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PRECAMBRIAN STRUCTURES IN CANON DEL TRIGO MANZANO MOUNTAINS, CENTRAL NEW MEXICO

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INTRODUCTION

The Manzano Mountains form a block-faulted uplift along the eastern edge of the Rio Grande rift. They extend northward from Abo Pass to Escabosa, a distance of some 55 km. The range consists of a large mass of Precambrian metamorphic and igneous rocks, overlain by a thin veneer of east-dipping Paleozoic sedimentary rocks. Recent sands and gravels bury the basement rocks to the west. Details of the regional geology are summarized in Reiche (1949).

According to most previous workers, Precambrian geology of the Manzano Mountains is fairly simple (Reiche, 1949; Stark, 1956; Myers and McKay, 1971; Condie and Budding, 1979). These workers have recognized two periods of deformation in the basement rocks. The first deformation affected units in the northern part of the area, where rocks seem to be multiply deformed. This region, north of Comanche Canyon, is separated from rocks to the south by an apparent angular unconformity (Reiche, 1949; but see Blount, road-log segment I-C of this guidebook, for a different explanation). According to published interpretations, rocks south of the unconformity are younger and have been deformed only once.

Estimates of stratigraphic thickness in metamorphic rocks south of the unconformity are in excess of 5 km, whereas the overall thickness of Precambrian units in the Manzanos is estimated as 12.5 km (Condie and Budding, 1979).

The present report summarizes results of detailed mapping of a 3.5 km' area in Callon del Trigo. The mapped area lies entirely within the "Blue Springs schist" of Stark (1956), and it lies approximately 5 km south of the apparent angular unconformity. According to previous structural interpretations, rocks in Callon del Trigo have been deformed only once. However, Precambrian exposures show numerous features which are inconsistent with this simple structural history. These features include multiple schistosities, folded foliations, and refolded folds. Such features are characteristic of multiply-deformed terranes. If basement rocks in the area are multiply folded, then their stratigraphic thickness may be considerably less than published estimates.

LITHOLOGIES

Rocks in Callon del Trigo have been divided into three mappable units (fig. 1). Green to brown phyllite, typically containing discontinuous stringers of vein quartz, is most abundant. Mineral assemblages include chlorite-calcite-albite and chlorite-chloritoid-biotite, both with muscovite and quartz, indicating upper greenschist facies metamorphism.

The second map unit is a fine-grained, well-laminated micaceous quartzite. Laminae range from 1 mm to 10 cm thick and alternate from buff to gray to green, giving the rock a striped appearance. Specimens contain 80-90 modal percent quartz, but individual crystals are smaller than 10 microns and are barely visible in a petrographic microscope. Due to its fine grain size and quartzose composition, this rock is interpreted as metamorphosed impure chert.

The third map unit consists of lenticular, discontinuous layers of orthoquartzite. Most layers are less than 700 m in length but the largest layer, which crosses Cation del Trigo in the eastern part of Figure 1, extends for at least several kilometers. Quartzites are almost pure quartz, with individual crystals averaging 0.1-1 mm in size. They contain FeTi oxides, trace amounts of zircon, and rare tourmaline and apatite as the only impurities. Structures resembling primary bedding are rare. Stark (1956) called these layers "quartz reefs" and interpreted them as pods of intrusive vein quartz which were deposited from solution. Keyes (1920) and Condie and Budding (1979) ascribed a sedimentary origin to the layers, interpreting them as the metamorphosed equivalents of lenticular sand bodies. The rare occurrence of crossbedding (fig. 2) supports a sedimentary origin for the "quartz reefs." The poor preservation of primary sedimentary features is probably due to solution and redeposition of FeTi oxides, combined with recrystallization and ductile flow during deformation.

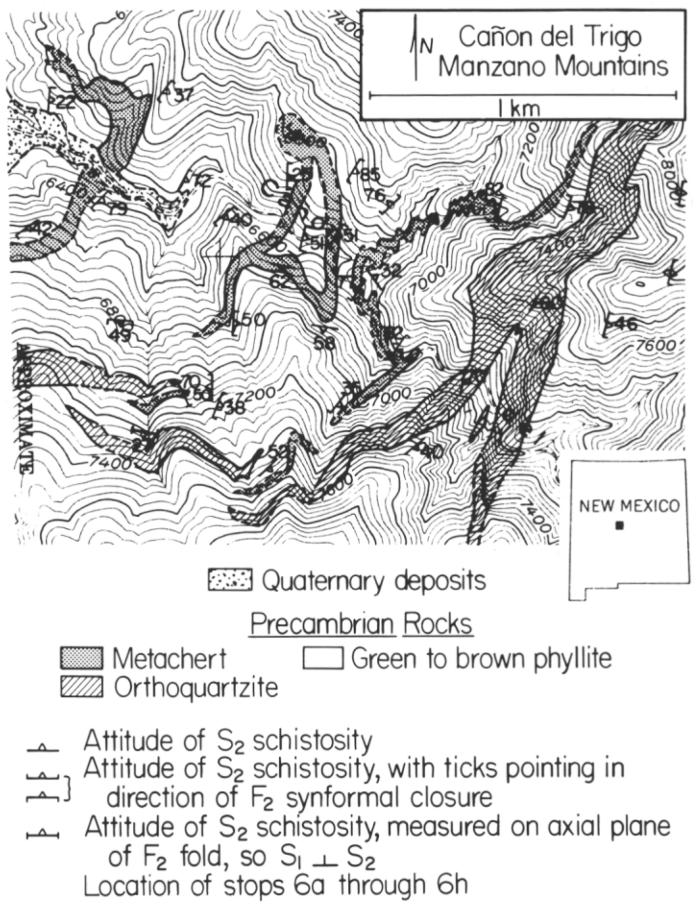

Figure 1. Geologic map of Cañon del Trigo, Manzano Mountains, New Mexico. Foliation symbols identify S_2 cleavage, with data on bedding and other schistosities omitted for clarity. Structural facing symbols (foliation symbols with ticks on their ends) point in the direction of F_2 synformal closure, as indicated by vergence relationships between S_1 and S_2 schistosities.

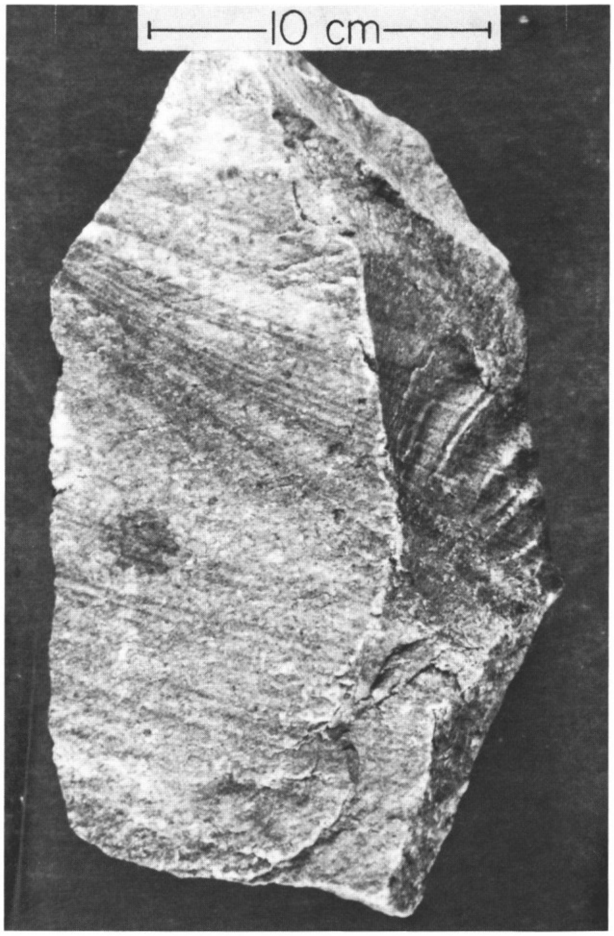

Figure 2. Relict crossbedding in orthoquartzite from Cañon del Trigo. Sample comes from quartzite outcrop in the southwestern corner of Figure 1, at elevation 2,195 m (7,200 ft). This quartzite layer can be mapped directly into massive, featureless quartzite that was interpreted as a "quartz reef" (i.e., vein quartz) by Stark (1956).

Quartzites, however, may not have originated as lenticular sand bodies. Both quartzite and metachert form lenticular pods in the southern part of Figure 1. A quartzite lens lies immediately south of each metachert lens, and successive pairs of lenses are offset relative to one another in en echelon fashion. It seems unlikely that sand lenses should be coupled with chert lenses in a primary stratigraphic relationship, especially with en echelon offsets in each lithology. Instead, a more likely explanation is that these units owe their lenticular shapes to deformation. That is, they were initially continuous layers of quartzite and chert, but the once-continuous layers were broken apart during deformation by transposition or by brittle failure and rigid rotation.

Structural data show that all metachert exposures lie at or near a single stratigraphic horizon. Assuming chert formed continuous layers prior to deformation, this suggests that all exposures of metachert form part of a single, complexly folded sheet.

STRUCTURAL FABRICS

Precambrian rocks in Canon del Trigo show at least three distinct structural fabrics. These can be correlated from unit to unit and used to reconstruct the deformational history of the area. Fabric elements are summarized in Table 1 and are described in this section using terminology of Turner and Weiss (1963).

Phyllites show several foliations. In more quartzose layers, an early schistosity (S,) lies parallel to compositional layering and is defined by

Table 1. Structural fabric elements in Precambrian rocks of Cation del Trigo.

Element	Description
S„	Bedding: recognizable in quartzite as layering outlined by FeTi-oxide minerals, rarely crossbedded, commonly intensely folded. Transposed into S, in metachert and phyllite.
S,	Schistosity: parallel to bedding planes in most outcrops, as a result of transposition. S, forms the dominant layering in metachert, defining a strong micaceous schistosity parallel to layering. Transposed fold remnants are abundant. Phyllites show S, as an early cleavage, largely overprinted by younger structures.
S,	Crenulation cleavage: S, forms the dominant cleavage visible in hand samples of phyllite. Close examination of hand samples, or microscope study, reveals that S2 is a crenulation cleavage folding the early schistosity (S,) in phyllites, largely obliterating vestiges of S,. Adjacent metacherts show S, surfaces as broad, open to tight folds of S, schistosity.
S,	Broad folds or rarely crenulations: best seen in phyllites, the S3 surfaces rarely appear as crenulations of S2. These crenulations are broadly spaced and have small amplitudes. In most of the map area, S, surfaces are defined only by their effect of folding S_2.

parallel alignment of muscovite. This fabric is folded about a second cleavage (S,), which varies from closely spaced crenulations to widely spaced, tight-to-isoclinal minor folds. More typically S, is overprinted and nearly obliterated by the younger schistosity. Minor folds and lineations associated with S, plunge shallowly northeast or southwest (rarely in other directions).

Specimens of metachert show abundant, intense minor folding. The most obvious minor folds warp micaceous schistosity into nearly concentric structures with shallow but variable plunge. Close examination reveals that concentric folds are second-generation structures, refolding earlier isoclines whose axial planes lie parallel to the dominant schistosity. The early folds are poorly preserved, apparently due to intense transposition, but remnants of small, isoclinal closures can be seen in numerous samples. Bedding (S_0) is parallel to schistosity in all samples, a result of transposition during F, folding. Micaceous schistosity and intrafolial isoclines in metachert are parallel to S, schistosity in phyllites, and concentric minor folds in metachert have axes and axial planes parallel to S, surfaces in phyllites. Thus, structural fabrics seem to correlate between the two lithologies.

A pervasive feature in metachert is the presence of intense, closely-spaced lineations on schistosity surfaces. Lineations wrap around minor concentric folds but maintain a consistent geometric relationship to fold axes. Individual lineations are always oriented 90° from axes of the concentric folds, and planes that enclose the curving lineations are exactly perpendicular to fold axes. These features suggest that the curving lineations have not been folded but instead are features similar to slickensides, which formed by slip between adjacent layers of metachert during concentric folding (cf. Wilson and Cosgrove, 1982, p. 18-20).

Poorly preserved, weakly developed crenulations with orientations close to easterly reveal the presence of a third set of S-surfaces (S_3). These weak crenulations are best developed in phyllites, although they are rarely visible in metacherts. The presence of a third set of axial planes is also indicated by trends of S, and S2. Neither shows a consistent orientation in the mapped area: both are folded about S, which strikes about N80°W and has near-vertical dip. Stereograms showing orientations of structural fabrics appear in Figure 3.

DEFORMATIONAL HISTORY

At least three deformational events have affected rocks in the study area. These are referred to as F, , F2, and F_3; they correspond to schis-

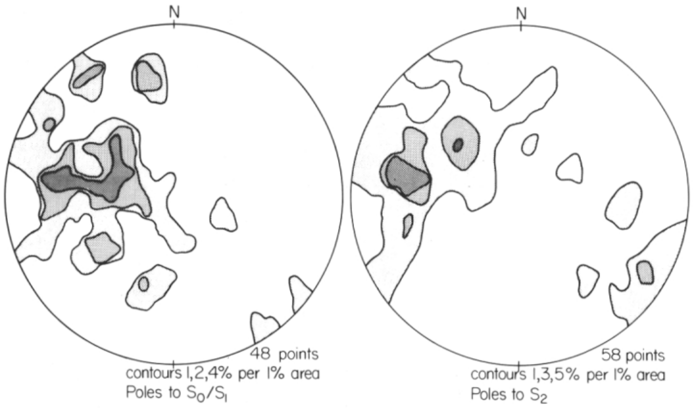

Figure 3. Stereograms showing orientations of S_0/S_1 and S_2 in Cañon del Trigo. Bedding and S_1 are plotted together, because they are parallel in most outcrops as a result of isoclinal F_1 folding. S_3 is not shown due to insufficient data.

tosities S, through S,. Each deformational event has resulted in development of a set of mappable folds. As a result, the geologic map (fig. 1) displays a complex interference pattern.

The purpose of this section is to explain the interference pattern in Figure 1 by unravelling the deformational sequence. This is done using two different methods. First, wherever possible, fold geometry and structural vergence relationships are employed, following techniques of Hobbs and others (1976) or Grambling and Codding (1982). This method is fairly simple. It involves identifying and measuring orientations of bedding and S, through S, schistosities as a first step. Next, relative strikes and dips of a pair of S-surfaces, such as bedding and S, schistosity, are used to locate folds and to identify their relative ages (e.g., F,). Because any number of S-surfaces can be compared and evaluated using this technique, the method affords a powerful tool for studying multiply-folded terranes.

Unfortunately, this method of fold identification fails when two S-surfaces are strictly parallel as a result of isoclinal folding. Such a relationship occurs between bedding and S, in Cafion del Trigo. Therefore vergence data yield no information on the presence of F, structures. To locate F, structures a different, less accurate approach is used. If mapped folds occur without reversals in vergence between S, and S2. or between S2 and S,, they are assumed to be F, folds. This approach should be valid in Carlon del Trigo. Such folds cannot be F, or F, structures, and rock fabrics indicate only three periods of deformation.

Two folds that appear to be F, structures occur in Canon del Trigo. One is antiformal and forms part of a complex dome in the center of the area. The second is synformal and lies about 350 m west of the first. These are indicated in Figure 4 and are identified with question marks because their relative ages cannot be proven. Structural data require only that they are neither F, nor F, folds. To be consistent with the observed map pattern, the two inferred F, structures must form a minor fold pair, as is documented below. Axial planes of F, structures strike between N45°E and N30°W, dipping steeply east, but complex map patterns suggest that the shapes of F, folds have been modified by younger structures.

Several large F2 folds and a number of smaller ones crop out in the study area. These are tight to isoclinal, overturned to the west. All are suggested by the map pattern and confirmed by reversals in structural facing between S, and S2. The most obvious F2 fold is a northeast-plunging antiform which folds quartzite in the eastern part of Figure 3, but numerous smaller F2 folds have significant effect on the map pattern. Additionally, structural disruption of orthoquartzite and metachert may be an **F2** feature. These F2 folds are the structures mapped by Reiche (1949), Stark (1956), and many subsequent workers. Axial

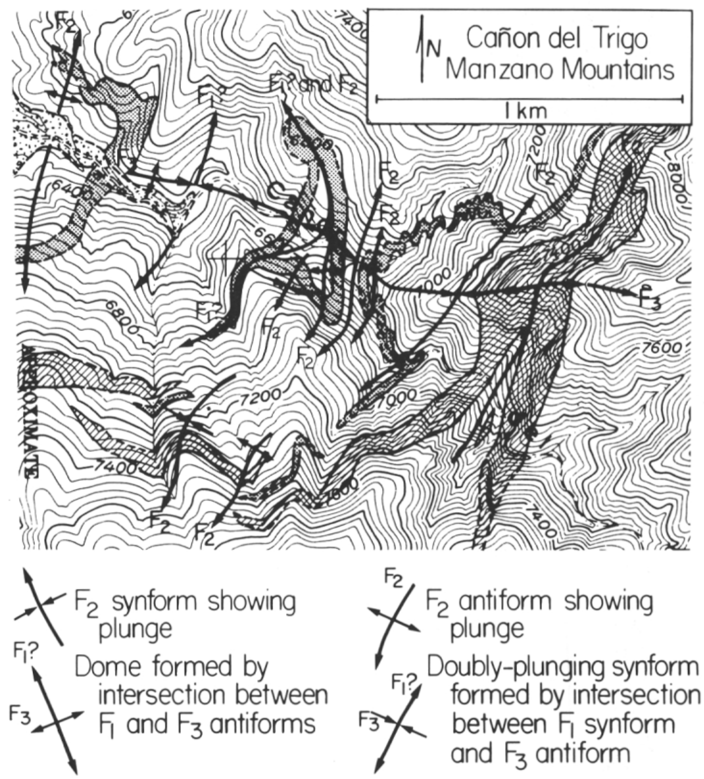

Figure 4. Structural interpretation of map units in Cañon del Trigo. Traces of folds are identified as F_1, F_2, F_3. F_1 structures are identified with question marks because their age cannot be demonstrated conclusively (see text). Arrows show plunges of folds. Note disharmonic folding of the various units.

planes of F, folds have an overall trend near N20°E, 50-60°SE, but local orientations of S2 schistosity are largely controlled by F, structures. Second-generation fold axes show variable plunge as a result of F, deformation.

A single F, fold is exposed in the study area. It forms a broad, relatively open antiform with dips of 30-40°N in its northern limb and 40-50°S in its southern limb. Its axial plane strikes about N80°W, such that the F, fold is oriented nearly perpendicular to earlier structures.

Because rocks in Carlon del Trigo have been deformed three times, the geologic map shows a complex interference pattern involving all periods of deformation (fig. 4). The most obvious manifestation of the interference pattern is the series of complete and partial domes, outlined by metachert, aligned along Callon del Trigo. Each dome shows where the trace of the F, antiform crosses the axial plane of a F, or F_2 antiform. The mapped area contains one complete dome which is an interference feature between F, and F, folds. This dome is slightly modified by F_2 structures. At least three partial domes show F_2-F, interference.

Plunges of F,, F_2, and F, structures are indicated in Figure 4. Figure 5 shows a north-south cross section along the western edge of the mapped area. It illustrates relatively open F, folds, refolding tight F2 structures. The projection exaggerates tightness of F_2 folding, because the projection plane lies at a low angle to F_2 fold axes.

An unusual feature occurs in the cental part of the region, where axial planes of F,, F_2, and F, folds nearly intersect (fig. 4). The intersection results from convergence of F, and F, folds. North of the intersection, axial planes of F, and F_2 antiforms coincide, and F_2 deformation merely acted to tighten a pre-existing F, fold. South of the intersection the F, and F_2 axial planes diverge, and a discrete F2 synform-antiform

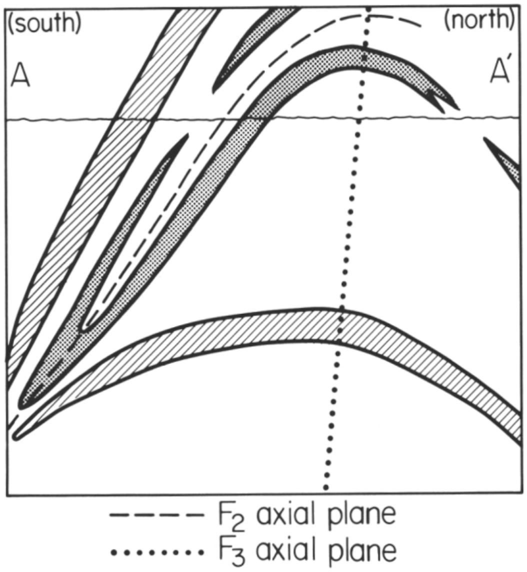

Figure 5. North-south cross section of Cañon del Trigo. Section line is indicated in Figure 4. Wavy horizontal line indicates schematic location of ground surface.

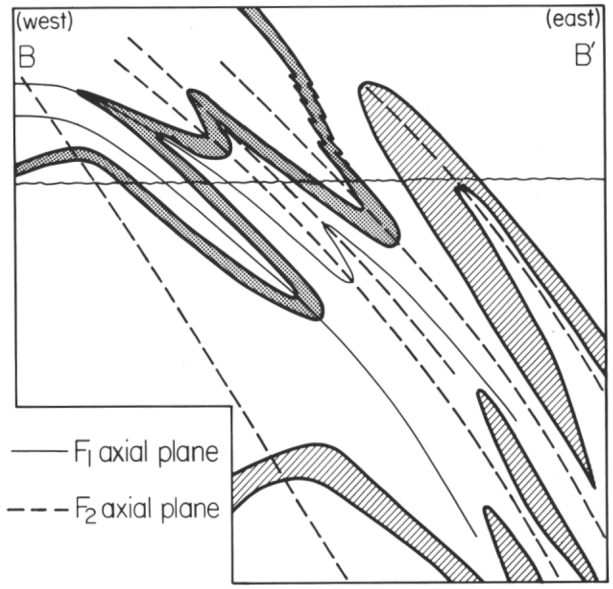

Figure 6. East-west cross section of Cañon del Trigo. Section line trends N80°W, following the general trend of the canyon. It passes just south of the spot where F_1, F_2, and F_3 axial planes nearly intersect, in the center of Figure 4. Horizontal line indicates schematic ground surface.

pair develops as a result. An east-west cross section along Cafion del Trigo, located just south of the point where F, and F, structures diverge, illustrates the F,-F, relationship in this area (fig. 6). If one were to start with Figure 6 and draw a closely-spaced series of east-west cross sections, each located just north of the preceding one, the minor F, synform would diminish in amplitude and the F_2 antiform would gradually merge with the F, fold.

The two F, folds shown in Figure 6 are interpreted as a minor "S" fold pair, because it is impossible to draw a geometrically reasonable cross section otherwise. The sequence of repetition of units, thicknesses of mapped units, and the geologic-map pattern support this conclusion.

CONCLUSIONS

Contrary to previous reports, rocks in Cafion del Trigo have not undergone a simple structural history. They have been deformed at least three times, with each deformation resulting in a set of mappable folds. First-generation deformation accompanied greenschist-facies metamorphism, as indicated by the alignment of chlorite and muscovite parallel to S,. The metamorphic peak, between 425° and 500°C according to mineral assemblages (Hoschek, 1969; Winkler, 1979), occurred after F, folding but before or during F, deformation, because chloritoid and biotite are aligned in S,. Therefore, greenschist-facies conditions prevailed during F, and F, events. Metamorphic pressures were above 1.5 kb based on coexistence of chloritoid + quartz, but they cannot be constrained further using available data.

The total thickness of Precambrian stratigraphy in the mapped area cannot exceed 350 m (figs. 1, 4). With reasonable corrections for unmapped minor folding, it is likely that the stratigraphy is closer to 200 m thick. This compares with Condie and Budding's (1979) estimate of 5,000 m, taken from outcrops in or near Cafion del Trigo. Such discrepancies point out the need for detailed mapping of structural fabrics in addition to lithologies in metamorphic terranes. If results of this study can be generalized to other portions of the Manzano Mountains, then it is likely that much work remains to be done in basement rocks of these other areas.

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REFERENCES

Condie, K. C. and Budding, A. J., 1979, Geology and geochemistry of Precambrian rocks, central and south-central New Mexico: New Mexico Bureau of Mines and Mineral Resources Memoir 35, 58 p.

Grambling, J. A. and Codding, D. B., 1982, Stratigraphic and structural relationships of multiply deformed Precambrian rocks in the Rio Mora area, New Mexico: Geological Society of America Bulletin, v. 93, p. 127-137.

Hobbs, B. E., Means, W. D., and Williams, P. F., 1976, An outline of structural geology: John Wiley, New York, 571 p.

Hoschek, G., 1969, The stability of staurolite and chloritoid and their significance in metamorphism of pelitic rocks: Contributions to Mineralogy and Petrology, v. 22, p. 208-232.

Keyes, C. R., 1920, Geological setting of New Mexico: Journal of Geology, v. 28, p. 233-254.

Myers, D. A. and McKay, E. J., 1971, Geologic map of the Bosque Peak quadrangle, Torrance, Valencia and Bernalillo Counties, New Mexico: U.S. Geological Survey Map GQ-948.

Reiche, P., 1949, Geology of the Manzanita and North Manzano Mountains, New Mexico: Geological Society of America Bulletin, v. 60, p. 1183-1212.

Stark, J. T., 1956, Geology of the South Manzano Mountains, New Mexico: New Mexico Bureau of Mines and Mineral Resources Bulletin 34, 46 p.

Turner, F. J. and Weiss, L. E., 1963, Structural analysis of metamorphic tectonites: McGraw Hill, New York, 545 p.

Wilson, G. and Cosgrove, J. W., 1982, Introduction to small-scale geological structures: George Allen and Unwin, London, 128 p.

Winkler, H. G. F., 1979, Petrogenesis of metamorphic rocks: Springer-Verlag, New York, 5th edition, 348 p.