.0032 0.006 0.027 -------- 0.0084	
[2]525±25 m.y.	Carlton Rhyolite	Exposures at Bally Mountain	NA	21(6)	0.012±0.0031	0.013	
		Blue Creek Canyon	NA	8(27)	0.0028±0.00088	0.0026	
		Fort Sill	NA	15(21) ------------------ 8+15+21(54)	0.0068±0.0053 ---------------- 0.0053±0.0045	0.0045 -------- 0.0034	
[3]552±7 m.y.	Raggedy Mountain	Roosevelt Gabbros	Sandy Creek Gabbro Member	16(20)	0.029±0.0057	0.027	
			Mount Sheridan Gabbro Member	4(15) 5(13) ------------------ 4+5(28)	0.076±0.021 0.12±0.051 ---------------- 0.099±0.044	0.08 0.1 -------- 0.087	
[4]577±165 m.y.		Glen Mountains Layered Complex	Unassigned	6(4) 10(17) 12(11) 13(12) ------------------ 6+10+12+13(44)	0.099±0.048 0.022±0.015 0.035±0.017 0.046±0.012 ---------------- 0.039±0.028	0.11 0.023 0.028 0.046 -------- 0.03	
			N zone M zone L zone	1(20) 18(18) 29(4) 17(19) ---------------- 1+17+18+29(61)	0.031±0.035 0.0026±0.0017 0.02±0.025 0.0027±0.0039 ---------------- 0.013±0.025	0.018 0.0021 0.0081 0.0015 -------- 0.003	

[1]And its equivalent, Post Oak Conglomerate.
[2]Approximate age from Ham and others (1964).
[3]Bowring and Hoppe (1982).
[4]Lambert and others (1988).

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TOTAL-FIELD AEROMAGNETIC AND DERIVATIVE MAPS OF THE LAWTON AREA, SOUTHWESTERN OKLAHOMA

By
Meridee Jones-Cecil
1995

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