Tectonic history of the metamorphic basement rocks of the Sierra del Carmen, Coahuila, Mexico

Danielle L. Carpenter*  Department of Geological Sciences and Institute for Geophysics, University of Texas at Austin, Austin, Texas 78712

ABSTRACT

The Ouachita interior zone, arching through central Texas into the Big Bend region of the Rio Grande, has been defined chiefly on the basis of well samples and its associated gravity high. The only known exposure of rocks of the interior zone lies at the base of the Sierra del Carmen normal fault escarpment under approximately 1100 m of Lower Cretaceous carbonates in northern Coahuila, Mexico. These lower greenschist facies metamorphic rocks are thinly interlayered graphitic muscovite schist, graphitic marble, and minor quartzite with abundant quartz veins.

The dominant structures shown by the metamorphic rocks are isoclinal folds (F2) that have a well-defined axial planar foliation (S2). Earlier (F1) folds refolded by F2 are locally preserved. At the microscopic scale, S2 is a crenulation cleavage of the S1 foliation that has been enhanced by pressure solution. Feldspar porphyroblasts overgrow a folded graphitic S1 foliation. Throughout the area a second crenulation cleavage (S3) is present on the S2 foliation surface as an intersection lineation (L3). Subsequent Laramide regional to outcrop-scale folds and associated thrusts overprint these ductile structures. Lower Cretaceous carbonates form a large, east-verging overturned to recumbent anticline-syncline fold pair. The maximum offset along the normal fault is 1600 m.

Rb/Sr data from the muscovite schist gives a six-point, mineral–whole-rock isochron age of 277 ± 10 Ma, indicating that polyphase deformation and metamorphism occurred during the late Paleozoic Ouachita orogeny. An initial Sr ratio of 0.721 suggests an ancient source for these metasediments. Common Pb ratios of the metamorphic whole rock and galena from the Puerto Rico Mine are isotopically distinct from those of the North American craton, suggesting that they were derived from a non-North American source and thus may be equivalent to unmetamorphosed Ouachita flysch sediments. This indicates that the metamorphic rocks of the Ouachita interior zone were derived from an ancient allochthonous source that was deformed and metamorphosed during the late Paleozoic Ouachita orogeny.

INTRODUCTION

The late Paleozoic Ouachita deformational front trends south in the subsurface from Oklahoma through central Texas, and into the Big Bend region (Fig. 1A). Fold-and-thrust belts associated with Ouachita deformation involve unmetamorphosed Paleozoic sedimentary rocks and are exposed in the Marathon uplift and the Solitario dome in Trans-Pecos Texas (King, 1937; Tauvers and Muehlberger, 1989). These pre-Permian Paleozoic rocks are allochthonous (King, 1980) and have undergone a maximum of 84% shortening (Coley, 1987; Tauvers and Muehlberger, 1989). The trailing edge of exposed rocks has been transported approximately 200 km to the northwest (Muehlberger et al., 1984). Low-grade metamorphic rocks associated with Ouachita deformation have been encountered in the subsurface in central and west Texas, south of the zone of thin-skinned deformation, and are thought to represent the Ouachita interior zone (Flawn et al., 1961). This zone has been chiefly studied using gravity and well data. The interior zone is associated with a positive gravity anomaly that can be traced from the Texas-Oklahoma border into the Big Bend region of Texas. The primary cause of the gravity anomaly is the thinning of Laurentian continental crust and its transition into possible oceanic crust (Keller et al., 1989). The trend into Mexico of the Ouachita interior zone has been defined on the basis of the gravity high (Hendry et al., 1987) and by basement common Pb isotope signatures in west Texas and northern Mexico (James and Henry, 1993a, 1993b). Common Pb isotopic data of James and Henry (1993b) indicate that the Ouachita interior zone has a different Pb isotopic signature than does that of the North American craton farther north in west Texas, suggesting that the interior zone is allochthonous and exotic with respect to North America.

The only outcrop of the proposed Ouachita interior zone is an inlier of basement rocks beneath approximately 1100 m of Lower Cretaceous limestones at the Sierra del Carmen, northern Coahuila, Mexico, on the southern side of the Rio Grande at Big Bend National Park (Fig. 1B). Baker (1935) first noted the metamorphic rocks at the Sierra del Carmen, and Flawn and Maxwell (1958) and Flawn et al. (1961) were the first to describe them in any detail. The only previously published semidetailed geologic map of this area is a 1:250 000 scale geologic map of northern Coahuila (Smith, 1970). Prior to this study, no geologic map of the exposed metamorphic rocks existed, and the relationship with the overlying sedimentary rocks was unknown.

Initial geochronological work on rocks of the Sierra del Carmen exposure of the Ouachita interior zone yielded Early Permian metamorphic ages. These ages were determined by means of

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*e-mail: dcarpent@utig.ig.utexas.edu
Rb/Sr muscovite and whole-rock analyses (two-point isochrons) as well as a K/Ar model age of muscovite (Denison et al., 1969). These ages represent the youngest metamorphic event to have affected these rocks. The protolith age of the Sierra del Carmen metasedimentary rocks is undetermined.

This study documents the deformation and metamorphism affecting the rocks at the Sierra del Carmen and provides evidence supporting the hypothesis of Viele and Thomas (1989) that the Ouachita interior zone is allochthonous to North America and was derived from an exotic source. This study is the first to produce a geologic and structural map of this area, to determine the structural and stratigraphic relationship between the metamorphic rocks and the overlying Mesozoic strata, to document the extent of the carbonate strata, and to recognize a distinct isotopic character to the Ouachita interior zone.

GEOLOGIC SETTING

Late Cretaceous–early Tertiary Laramide and late Cenozoic Basin and Range deformational events affected rocks in the Big Bend region. Laramide deformation is characterized by northwest-trending reverse faults, folds, and monoclines with a general N70°E shortening direction (Muehlberger, 1989).

Extensional deformation associated with the development of the Rio Grande rift is localized within the Texas lineament zone (Dickerson and Muehlberger, 1994) and encompasses both the area of Big Bend National Park and the Sierra del Carmen exposure of metamorphic rocks (Fig. 1B). Basin and Range extension is characterized by the formation of deep grabens with steep, north-northwest–trending, bounding faults, which in some places show right-lateral movement and generally reactivate earlier Laramide contractional structures in an extensional sense (Muehlberger, 1989; Dickerson and Muehlberger, 1994).

Metamorphic rocks of the interior zone, as well as a thick section of overlying Cretaceous conglomerates in the Big Bend region, are well exposed along a normal fault escarpment at the Sierra del Carmen (Fig. 2). The normal fault, which has a minimum throw of 1600 m down to the west, is the eastern bounding fault to the Big Bend graben system (Fig. 1B). The metamorphic rocks are exposed at the base of the escarpment within and between the four canyons north of the abandoned Puerto Rico lead mine. Flawn and Maxwell (1958) noted that no metamorphic rocks exist south of the mine.

At the base of the Lower Cretaceous carbonate section, a red pebble-cobble conglomerate non-conformably overlies the metamorphic rocks. The well-exposed contact is an erosional surface that locally undulates and is both parallel and oblique to the metamorphic foliation. The conglomerate contains clasts of black and white cherts, a variety of limestones, vein quartz, pieces of sandstone and minor amounts of schist and rare volcanic rocks (Carpenter, 1996). These fragments are equivalent to the Paleozoic Marathon facies of the Fort Peña, Maravillas, Caballos, and Dimple Formations (M. Helper, 1995, personal commun.). Further discussion of this conglomerate and its relationship to other basal Cretaceous conglomerates in the Big Bend region can be found in Carpenter (1996). Detailed discussions of the carbonate stratigraphy at the Sierra del Carmen can be found in Smith (1970) and Barcelo-Duarte (1983).

The metamorphic rocks are exposed in three main areas—northern, central and southern (see Fig. 2). Reconnaissance of canyons farther to the north indicates that some metamorphic rocks are present; they were not studied, however. The metamorphic rocks in the northern study area are well exposed for approximately 1.5 km. The central exposure is confined to El Socavón Canyon. This smallest exposure of metamorphic rocks, approximately 0.25 km in extent, is structurally and topographically higher than either of the other study areas. The southern exposure is confined to the canyon south of El Socavón and just north of Puerto Rico Mine. On the basis of aerial photograph interpretation of the conglomerate contact, the metamorphic rocks are interpreted to be exposed for approximately 0.75 km and occur south of the arroyo, but no exposures south of the arroyo were found. Outcrops of metamorphic rocks are scattered north of the arroyo for approximately 0.5 km.

LARAMIDE AND BASIN AND RANGE STRUCTURES

The Lower Cretaceous carbonates along the base of the Sierra del Carmen escarpment form a large, north-northwest–trending, east-verging overturned to recumbent anticline-syncline fold pair within the Cupido Formation with approximately 150 m of structural relief (Figs. 3 and 4). Units are relatively horizontal east of the escarpment. Laramide deformation in the Mexican Sierra del Carmen was previously unknown. The anticline is only expressed in the carbonates in the northern part of the map area. The morphology and orientation of the syncline changes slightly southward from canyon to canyon. The axial plane dip varies in steepness, and becomes recumbent southward at the highest structural levels. In the central area, the folds are generally box-like, but become overturned with higher
structural relief to the south. Other shortening structures are exposed in a block between two normal faults exposed in the Cupido Formation on a large cliff face. Angular discordances between beds result from east-directed thrust faults. Folds of bedding are spatially related to the thrust faults with similar senses of motion.

Foliations within the metamorphic rocks define an antiform that is nearly coincident in trend to that of the fold in the overlying Lower Cretaceous rocks (Figs. 5 and 6), indicating that the metamorphic rocks were also folded by this regional structure. Stereonet plots of poles to metamorphic foliation indicate that the foliations are folded with a fold axis of N16°W, 8° (Fig. 5a). The difference in orientation (primarily in plunge) between the fold axes in the carbonates (N11°W, 10°) and in the metamorphic rocks results from the angular discordance between the metamorphic foliation and the overlying sedimentary bedding. The metamorphic foliation is somewhat oblique to bedding in the overlying Lower Cretaceous rocks (Fig. 5).

In one location, fault-propagation folds and associated thrusts are exposed in the metamorphic rocks along strike from similar folds and faults in the Cupido Formation. These thrusts within the metamorphic rocks show top to the east direction of motion. In general, rocks of relatively uniform dipping metamorphic foliation are present on the hanging wall of the thrusts. Folds of the metamorphic foliation are tight and kink-like at the fault contact and gradually grade into broader folds below the fault.

Kink, box, and chevron folds of the metamorphic foliation (Fig. 7) are also found adjacent to
the vertical conglomerate contact where the more competent conglomerate is in irregular contact with the well-foliated metamorphic rocks. Small-scale kinking of the metamorphic rocks near the vertical conglomerate contact, and small-scale faulting of the contact, apparently occurred as the entire package of rocks was folded into a syncline. Some of these folds are associated with thrusts and appear to be fault-propagation folds.

Similar kink, box, and chevron folds of the metamorphic foliation are found along the trace of the Basin and Range normal faults. In one location, west of the normal fault in El Socavón Canyon, the Santa Elena Formation is also kinked and folded on the hanging wall of the normal fault. These folds are localized within layers consisting predominantly of rudist fossils and are similar to contractional structures in the adjacent metamorphic rocks.

In all localities, the kinks form sharp bends of the S2 foliation and generally grade vertically into box folds. The kinks show no thickening of layers into the hinge region and can be characterized by flexural-slip folding. At the thin-section scale, the foliation planes, defined by alternating quartz-rich layers and micaceous layers, are physically broken in the kink hinge regions. No signs of internal flow were observed. Analysis of the orientation of the kink planes and axes for those related to thrusts indicate that no preferred kink orientation exists, although the axial planes and many axes generally trend northwest. Associated thrust faults show a distribution similar to that of the kink axial planes. Most of the kink axial planes not related to thrusts dip northward and generally have a northwest (from west to south) trend. Those with a northeast trend are primarily those near the vertical conglomerate contact.

One or two north-northwest–trending normal faults cut the carbonates and metamorphic rocks and truncate the anticline-syncline fold pair (Figs. 2 and 3). The maximum offset along the normal faults is 1600 m where the western normal fault juxtaposes the Lower Cretaceous Santa Elena Formation and the metamorphic rocks. In the northern study area, the eastern normal fault is steeper than the western normal fault, whereas in the southern area, the eastern fault is shallow (Figs. 3 and 5).

Two monoclines within the Santa Elena Formation are present in the hanging wall of the major normal fault. In the northern area, a broadly defined monocline has a northwest-trending axis, oblique to the overturned fold axis but subparallel to the strike of the normal fault at that location (Fig. 3a). A broad monocline is also present in the hanging wall of the normal fault in El Socavón Canyon. The position and orientation of these monoclines in the hanging wall of the large normal fault and their differences with the

Figure 3. Cross section across the Sierra del Carmen escarpment. Laramide contractional structures are in a fault-bounded block exposed on a vertical cliff face in northern area. See Figure 2 for section line and abbreviations. (a) Stereonet of poles to bedding in the Santa Elena Formation around monoclines. Dots correspond to measurements from the northern monocline; squares correspond to measurements from the El Socavón monocline. The great circle represents the orientation of the western normal fault. (b) All measurements from Laramide fold in Lower Cretaceous rocks. The great circle represents the axial plane of the fold determined by three-point method and constrained by the fold axis.

Figure 4. Anticline-syncline fold pair in the Cupido Formation in the northern study area. Anticline is not exposed in the carbonates to the south of this locality (see Fig. 2). View is to the north; the cliff is approximately 100 m in elevation.
Laramide contractional folds suggest that they may be drag folds related to the Basin and Range normal fault.

**Interpretation**

Well-studied contractional deformation of Lower Cretaceous strata in Texas, similar to that observed in the Sierra del Carmen area, has been interpreted to be Laramide (Cobb and Poth, 1980; Moustafa, 1988; Muehlberger, 1989; Maler, 1990). The anticline-syncline fold pair described above is interpreted to be a Laramide fault-propagation fold pinned within the metamorphic basement rocks. This interpretation is supported by fault-propagation fold-and-thrust sets exposed both within the metamorphic rocks and within the Cupido Formation. Shortening was approximately N79°E with movement toward the east, slightly oblique to the maximum compressive stress direction determined for the Sierra del Carmen in Texas for Laramide deformation of N65°E (Moustafa, 1988).

The westernmost horizontal limb of the fold pair has subsequently been cut by two Basin and Range down-to-the-west normal faults, one of which is the Sierra del Carmen escarpment-defining normal fault, which dips approximately 45° to 55°W. Laramide structures are preserved in the zone between these faults. In both the hanging wall and foot wall of the Basin and Range normal fault, contractional structures (e.g., kinks, chevron to box folds, overturned folds, fault-propagation folds, and thrust faults) are preserved locally adjacent to the fault. The spatial relationship between these contractional features and the large Basin and Range normal fault, as well as the location and dip of the fault relative to the Laramide anticline, suggest that this fault reactivated the thrust responsible for the contractional deformation. The western normal fault most likely joins the reverse fault at depth. This hypothesis of Basin and Range reactivation of Laramide reverse faults also has been proposed for adjacent areas in Texas (Cobb and Poth, 1980; Moustafa, 1988; Muehlberger, 1989; Tauvers and Muehlberger, 1989). Other contractional deformation spatially related to the folded conglomerate contact is interpreted to be related to movement along this irregular contact during folding.

The monoclines observed within the Santa Elena Formation are immediately adjacent to the normal fault in the hanging wall and have the same trend as the normal fault, suggesting that they formed in response to extension. Analysis of the aerial photographs indicates that the Santa Elena Formation is relatively horizontal west of the monocline, indicating that the normal fault is not listric at depth. The normal fault is shown with a vertical bend in the cross section in Figure 3 because it is required to restore the cross section to post-Laramide conditions and remove the monocline. Restoration of a cross section with a planar normal fault would not remove the monocline (M. Maler, 1996, personal commun.). Thus, it is most likely that upon reactivation of the Laramide thrust fault as a Basin and Range normal fault, an extensional monocline was created on the hanging wall of the normal fault over a vertical bend in the fault plane.

The monoclines in the Texas Sierra del Carmen, however, have been interpreted to be Laramide contractional monoclines (Moustafa, 1988; Maler, 1990). The nearest monocline in Texas, the Big Bend monocline, is similar in morphology to the Sierra del Carmen monoclines, but it is not along strike and has a different trend. In this study, structures within the carbonates were only mapped in enough detail to determine their relationship to structures within the metamorphic rocks. Future work is necessary to substantiate the extensional interpretation for the Mexican monoclines, to determine the relationship between the Mexican and Texas monoclines, and to determine if any of the previously interpreted contractional monoclines could in fact be extensional.
In summary, the Lower Cretaceous conglomerate and carbonates were deposited nonconformably upon the metamorphic rocks. During the Late Cretaceous–early Tertiary Laramide orogeny, a fault-propagation fold developed deforming the entire metamorphic rock–Lower Cretaceous package. It is unknown whether this fold propagates to higher structural levels to deform the Santa Elena Formation. The Laramide thrust responsible for the folding was subsequently reactivated during late Cenozoic Basin and Range extension preserving and/or forming monoclines in the hanging wall.

PRE-CRETACEOUS DEFORMATION AND METAMORPHISM

The metamorphic rocks consist of graphitic marble, graphitic muscovite schist, and quartzite containing abundant quartz veins. These units vary along the length of the exposure. In the southern study area, the dominant rock type is a dark gray to black pelitic schist containing minor quartzite lenses. North of El Socavón Canyon, 2–3-cm-thick marble layers containing thin schistose interlayers dominate (Fig. 7). No marbles were found in the southern area, and no quartzites were mapped farther north. The metamorphic rocks generally dip to the north, indicating that the pelitic schists lie structurally below the marbles with respect to the S2 foliation. Primary stratigraphic indicators are absent.

Thin-bedded, light gray, fine-grained marbles are commonly stained with iron oxide near the conglomerate contact. Minerals include calcite, quartz, muscovite, and pyrite. Muscovite, chlorite, graphite, and quartz are generally found in the schist interlayers. The calcite has been totally recrystallized and no fossils were found. Locally, the marble is more of a calcareous schist than a schistose marble.

Dark gray to black, fine-grained pelitic schists contain sericitic muscovite, chlorite, quartz, albite feldspar, graphite, and rare biotite. Accessory minerals include tourmaline, rutile, ilmenite, and apatite. Graphite is pervasive throughout the rocks, indicating a carbon-rich depositional environment in which significant organic material was preserved. Quartzite lenses are rare and may represent small channel fills or parts of a turbidite sequence. The quartzites do not constitute a major rock type and are not mappable. In addition to quartz, the quartzites contain muscovite, chlorite, tourmaline, and rare biotite.

Feldspar porphyroblasts contain graphitic inclusions defining a foliation that is oblique to the dominant metamorphic foliation. Folds of the graphitic foliation are preserved locally in the feldspar porphyroblasts. Rims of inclusion-free feldspar surround inclusion-rich cores. Tourmaline is zoned with blue-green cores and light brown rims and is localized in schistose areas along with abundant chlorite and quartz. Rutile is replaced by ilmenite and forms long thin blades. The dominant S2 foliation commonly wraps the ilmenite blades, suggesting that rutile growth is pre-S2 foliation development. Individual chlorite and muscovite grains locally

Figure 6. Detailed geologic map of the northern study area illustrating the Laramide anticline expressed in metamorphic rocks. The topographic contour interval is 20 m. See Figure 2 for location of map area and key to symbols.
control the grain boundaries of quartz grains. In addition, some muscovites are large, oblique to the S2 foliation, and undeformed, suggesting that they were recrystallized during and after development of the S2 foliation. This evidence of some degree of static recrystallization is supported by 120° grain boundaries in some quartz. However, quartz grains also show the development of tilt walls and subgrains, suggesting that static recrystallization was minor.

Structure

The metamorphic rocks are consistently very well foliated (Fig. 7), the foliation developing axial planar to isoclinal folds of bedding. The centimeter- to meter-scale isoclinal folds show thickening in the hinge regions and thinning on the limbs (Fig. 8), characteristic of ductile deformation. Fold hinges in the marbles show heavily twinned calcite and evidence of subgrain development. Quartz deformation fabrics are dominated by evidence of low- to moderate-temperature deformation and incipient dynamic recrystallization including undulose extinction, bulge nucleation, grain boundary migration, and subgrain and tilt wall development. In quartz-rich layers and veins, quartz grains are elongate parallel to the dominant foliation and axial planar to isoclinal folds.

Locally, isoclinal folds (F2) are observed to have ductilely refolded earlier folds (F1). In the marble and schist interlayers in the northern area, the dominant foliation is axial planar to the second generation of folds (F2), and is therefore a S2 foliation. This refolding is not observed on an outcrop scale in the southern area, but evidence of an S1 foliation and F2 fold hinges is observed petrographically.

Microstructural analysis indicates that the dominant foliation (S2) is a crenulation cleavage of an earlier foliation (S1) and has been enhanced by pressure solution (Fig. 9), indicating low-temperature deformation and the presence of fluids. Where the foliation is not a crenulation cleavage, it is defined by alternating muscovite-rich and quartz-rich layers. S1, where preserved in the crenulation hinges, is defined by fine-grained muscovite and graphite. F1 fold hinges refolded by the F2 isoclinal folds are also observed. Muscovites form small recrystallized grains in both F1 and F2 fold hinges, indicating that temperatures high enough for muscovite recrystallization occurred during both folding events.

Additional evidence of the S1 foliation is found where feldspar porphyroblasts overgrow an early graphitic foliation. Feldspar porphyroblasts contain graphite inclusions defining a foliation that is commonly oblique to the dominant foliation (Fig. 10). In places, layers of feldspar containing and parallel to S1 have been folded by F2 (Fig. 11). It is rare that feldspar overgrows entire folds of this preserved foliation. Therefore, it is interpreted that feldspars overgrow the S1 foliation and are post S1 and locally post S2. Overgrowths of inclusion-free feldspar occur around graphite-inclusion–rich feldspar cores (Fig. 10). Usually these inclusion-free zones are preferentially located parallel to the dominant foliation (S2) on opposite sides of the grain, but some completely rim the inclusion-rich core. The feldspar with inclusion-free rims results from feldspar porphyroblasts overgrowing a graphitic foliation post-S1 and pre-S2 with continued rim growth syn-S2. As the foliation was shortened perpendicular to the limbs of the crenulation cleavage, feldspar crystallized as overgrowths in the low-stress regions behind the rigid feldspars. Feldspars do not exhibit evidence of dynamic recrystallization or any other plastic strain.

Strain shadows exist around the feldspar porphyroblasts and are composed of large grains of muscovite and chlorite. Both the strain shadows and the micaceous minerals are parallel to the dominant foliation. Strain shadows of large chlorite grains are also present around ilmenite blades. The long axes of the ilmenite blades are oblique to the dominant S2 foliation. Where chlorine is not located in strain shadows, it is usually oriented parallel to the dominant foliation. Therefore, chlorite most likely recrystallized during the development of the S2 foliation. Pressure solution seams commonly wrap feldspar strain shadows where F2 folds are present in the feldspar. Therefore, the pressure solution continued after F2 folding. Pyrites that have fibrous quartz overgrowths are also present in schists; the quartz is present on the ends of the long axes of rectangular pyrite crystals. The fibrous quartz is oblique to the dominant foliation.

A crenulation cleavage forms lineations on the S2 foliation surfaces throughout the area. In thin section, this lineation is seen as a slight crenulation cleavage of S2. It is oblique to the S2 foliation (Figs. 9 and 11) and is therefore a L3 lineation. Muscovites in the hinges of this later crenulation are not recrystallized. Therefore, this last phase of ductile deformation was at a lower temperature than the previous deformation phases. L3 is folded by Laramide contractional deformation affecting the metamorphic rocks discussed earlier. The dominant S2 foliation is at an angle to, and truncated by, the sub–Lower Cretaceous unconformity.

Milky white quartz veins are present throughout the study area and formed during every phase of deformation. They range from undeformed to folded and refolded, and commonly form knobs of
quartz indicative of floating fold hinges or boudins. Quartz veins also exhibit similar deformation fabrics described earlier for the schist. Calcite veins are also present but tend to fill spaces in the hinges of kinks and are probably associated with Laramide contractional deformation. These calcite veins contain coarse-grained calcite and do not appear to be recrystallized.

**Orientation of Metamorphic Foliation Prior to Pre-Cretaceous Unconformity**

As discussed previously, the metamorphic foliation has been deformed by the same Laramide fold expressed within the Lower Cretaceous carbonates. Removing the effects of the Laramide fold on the metamorphic rocks, where best expressed in the northern area, yields an average S2 restored orientation of N45°E, 40°W (Fig. 5c). Note that the restored data (poles to foliation) scatter throughout the entire southeast quadrant; the mean of the data is taken to be the average restored orientation. The metamorphic foliation in the southern area, or the “horizontal” limb of the fold, also plots in the southeast.
quadrant. The trend of the pre-Cretaceous foliation is northeast to southwest, indicating a shortening direction from the southeast or northwest. This direction correlates well with the widely accepted idea that the direction of collision during the Ouachita orogeny was along this azimuth. However, dominant vergence is to the northwest, and the restored metamorphic foliation dips in that direction. Because this is the only recognized outcrop of the interior zone rocks, it is unknown whether the exposure is on the limb of a larger structure or if it has been rotated from its original orientation.

Summary of Pre-Cretaceous Deformation and Metamorphism

Figure 12 summarizes the development of the pre-Cretaceous structures within the metamorphic rocks. Folding and transposition of bedding generated a S1 foliation that was subsequently folded into tight isoclinal folds during the F2 folding event. A crenulation cleavage of the S1 foliation, enhanced by pressure solution, generated the S2 foliation axial planar to the isoclinal folds. The S1 foliation was overgrown by feldspar porphyroblasts prior to or during development of the S2 foliation. A crenulation developed oblique to the main S2 foliation (S3). All of these structures were folded by the Laramide contraction. Petrographic analysis indicates that the metamorphic rocks are dominated by low-temperature deformation and fabrics. Temperatures were high enough during F1-S1 to grow muscovite and biotite and during F2-S2 to grow albite and to recrystallize quartz and muscovite, but not high enough to strain feldspar. Large grains of muscovite oblique to the S2 foliation suggest that metamorphism continued after the F2 folding event.

ISOTOPIC DATA

Several methods were employed to determine the age of deformation and metamorphism of the lower greenschist facies metamorphic rocks at the Sierra del Carmen. Initial attempts concentrated on determining the age of maximum metamorphism. Metamorphic rutiles were separated and analyzed for U/Pb, but no uranium was found in the rutiles. Subsequent analyses concentrated upon the Rb/Sr system and isotopic analysis of common Pb.

Sampling Locations

Figure 2 shows the sampling location for #SDC-GC1, a pelitic muscovite schist from a prospect pit in the southern study area. The sampling location for #SDC-GC1 is most likely the same as the location for both Flawn’s and Denison’s samples; however, the exact location for Denison’s samples is unknown (R. Denison, 1995, personal commun.). Galena from the Puerto Rico Mine was also analyzed for common Pb.

Rb/Sr Analyses

Denison et al. (1969), in their attempt to find the source of the granitic gneiss boulders in the Pennsylvaniaan Haymond Formation of the Marathon uplift, analyzed the metamorphic rocks from the Sierra del Carmen using Rb/Sr and K/Ar methods. The Rb/Sr data are summarized in Table 1. Metamorphic ages from the Rb/Sr muscovite-whole rock two-point isochron and from K/Ar analyses of muscovite yield Permian metamorphic ages of 286 ± 21 Ma (Rb/Sr) and 270 ± 5 Ma (Denison et al., 1969). These ages have been adjusted from the published ages using currently accepted decay constants (Rb/Sr: Steiger and Jäger, 1977; K/Ar: ...

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<td>0.7323</td>
<td>3.11</td>
<td>285 ± 21†</td>
</tr>
<tr>
<td>61 Mica</td>
<td>-</td>
<td>288.1</td>
<td>125.5</td>
<td>0.7466</td>
<td>6.62</td>
<td></td>
</tr>
</tbody>
</table>

*Data reported to 2σ; isotopic errors are reported in percent; mean standard weighted deviation (MSWD) = 461.
†From Denison et al. (1969); age recalculated with new decay constant (λ = 1.42 × 10⁻¹⁰/yr).
Dalrymple, 1979). Flawn et al. (1961) reported ages of 240 Ma and 370 Ma on muscovites, but stressed that these ages are inconclusive; these ages were achieved by Rb/Sr analysis (G. Edwords, Shell Development Company, U.S.A., as reported by Flawn et al., 1961). All of these data suggest that the metamorphic rocks have a late Paleozoic history. The Rb/Sr two-point isochron of Denison et al. (1969) represents the youngest metamorphic age, which falls near the end of the Ouachita orogeny age range of 280 to 340 Ma (Viele and Thomas, 1989).

**Analytical Technique.** Whole-rock and mineral samples were analyzed by the isotope dilution method using a $^{87}\text{Rb}^{84}\text{Sr}$ spike. After being spiked, samples were digested in a mixture of HF and HNO$_3$ in 15 ml screw-top Teflon containers. Samples were dried down and taken up in HCl, and were centrifuged to ensure that no solid material had escaped dissociation. Residual solids was returned to acid dissolution until all solids had decomposed. Rb and Sr were isolated from all other elements and from one another by using a standard cation exchange column (BIO-RAD AG 50W-X8 resin) eluted with 2.3N HCl.

Rb isotope compositions were determined using an automated, 30-cm-radius, single-collector mass spectrometer equipped with double tantalum filaments. Sr isotope compositions were determined using a Finnigan MAT-261 multicollector mass spectrometer operated in static acquisition mode. A more complete description of the analytical technique can be found in Awwiller (1994) and Long et al. (1997).

**Sierra del Carmen Data.** Five mineral fractions (apatite, feldspar, and several fractions of muscovite, all handpicked to near 100% purity) and one whole-rock fraction from sample #SDC-GC1 were analyzed by the Rb/Sr method by L. E. Long at the University of Texas at Austin (Table 1) (Carpenter and Long, 1995). Muscovite was separated into three size fractions. Smaller crystals correspond to muscovite aligned during the development of the S1 foliation; larger crystals are undeformed and record post-S2 foliation mineral growth and recrystallization. Thus, the three analyzed muscovite fractions are a bulk fraction, grains larger than 200 µm, and grains between 70 µm and 100 µm.

The data points from SDC-GC1 and from Denison et al. (1969) are plotted in Figure 13. The apatite fraction, which contains virtually no rubidium, preserves the strontium ratio of the rock at the time of metamorphic recrystallization. This value, 0.721, is quite a bit higher than the initial ratio of 0.706 assumed by Denison et al. (1969) in his model-age calculation.

The slope of the Rb/Sr isochron yields an age of 277 ± 10 Ma, which is interpreted as the youngest metamorphic event that affected these
rocks and corresponds to the latest pulse of metamorphism associated with the Ouachita orogeny. The data points define a well-constrained line, suggesting that these rocks reached isotopic equilibrium during the Ouachita metamorphic event.

Similar ⁸⁷Sr/⁸⁶Sr initial ratios are observed in metamorphic samples from elsewhere in the Ouachita interior zone (Long et al., 1995). The Exxon #1 Gatlin well in Terrell County, Texas, 100 km northeast of the Sierra del Carmen, was drilled 7620 m (25000 ft) into metamorphic basement rocks of the Ouachita interior zone (Marsaglia et al., 1994). Values of ⁸⁷Sr/⁸⁶Sr at the time of low-grade metamorphism range from 0.7093 to 0.7149, and average 0.7112 (L. Long, 1995, personal commun.). Cameron et al. (1992) performed isotopic analyses of crustal paragneiss xenoliths from the La Olivina cinder cone in Chihuahua, Mexico, and obtained similar high ⁸⁷Sr/⁸⁶Sr data ranging from 0.720 to 0.746, averaging 0.728. La Olivina samples are discussed further. Data from rocks of the Ouachita interior zone summarized here indicate that it may be possible to characterize the Ouachita interior zone by a high initial Sr ratio, suggesting that the protolith had a substantially older history (i.e., these metasedimentary rocks have been recycled from a much older continental crustal source).

Common Pb Isotopes

Background. Common Pb isotope data from west Texas and northern Mexico have been published by James and Henry (1993a, 1993b). Their data represent isotope compositions of Tertiary igneous intrusions and ore deposits, the ratios of which are inherited from the basement. James and Henry (1993b) delineated two basement compositions divided into three zones in west Texas and northern Mexico (see Fig. 14). The northwestern zone corresponds to the North American data; the central or intermediate zone corresponds to the Ouachita frontal zone, the fault overlap of the Ouachita facies over the North American basement; the southeastern zone corresponds to the Ouachita interior zone (James and Henry, 1993b). The Ouachita frontal zone is characterized by fold and thrust belts of Paleozoic rocks deformed during the Ouachita orogeny, expressed in the west Texas Marathon uplift and the Solitario dome. These deformed upper Paleozoic terrigenous flysch units had a southern depositional source and are interpreted as allochthonous to North America (Nicholas and Waddell, 1989). The Ouachita interior zone consists of low-grade metamorphosed rocks that have been involved in polyphase deformation and metamorphism during the Ouachita orogeny. James and Henry’s data suggest that the Ouachita interior zone is isotopically distinct from cratonal North America.

Sierra del Carmen Data. Two samples from the Sierra del Carmen were analyzed, the metamorphic whole rock and galena from the Puerto Rico lead mine, for their common Pb isotopes by E. W. James at the University of Texas at Austin. The data from this study are presented in Table 2 and plotted in Figure 14. See James and Henry (1993b) for a discussion of analytical techniques. Cumming et al. (1979) analyzed galena from the Puerto Rico Mine and reported Pb isotope data similar to those from this study. The study between the metamorphic rock and the galena Pb signatures supports the hypothesis that the intrusion retains much of its Pb isotopic signature from the rocks through which it passed.

Relationship Between Sierra del Carmen and Regional Pb Data. These new common Pb data from the Sierra del Carmen are compared with the regional common Pb picture (James and Henry, 1993b) in Figure 14. It is evident that the metamorphic basement sample from the Sierra del Carmen is distinct from all the data for the northwest and central zones in west Texas and northern Mexico. The data plot at the edge of the southeast zone, probably because the data from the Tertiary intrusives (James and Henry, 1993b) contain Pb from the mantle as well as basement crustal sources, whereas the Sierra del Carmen data are from the basement. This conclusion is supported by Pb isotopes of paragneiss xenoliths (basement?) from the La Olivina cinder cone in Chihuahua, Mexico (Cameron et al., 1992). These data plot near the data from this study (the Sierra del Carmen rocks). In addition, both the La Olivina xenoliths and the Sierra del Carmen samples are isotopically similar to those from the southeastern zone of James and Henry (1993b).

Cameron et al. (1992) also analyzed samples from the Paleozoic Haymond, Tesnus, and Dimple Formations (Ouachita flysch) from the Marathon uplift. These three units represent approximately 3,000 m of flysch deposits (McBride, 1989). The Tesnus Formation is composed of thick-bedded sandstone and shales of Late Devonian to Early Pennsylvanian age with a southerly source. The Dimple Formation is a calc-arenite turbidite of early Pennsylvanian age that had a northerly source, and is the middle flysch unit (McBride, 1989). The Haymond Formation, thin-bedded sandstones and shales and a conglomerate unit, is early Middle Pennsylvanian in age, and had a southerly source. The Dimple Formation lead data are different than those for the other two flysch units with a southerly source (Cameron et al., 1992). Figure 14 shows that the Sierra del Carmen, La Olivina paragneiss xenoliths, and the Haymond and Tesnus Formations plots are in the same range, whereas the Dimple Formation is quite different. The Dimple Formation flysch is even more unlike any of the lead data for cratonal North America than the other flysch. Cameron et al. (1992) suggested that the paragneiss xenoliths are the metamorphosed equivalents of Paleozoic Ouachita flysch units. This hypothesis was suggested by Nicholas and Waddell (1989) and Viele and Thomas (1989). However, no other geologic correlation exists to confirm that the metamorphic rocks of the Ouachita interior zone are the metamorphosed equivalent of the Paleozoic Ouachita flysch units.

Implications. The common Pb isotope data from west Texas and northern Mexico indicate that the Ouachita interior zone is isotopically distinct from cratonal North America in this region (James and Henry, 1993b), and this observation is supported by the new data from the Sierra del Carmen. The probable source for the metamorphic rocks of the Mexican Sierra del Carmen, and the Ouachita interior zone as a whole, is not North America. Presumably, the source of the sediments was the oncoming continent involved in the continent-continent collision during the Ouachita orogeny. The approaching continent was either northern South America or a microplate (such as the Coahuila terrane or the Yucatan block), caught in the collision between southern North America and northern Gondwana that resulted in the Ouachita orogeny.

CONCLUSIONS

This study has confirmed the presence of the Ouachita interior zone in northern Mexico and has characterized the late Paleozoic deformation and metamorphism associated with this orogeny. Pre-Cretaceous tectonism intensely deformed the lower greenschist facies rocks of the Sierra del Carmen. The dominant well-developed foliation is axial planar to isoclinal folds of bedding; locally an earlier fold generation is observed. This foliation was subsequently refolded during Laramide deformation, which penetrates the basement and involves Cretaceous sedimentary rocks, as in the Texas Sierra del Carmen. Structural mapping has identified a regional north-trending, east-vergent Laramide fold, extending the extent of known Laramide deformation. A Basin and Range normal fault reactivated a buried reverse fault associated with this fold.

The youngest metamorphic event to affect these rocks is dated as 277 ± 10 Ma, part of the late Paleozoic Ouachita orogeny. These rocks could, however, have an older history indicated by the high initial Sr ratio. This value, 0.721, suggests that the protolith had a substantially older history (i.e., that these metasedimentary rocks have been recycled from a much older continental crustal source). Pb isotopes show that the basement of the Sierra del Carmen, and of the
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