GEOLOGY OF FORT BURGWIN RIDGE,

TAOS COUNTY, NEW MEXICO

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GEOLOGY OF FORT BURGWIN RIDGE,

TAOS COUNTY, NEW MEXICO

by

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THESIS

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ABSTRACT

A strip in Taos County, New Mexico, eight miles long and three miles wide along New Mexico Highway 3 from U.S. Hill to Talpa, Fort Burgwin Ridge, has outcrops of Precambrian metaquartzite (1800 my bp) and cataclastic granite (1760 my bp) overlain by Mississippian limestone and Pennsylvanian fan delta sediments of the western Taos Trough.

Twelve cross sections are used to demonstrate (1) Pennsylvanian syndepositional vertical movements, (2) Laramide thrusting that folded the Mississippian and Pennsylvanian rocks along splay faults to the east of the Pecos-Picuris Fault (see Fig. 40), and (3) Late Miocene to present-day normal and strike slip faulting, which is related to Rio Grande Graben rifting and appears to have reactivated earlier fault zones.

The north-trending structures appear to be subsidiary to the Pecos-Picuris Fault, a major geofracture that parallels the map area to the west.

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INTRODUCTION

Geography

Taos County is in north-central New Mexico, and it covers 2,256 square miles (Schilling, 1960). The county can be divided into two physiographic provinces: the Rio Grande depression in the western half, and the north-trending Sangre de Cristo Mountains in the southern and eastern parts of the county (see Fig. 1).

The Rio Grande depression is the southern extension of the San Luis Basin of Colorado which locally is called the Taos Plateau. Although most of the plateau is level, its western edge is dotted by low volcanic hills. The plateau is dissected by the Rio Grande Gorge which cuts southward through unconsolidated sediments and volcanic flows. The town of Taos rests at the junction of the plateau and the Sangre de Cristo Mountains, in an indentation in the mountain front called the Taos Reentrant.

The Sangre de Cristo Mountains are the southern continuation of the Rockies. Taos is flanked to the north by the Taos Range, which forms snow capped peaks; to the east by the low, wooded Tres Ritos hills which are almost entirely made of Pennsylvanian sediment; and to the south by the Picuris Range, which is almost entirely Precambrian rock. The field area lies along the eastern border of the Picuris Mountains and includes rocks and structure related to both the Picuris and Tres Ritos provinces (see Fig. 1).

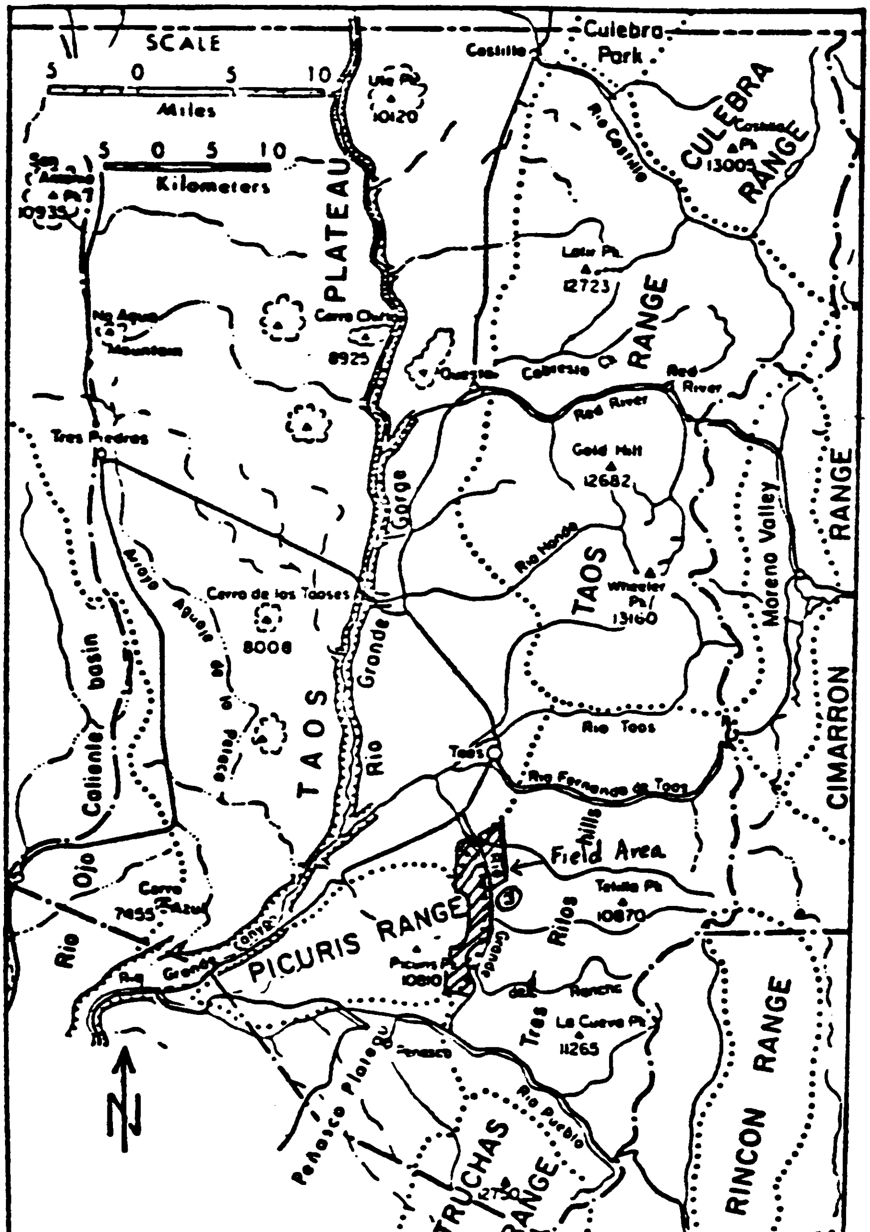


Figure 1. Physiographic map of Taos County. From Schilling (1960).

The climate of the Taos Plateau and the lower reaches of the mountains is semi-arid, receiving approximately 13 in of rain a year. The mountains receive approximately 23 in a year. Consequently, the plateau is covered by sagebrush and grasses, the lower reaches by piñon, juniper and scrub oak and the higher areas are wooded with ponderosa pine and aspen. There is a direct relationship, therefore, between altitude and ground cover and finding good exposures is increasingly difficult as one moves southward, up into the mountains. Furthermore, on the lower slopes, the southern flanks of the hills tend to be less well covered than the northern flanks due to the different number of hours of direct sunlight received per diem.

Location and Access

The area of concern in this thesis involves the southern portion of Taos County along both sides of New Mexico Highway #3 from Talpa to U.S. Hill (see Fig. 1). The highway is paralleled along much of its length by Rio Grande del Rancho which must be crossed to reach outcrops to the west. Fort Burgwin Ridge (named in this paper) lies along the west side of the road. The back of the ridge is accessible from both the north and south by unpaved Forestry Service roads. U.S. Hill can be reached from New Mexico Highway #3 and from a service road which goes to the top of Picuris Mountain. Almost all the land in the field area is in the care of the National Forestry Service.

Scope

Fort Burgwin Ridge is an isolated hill of Precambrian rock which has a prominent double crest and steeply dipping Mississippian limestone beds. This area was chosen for study because Dr. W. R. Muehlberger, my advisor, suspected that the steep dip of the Mississippian rocks was the product of rotation along a Laramide thrust fault. Previous structural studies in the region were of a reconnaissance nature and the best available maps are not sufficiently accurate to make adequate east-west cross-sections of the area. For example, on Montgomery's (1953) map of the area, the double ridge configuration is shown to be the product of northwest-trending fault slivers, but the New Mexico State Highway Department-Geology and Aggregate Resources map (1975) shows these same faults trending due west. The map presented in this thesis shows that the double ridge crest is the product of down to the west normal faulting along a north-northwest trending fault.

The purpose of my thesis is to resolve these mapping disagreements by producing a detailed map and from this new map create cross-sections which show the nature of the Laramide deformation. During the study, the problem became larger and more complex because the field area is located west of the Pecos-Picuris Fault, at the boundary of the Picuris Mountains and the Tres Ritos hills which represent a Pennsylvanian structural trough. My region has had a tectonic and sedimentary history which has been affected by movements along this fault since Precambrian time and reflects the interaction of the two provinces through the Pennsylvanian and Laramide. More recently, the opening of the Rio Grande

Graben and strike-slip motions along the Embudo Fault (see Fig. 43) have produced numerous post-Miocene fault movements within the field area. To create a meaningful interpretation of the structures in the area, it was necessary to unravel the entire tectonic history of the area from Precambrian time to the present.

Consequently, this paper focuses on four main episodes of deformation; the Precambrian, Pennsylvanian, Laramide and post-Miocene. The Precambrian rocks are discussed first. The last chapters of this paper concentrate on a detailed examination of the Pennsylvanian, Laramide and post-Miocene deformations.

The main data for this paper is the structural map presented in the pocket (see Plate 1). The area was mapped during the summers of 1978 and 1979 using a Brunton Compass and standard bed walking techniques. From the data compiled in the field, twelve cross-sections were made (see Plate 2). These cross-sections are the basis for the structural interpretations presented in this paper. Plate 3 is a brief stratigraphic summary.

PRECAMBRIAN ROCKS

Introduction

The Precambrian basement is exposed in north-central New Mexico (see Fig. 2). The basement relief map of Foster and Stipp (1961) shows that basement elevations rise 5,000' or more from the Las Vegas and Raton basins to form the Sangre de Cristo Mountains. The western edge of these mountains is bordered by the Rio Grande Rift grabens which cut the platform approximately down the middle. Further to the west, Precambrian rocks crop out near Petaca, Tres Piedras and the Tusas Mountains. The westernmost margin of the plateau is marked by the Nacimiento and San Pedro Mountains where the basement drops from 9,000' above sea level to 6,000' below sea level in the San Juan Basin.

Previous Studies

Since Just's (1937) and Cabot's (1938) early reconnaissance surveys, much work has been done in the region, especially in the Picuris Mountains. Montgomery (1953) provided a good regional map, established a stratigraphic succession and described the petrography of rocks in the Picuris Mountains. His work has served as the basis of more recent detailed studies. Nielsen (1972) proposed a deformational history based on mesoscopic structural analyses and mineral parageneses. Long (1974, 1976) subdivided the intrusive episodes into four magmatic events. Isotopic studies by Fullagar and Schiver (1973), and Gresens (1975) provide a time reference which,in combination with other data,give a general

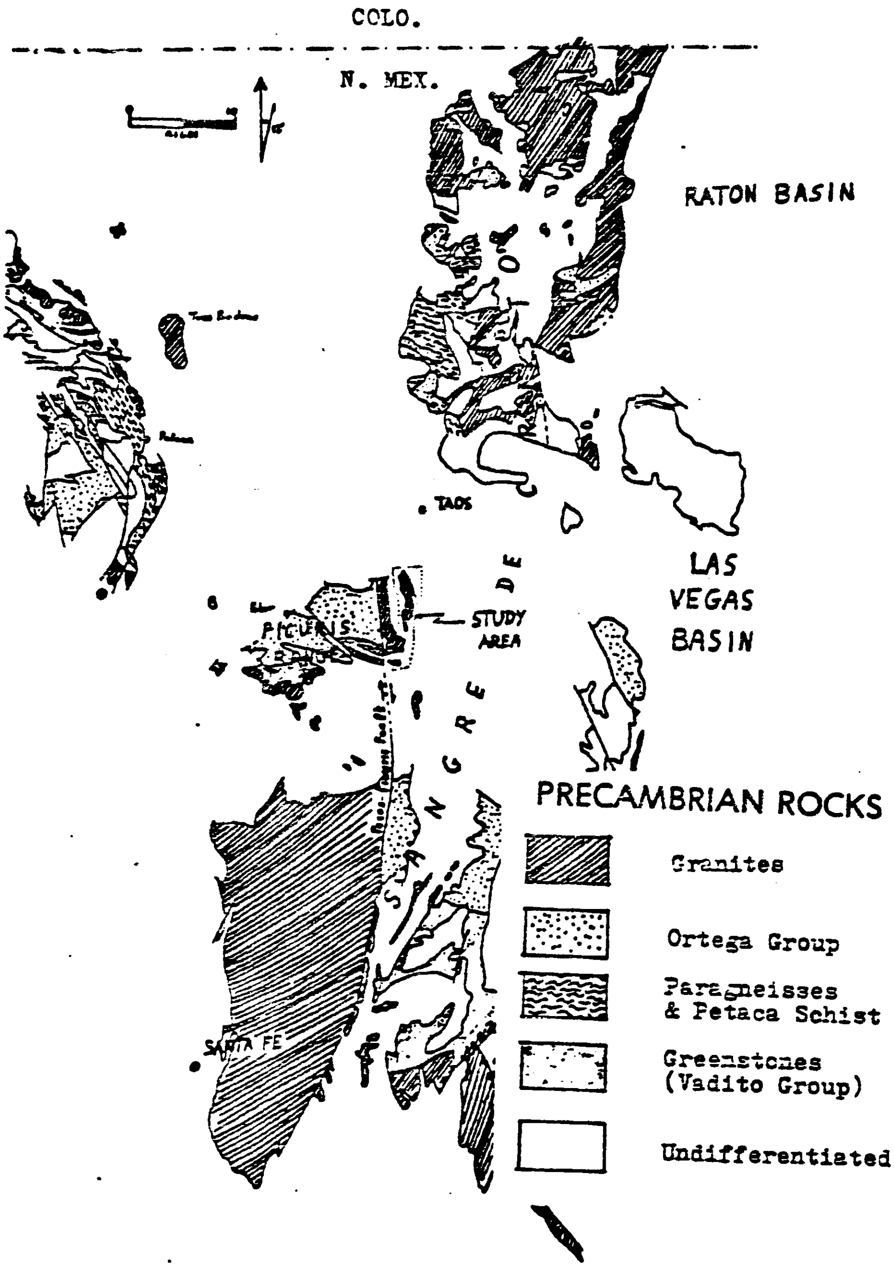


Figure 2. Basement map of northern New Mexico. Adapted from Condie (1979), Robertson and Mench (1979), Barker (1958), and Foster and Stipp (1961).

picture of Precambrian history from approximately 1800 m.y. B.P. to 1200 m.y. B.P.

General Stratigraphy

The Precambrian rocks in the Picuris Range are divided into the Ortega Group, the Vadito Group, and younger granites (see Fig. 3).

Ortega Group. The Ortega Group is considered to be approximately 1.8 b.y. old. The basal unit consists of a thick, typically blue gray, coarse-grained metaquartzite. Montgomery (1953) recognized a schistose, feldspathic quartzite subfacies which occurs on the eastern side of the Pecos-Picuris Fault. He mapped exposures of this rock at Osha Canyon and above U.S. Hill. This same rock is exposed in the Petaca region and is considered to be the metasomatized base of the Ortega quartzite.

The basal Ortega Quartzite is covered by the Rinconada Formation which is an alternating succession of phyllites, metaquartzites, and garnet-staurolite schists. The overlying Pilar Formation consists of carbonaceous slates and garnet phyllites.

Vadito Group. The Vadito Group is a series of metaquartzite and metaconglomerate beds which grade upward into schists and amphibolites with metavolcaniclastic units probably related to the granite intrusions. None of these rocks is present in the field area.

Plutonic Rocks. In the Picuris Range and in the Santa Fe Range, a complex group of plutons called the Embudo Granite has intruded the

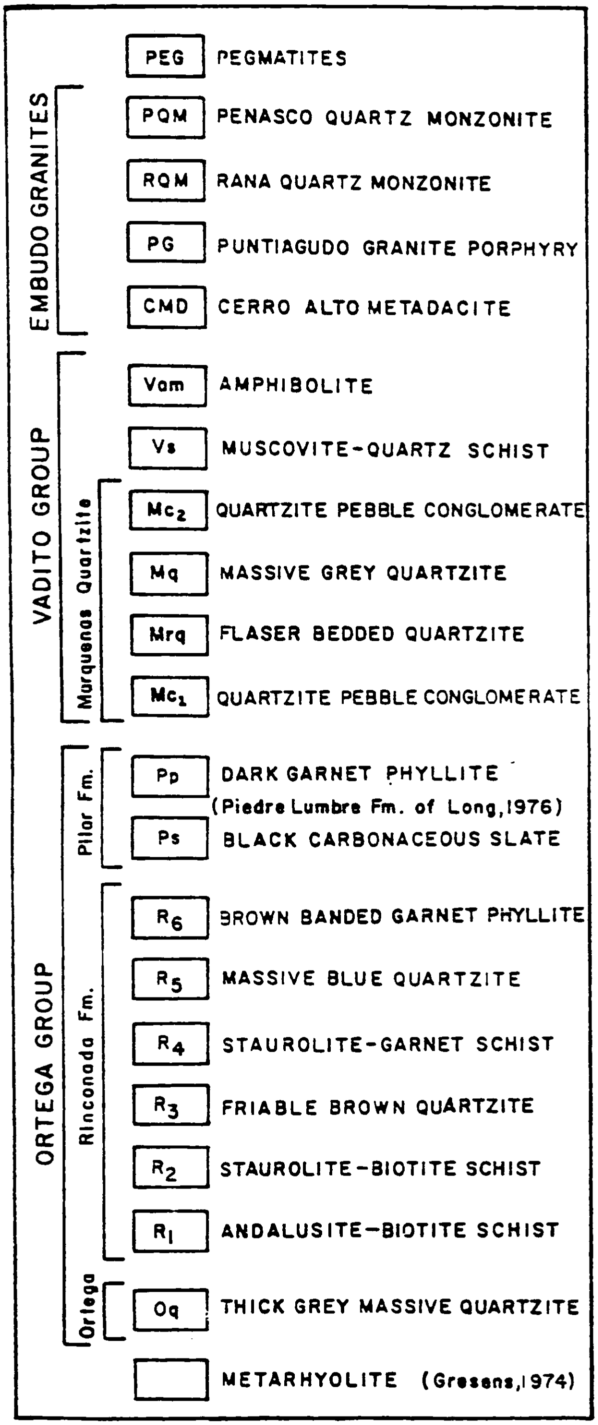


Figure 3. Stratigraphic succession of the Picuris Mountains (from Nielsen (1979)).

sediments (see Figs. 2 and 4). The oldest of these is the Cerro Alto Metadacite which was intruded at shallow depth. On the basis of field relationships, Long (1974) feels that the Puntiagudo granite porphyry is the next oldest rock followed by the deeper-seated Rana Quartz monzonite (Rb/Sr whole rock age 21.7 b.y. by Fullagar and Schiver, 1973). The youngest granite is the Peñasco Quartz Monzonite (21.4 b.y.) followed by aplite dike intrusions and 1.35 b.y. old pegmatites (e.g., the Harding Pegmatite).

Though the Embudo granites are near the granites in my field area, Montgomery (1963) proposed 23 mi of dextral slip along the Pecos-Picuris Fault which separates the Embudo granites from the study area. If this is the case, then the granites of the Petaca and Tres Piedras regions may be more closely related to the granite in the field area. The most notable of these northern intrusives are the Tres Piedras Granite [1.65 b.y. (Barker, 1974)] and the Maquinita Granodiorite.

General Deformational History

I assume that the deformational history of the Picuris Range is typical of that elsewhere in northern New Mexico. Nielsen (1979) has worked out a detailed sequence of events linking major deformations with metamorphic episodes and episodes of granite intrusion (see Fig. 4a). The most important deformation is F_2 which isoclinally folded the Ortega and Vadito sediments and the oldest granites. Axial surfaces of these folds strike east-west and dip steeply to the south. The folds have been truncated by the Pecos-Picuris Fault where they have been dragged sharply to the south. Following the F_2 deformation, metamorphic con-

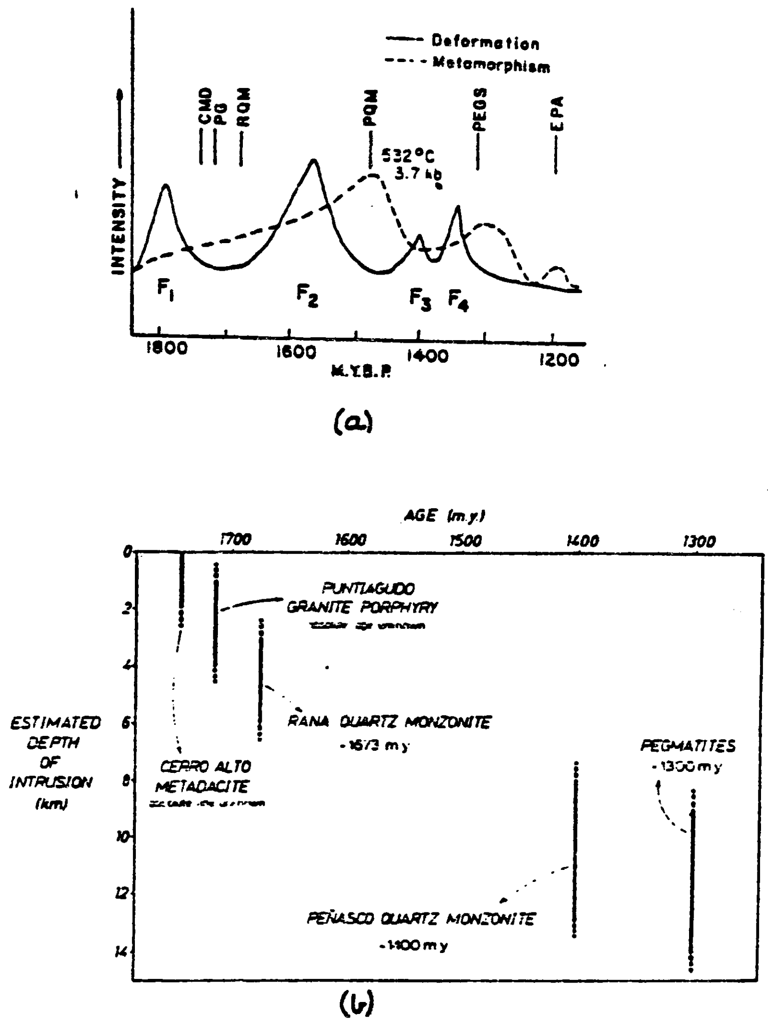


Figure 4. a) Summary of deformational, metamorphic and magmatic events in the Picuris Mountains (from Nielsen (1979). b) Plot of age versus estimated depth of intrusion for the granitic units (from Long (1974).

ditions reached a peak at roughly 530°C and 3.7 kb (Holdaway, 1978). F_3 and F_4 deformational episodes followed F_2 folding and two less intense, less well documented metamorphic episodes are inferred to have taken place between 1.4 and 1.2 b.y. ago.

It is important to note that granite intrusions occurred throughout the interval of burial, folding and metamorphism of the sediments. Granites that predate the F_2 folding should show the effects of deformation, whereas granites younger than the F_2 event should be less affected by stress. Long (1974) suggests that the younger granites were emplaced at greater depths than the older granites (see Fig. 4b). Thus the older granites would have been colder and would have acted more brittly under stress, whereas the younger granites would have been more ductile. The granite within the study area has a cataclastic texture and is presumed to be an older intrusion.

Precambrian Rocks of U.S. Hill and Fort Burgwin Ridge

Two kinds of Precambrian rocks are found in the field area; metasediments and granites. Their exact origin is problematical because they are separated from the Picuris Range by the Pecos-Picuris Fault which is interpreted to have moved 23 km dextrally during late Precambrian time. Therefore, the rocks in the field area were originally located at the same latitude as the Precambrian outcrops of the Petaca and Tres Piedras regions as well as the Taos Range. Complex field relationships and petrographic evidence indicate that the rocks in the field area are similar to those of the Petaca region.

Plutonic Rocks

Field Observations. Several kinds of plutonic rocks occur in the field area (see Plate 1). These consist of a main body of pink, foliated porphyritic granite, numerous aplite dikes, amphibolite dikes, pegmatite dikes, and quartz veins (see Table 1). By far the most important plutonic rock is a large intrusion of porphyritic granite, which commonly appears to be strongly foliated and in places looks like gneiss. It is deeply weathered, forming a grus of microcline crystals. The granite is best exposed along stream walls. Because of extreme fissility of the granite, vegetation covers much of the available exposure and the crest of Fort Burgwin Ridge is covered by a pine forest. Since the aplite and quartz dikes are relatively hard rocks, much of the forest floor is covered by aplite and bull quartz float which gives the impression that much more of this type of rock is present than is actually the case. Examination of stream beds, however, shows that the porphyritic granite is cut by numerous aplite dikes which range up to two feet wide.

The abundance of aplite dikes increases northward and the extreme north end of Fort Burgwin Ridge is largely aplite. Near Ponce de Leon Spring, several six-foot-wide amphibolite dikes are found. They have a composition of 50% plagioclase (An 68), 25% hornblende and the remainder is large amounts of chlorite and epidote. Pegmatite dikes are randomly scattered in the granite; two pegamtite bodies of notably larger size are located in the extreme south end of the map area.

Pegmatite-aplite pairs occur as small dikes near Peñasco and

	THIS STUDY				PICURIS RANGE				PETACA DISTRICT	
	Aplite	Porphyritic Granite			Penasco Quartz Monzonite	Cerro Alto Meta-Dacite	Puntiagudo Granite Porphyry	Rana Quartz Monzonite	Tres Piedras Granodiorite	Maquinita Granodiorite
	TC-15	TC-3	TC-6	TC-7	P171-56	P171-53	P171-44	P171-14	36-D-31	36-B-16
Quartz	34	32.4	24	24	27	41	29	34	42	27
Plagioclase	29	38.6	36.4	38	36	32	37	31	12	57
Microcline	35.6	18	33	29	18	12	24	21	40	1
Biotite	tr	4.6	1.8	2.6	13	10	4	6	3	10
Muscovite	0.8	2.5	0.4	0.8	3	2	4	5	3	4
Epidote	0.8	2.4	0.3	1.7	2	2	1	2	-	1
Chlorite	-	-	-	3.0	-	-	-	-	-	-
Sphene	-	-	0.7	-	1	tr	-	-	-	-
Opaques	tr	1.2	0.8	1.0	tr	1	1	tr	-	-
Apatite	tr	tr	0.2	0.3	-	-	-	-	-	-
	Tom Chapin (1981)				Long (1974)				Barker (1958)	

Table 1. Typical modes of the granitic rocks.

sporadically in the Picuris Range. They are considered to represent one of the latest stages of granitic magmatism. The pegmatites are considered to be younger than the aplites and quartz veins. The aplites may represent the residual liquid of the granites that enclose them inasmuch as they are restricted to the granites (Long, 1974). The amphibolite dikes are younger than all other Precambrian rocks and are presumed to be related to tear faulting in late Precambrian time (Montgomery, 1963).

The porphyritic granite shows several interesting features in the field. For example, megascopic cataclasis has penetrated most of the granite resulting in small-scale dislocations of quartz and aplite veins which form systematic step faults along parallel shears. In an attempt to measure this type of dislocation, no consistent pattern was found. Microcline cyrstals also show the effects of shear, and an individual crystal typically may be separated as much as ¼ inch along a microfracture.

One interesting megascopic feature of the granites is the occurrence of long parallel ribs and spines that trend approximately north-south. Montgomery (1963) said that in the Picuris Range they represent large silicified gouge zones that parallel the Pecos-Picuris Fault. They have been observed to occur at large distances from the fault, suggesting that the zone of deformation is very wide. Mapping of some of these outcrops (see Plate 1) indicates that the outcrops form straight lines, implying that they are steeply dipping. If one rotates the granite 80° so that the Mississippian peneplain surface is horizontal, then the shears would also be nearly horizontal. Muehlberger (personal communication) suggested that they may be related to stresses induced during the emplacement of the granite, or to a later period of faulting.

Though these shear features trend parallel to faults mapped within the field area, the former do not mark discrete zones of slip. More recent faults show slickensides, mullions, megascopically visible zones of penetrative shear, and generally a gouge zone of altered rock flour.

Clearly the porphyritic granite has been intensely deformed. Unlike the more competent aplite dikes which weather into 4" conchoidal blocks or have remained undeformed, the granite appears to have reacted to stress by both microscopic and megascopic slip. Therefore, the granite must be considered as an incompetent or semi-competent member.

Petrology and Origin of the Main Granite. The main body of granite is a coarse-grained, cataclastic, porphyritic granite. The most striking feature of the granite is its cataclastic texture. Euhedral to subhedral Carlsbad-twinned microcline phenocrysts up to 13 mm long are surrounded by rounded and fractured plagioclase crystals (moderately altered to sericite), flasers of recrystallized and strained quartz, unoriented muscovite in flakes and felted pools, epidote associated with biotite, muscovite, chlorite and opaque minerals, and small, extremely fresh microcline crystals which appear to have recrystallized from the crushed matrix.

Evidently the granite has experienced several phases of metamorphism. The recrystallized microcline, in the presence of plagioclase, quartz and muscovite, suggests metamorphism somewhat below the stability field of the reaction muscovite + quartz $\rightarrow$ K feldspar + Al_2SiO_5 + H_2O at low pressure. The temperature of this reaction is depressed by the presence of plagioclase. The presence of muscovite is complicated by evidence for retrograde muscovite alteration and retrograde formation of epidote.

Both the cataclastic texture and similar metamorphic effects have been observed elsewhere. Barker (1958) notes that the Maquinita, Tusas Mountain, and Tres Piedras granites show porphyritic textures with flasers of quartz and faint-to-pronounced foliation. The foliation indicates that the granites were emplaced prior to the main Precambrian deformation.

Nielsen (1979) worked out a deformational history for the Picuris Range. He observed that the greatest deformation, F_2, (see Fig. 4a) occurred prior to the major peak of metamorphism on the basis of the chloritoid-Al silicate association in the Ortega and the staurolite-almandine association in the Rinconada Formation. Long (1976) noted some recrystallization of muscovite, apparently followed by epidote formation, associated with F_3 folding in the Rana Quartz Monzonite.

The Rana Quartz Monzonite has a coarse-grained foliated texture. Biotite grains are partly shredded and recrystallized, and are crosscut by muscovite grains. Plagioclase is partly granulated and altered to sericite, epidote and clinozoisite. These textures were observed in the porphyritic granite of the study area.

If one assumes that the deformation in the Picuris Range occurred regionally, then the most likely time of emplacement of the porphyritic granite in my field area was prior to the peak of metamorphism, between

Nielsen's (1979) F_1 and F_2 deformations (see Fig. 4a). On the basis of textural similarities, the granite in the field area probably underwent the same kind of deformation and metamorphism as the Rana pluton. Therefore, the granite is probably as old as 1.7 b.y. and was emplaced at shallow depths (see Fig. 4b).

Because of the extreme internal variability of the granites and the possibility of local contamination by wall-rock inclusions, the modal analyses may not be typical of the pluton as a whole. The granite in the study area best matches the modal analyses of the Puntiagudo Granite Porphyry and the Rana Quartz Monzonite (see Table 1). The granites of the Petaca region do not resemble the granite in the field area but, given that there is a 23 mi offset on the Pecos-Picuris Fault during the Late Precambrian, the granite was probably located more than 20 km from any of the available granite exposures; hence, any compositional correlation is tenuous at best.

Metasediments

Several outcrops of a gray green, schistose, feldspathic muscovite metaquartzite occur in the south end of the field area (see Plate 1). They appear to have capped the granites in the area, later to have been tilted to 45°S by post-Pennsylvanian faulting. Closely associated with the sediments is a distinct lens of pure sugary muscovite schist which extends 4,000' east-west, dips 54° south and is approximately 700' wide. Located at the easternmost end of this schistose lens is a pegmatite which probably provided the fluids that altered the lens.

These metasediments belong to the lowest portion of the Ortega

Group. Sutherland (1963) considered these rocks to be the schistose facies of the lower Ortega. On the basis of my detailed mapping and literature search, I conclude that the assemblage described above is unique to the Petaca district in New Mexico. This observation is consistent with considerable dextral offset along the Pecós-Picuris Fault and suggests that the Petaca Schist should be considered to have extended further to the east than has been previously proposed.

In the Picuris Range, the Lower Ortega consists of a thick, blue-gray, coarse-grained, vitreous metaquartzite with andalucite- and kyanite-rich layers. In the Petaca region, the Ortega quartzite grades laterally into the muscovite-rich Petaca Schist. This relationship was first observed by Just (1937) who proposed that the schist represents metasomatized Ortega. The close relationship of muscovite-rich quartzites to pegmatite intrusions and muscovite-rich halos and lenses surrounding the pegmatites makes this proposal attractive (Just, 1937; Jahns, 1946; Barker, 1958). However, the quantity and bedded nature of the muscovite in these rocks suggests that much of the muscovite represents depositional layering. It seems unlikely that the abundance of mica and its bedded nature could have been produced by cross-cutting intrusions.

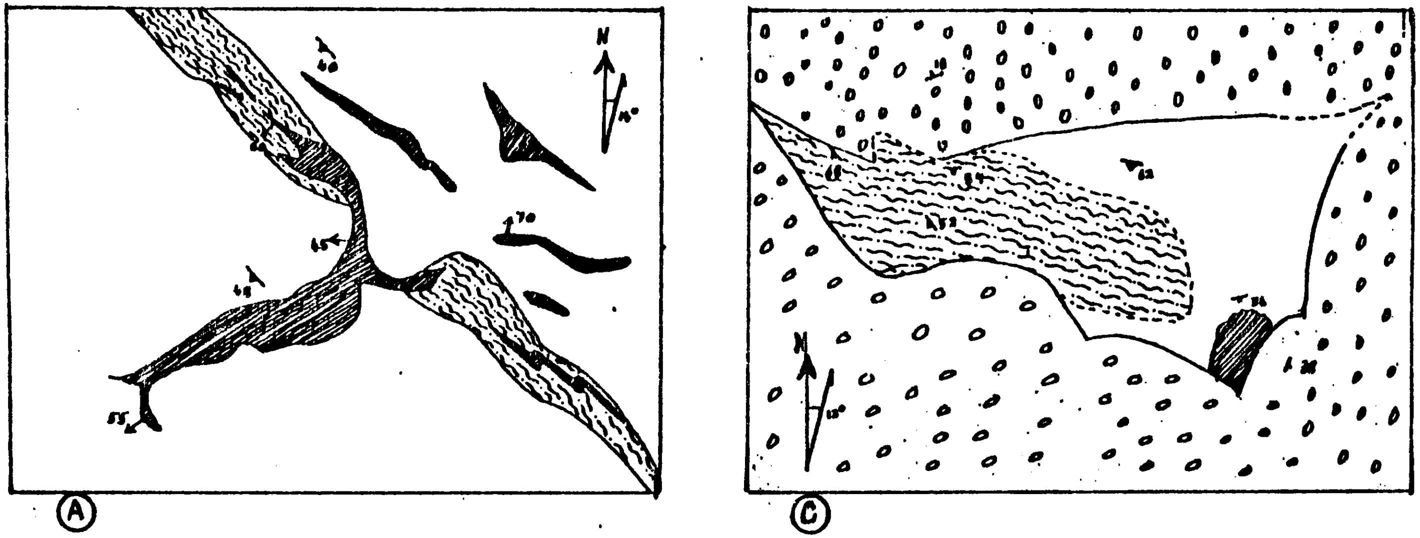
Metasediments in the field area resemble the Petaca Schist both petrographically and in field relationships. Megascopically, the rocks in my area have a wide range of muscovite content which increases dramatically towards the mica mine at U.S. Hill. At the greatest distance from the mine (two miles to the north) the rocks are found in three-foot slabs of gray-green, muscovite-rich quartzite. Further south, muscovite content increases sharply and the rocks become more foliated and thin-bedded.

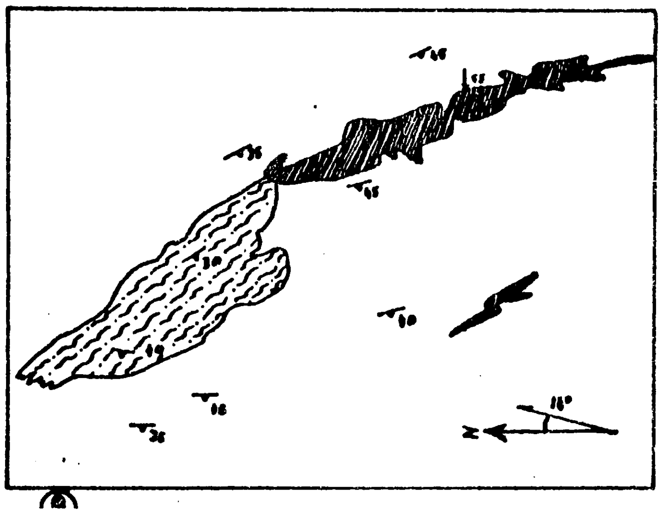
The granite-schist contact is extremely gradational, and in fact it can be missed quite easily. Just (1937) observed the same relationship in the Petaca region, stating that the schistose quartzite was a barrier to the granite and acted as a roof. Stoping and reaction with the intrusive resulted in a mixed region. He describes a gradational zone which is over one mile thick. The granites in the south end of my field area have the same contact between the granite and the schist and consequently the contact is mapped approximately on my field map.

As one approaches the pegmatite body (extreme south end of the map), mica content increases. A thin section of typical wall rock near the mica mine gives a modal anlaysis of 60% quartz, 25% muscovite and 15% microcline and plagioclase. The rock is highly foliated in which muscovite parallels the schistosity. This mode is entirely consistent with the descriptions of Petaca rocks by Jahns (1946) and Barker (1958).

The pegmatite at U.S. Hill (mined for mica during the 1940s) is made up of quartz, green muscovite and large albite crystals. To the west of the pegmatite, a lens of white, sugary muscovite schist containing large stretched boulders of bull quartz is presently being mined by Mineralogical Commodities of America to make spackle (see Fig. 5). Their ore consists of over 90% muscovite rock with a white crumbly schistose texture. The foliation within this body parallels that of the wall rock, whereas the pegmatite body is discordant to the wall rock.

On the basis of petrology and an intricate metasomatic relationship to other Precambrian rocks including the mica mine assemblage, the





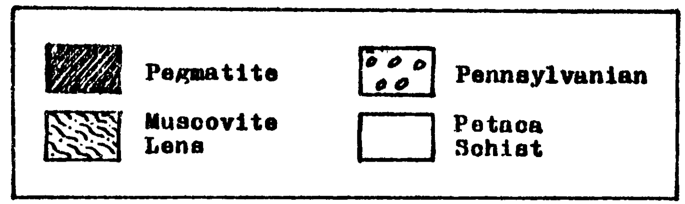


Figure 5 . Comparison of muscovite deposits:
a) Schematic diagram of the Queen Mine;
b) Schematic diagram of the Alma Mine;
c) Schematic diagram of the deposit in the field area. (a & b adapted from Jahns (1946))

metasediments in the field area are similar to Petaca Schist. Furthermore, if this is so, several observations can be made: (1) further evidence of a large displacement on the Pecos-Picuris Fault has been provided, (2) the Petaca Schist extended farther east than has been previously observed, and must underlie the Rio Grande Graben sediments, (3) the same kind of pegmatites were intruded into the Ortega, indicating that a very large body of granite probably underlies my area and (4) the Petaca Schist may be too large to attribute to metasomatism - it could represent a facies change within the Ortega.

Conclusions

The granite above Fort Burgwin Ridge and U.S. Hill need to be dated. The field observations and preliminary petrographic work presented here strongly suggest that the granite is approximately 1.7 b.y. old. The main evidence presented is that: (a) the cataclastic texture implies that the granites were intruded at shallow depth prior to F_2 folding (see Fig. 4a), (b) the presence of recrystallized microcline indicates that the granite experienced metamorphic conditions of 530^{o}C and 3.7 kb after deformation. These observations are consistent with an older age for the granites.

The metasediments in the region have the same assemblage as has been observed in the Petaca district.

ARROYO PEÑASCO FORMATION

Regional Description

Mississippian rocks of the Arroyo Peñasco Formation are of Osage to Meramec age (Armstrong, 1958a). They consist of fine-grained orthoquartzites overlain by conformable limestone. Outcrops range from 10' to 130' in thickness and have been recognized in the following mountain ranges: San Pedro, Nacimiento, Jemez, Sandia, Manzanita, Manzano, and the Sangre de Cristo of northern New Mexico.

The formation consists of a basal unit of 2 to 60' of quartz conglomerates, sandstones, and thin shaly beds. These interfinger and grade upward into three incomplete carbonate cycles (Armstrong, 1958a). The lowest (Cycle 1) contains dolomites, dedolomites and nodular chert lenses containing Late Osage microfauna. These are overlain by shallow marine carbonate mudstones followed by stromatolitic intratidal and supratidal carbonate rocks. Cycle 2 rocks, which are not found in the study area, are shallow marine intertidal wackestones, limestones and dolomites. Cycle 3, also absent, consists of shallow marine wackestones, and arenaceous oolitic packstones capped by subtidal lime mudstones.

Late Mississippian and Early Pennsylvanian uplift resulted in extensive erosion and removal of the Arroyo Peñasco Formation. Prior to Pennsylvanian deposition, the Mississippian strata in western central and northern New Mexico were folded and faulted, and subjected to extensive removal and truncation. Presently, only discontinuous disconformable erosional remnants are exposed from underneath the Pennsylvanian cover.

Previous Work

The Mississippian rocks of Fort Burgwin Ridge are part of a widespread group of rocks which crop out throughout New Mexico. Various names and ages have been assigned to these rocks (see Fig. 6). The classification of Read (1944), Baltz and Bachman (1956), Baltz and Read (1959) divides these pre-Pennsylvanian rocks into two formations. The basal sandy unit they call the Espiritu Santo Formation; the upper limestone member they call the Terrero Formation. They link these formations to rocks in Colorado. Armstrong (1955, 1958a, 1958b, 1967) disagrees with the above authors. On the basis of Meramec Endothyra fauna, he gives the rocks a Mississippian age and he considers the sandstone and limestone to be members of the same formation which he calls the Arroyo Peñasco Formation. He uses the rocks which crop out on Fort Burgwin Ridge for his type section of the basal sandy unit and the lower portions of the limestone upper member of the Arroyo Peñasco (Cycle 1).

Sutherland (1963) recognized Baltz and Read's nomenclature but added his own revision. He removed the basal sandstone from the Espiritu Santo Formation and reclassified it as the Del Padre Sandstone. In this he included 754' of conglomeratic sandstones found at his Rio Chiquito section (see Fig. 7) north of Truchas Peak near the Pecos-Picuris Fault. However, there is no fossil record in this section. Much of what Sutherland assigns to the Del Padre is coarser and much thicker than the basal sandstone of the Arroyo Penasco, and moreover, the top of the section is conformable to the overlying Pennsylvanian sediments.

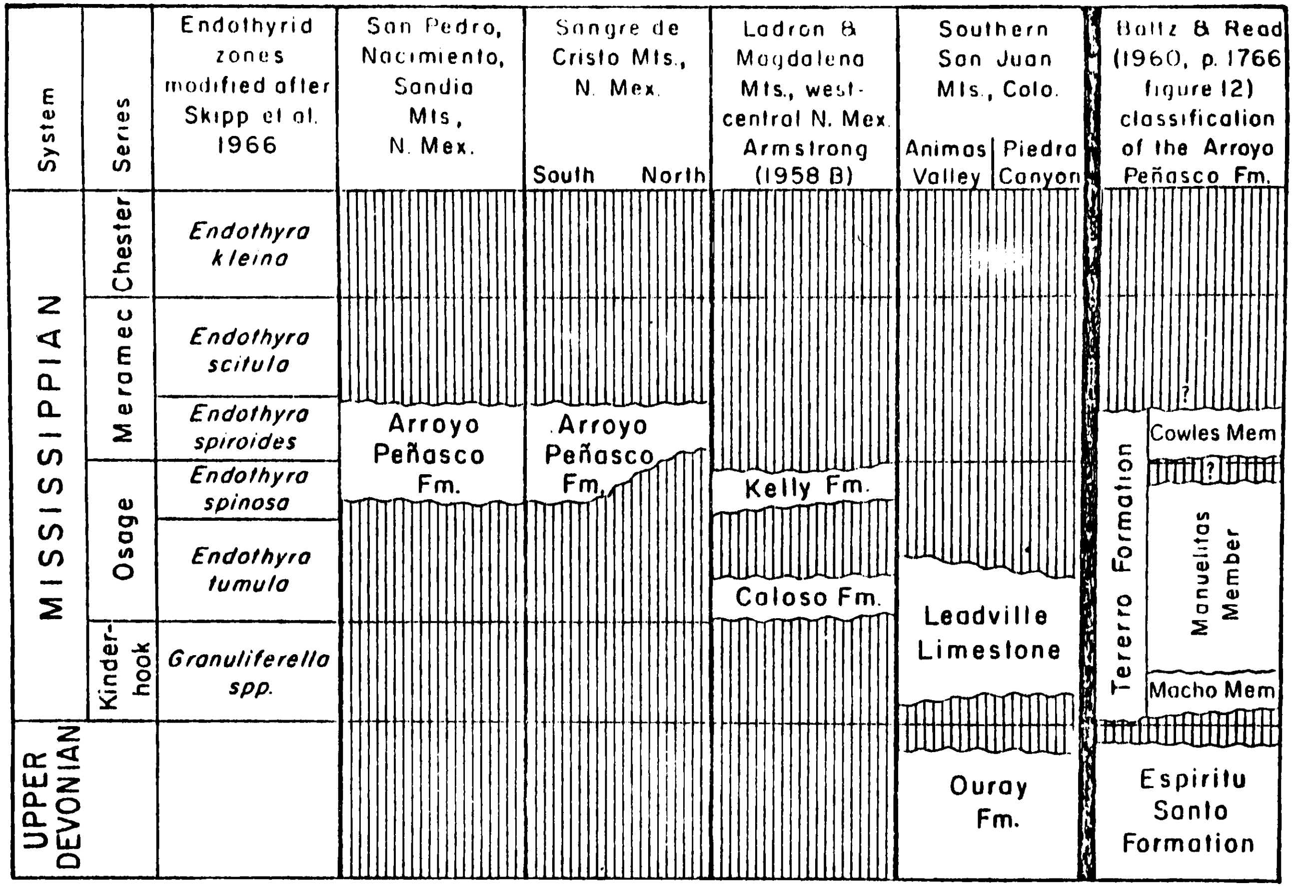


Figure 6. Correlation Chart of Mississippian Rocks of North-Central New Mexico and Southwestern Colorado. From Baltz and Read (1960).

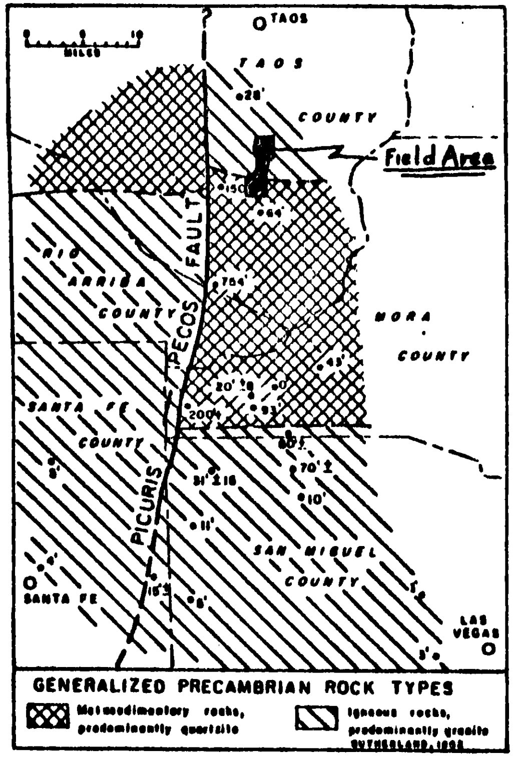


Figure 7. Points of known thickness of the Del Padre Sandstone in relation to the distribution of Precambrian Rock types. Note the anomalous thickness of the sandstone near the Pecos-Picuris Fault. Adapted from Sutherland (1963).

There is reason to believe that much of his Del Padre Sandstone is a local, unrelated Pennsylvanian alluvial fan deposit.

Recently, Turner Williamson (1978) extensively studied the basal sandstone above Ponce de Leon Springs. In his master's thesis, he adopts Armstrong's nomenclature because it is more consistent with his observations. For the purpose of this paper Armstrong's classification will be used; namely, the Arroyo Peñasco Formation which is made up of a basal sandstone unit, and overlying Mississippian limestone units. For comparison to Baltz and Read's correlations, see Figure 6.

Basal Clastic Rocks

The basal clastic rocks of the Arroyo Peñasco Formation were first studied by Armstrong (1955) who used the area above the Ponce de Leon Springs as his type section. Williamson (1978) followed by measuring several sections and doing detailed petrographic work. Here is a brief summary of his work.

Arroyo Penasco basal clastic rocks were deposited unconformably on a Precambrian peneplain. In many places they overlie thick granitic grus which has been interpreted to be a near-shore arid regolith. 6 to 15 meters of platy, pale yellow, unfossiliferous orthoquartzites were deposited on this surface. These rocks crop out along the crest of Fort Burgwin Ridge as steeply dipping, highly resistant beds.

The quartzites are extensive tabular beds, ranging from a few meters to one meter thick. The tops of the beds are scoured by the overlying beds and they are cut by several large, lenticular channels. Up-section, the sandstones merge into less defined, less resistant clastic carbonate layers. The top of the section is capped by a massive, gray, extremely resistant limestone approximately 28 meters thick.

Williamson proposed a depositional system of marine, tidal flat, beach, and braided streams which prograded and regressed over the Precambrian peneplain. The fine grain size and uniformity of the bedding indicate that tectonism during this time was nil.

After deposition, the unit underwent shallow-water diagenesis. Williamson recognized an extensive authigenic suite of minerals. These are principally quartz overgrowths followed by illite, chlorite, calcite, dolomite, chert, and chalcedony. As a consequence of these alterations, there is practically no porosity in the basal quartzite of the Arroyo Penasco Formation. The lack of porosity and the extreme induration caused by the silica cementation make this unit easily distinguishable from the more porous sandstones of the overlying Pennsylvanian sediments.

Carbonate Rocks

The clastic Arroyo Penasco rocks grade conformably into, or interfinger with the carbonate section. The transition zone is less than two meters thick, passing from a sandstone with clay cement, to a sandstone with calcite and spar cement, to a sandy micrite. In the Fort Burgwin Ridge area, the average outcrop width of the carbonate section is approximately 25 meters. This unit thins markedly south of Fort Burgwin, where it is only a few meters thick, or is locally absent. These resistant carbonate rocks form rounded ridge crests of steel gray to pearly gray dedolomites and coarse crystalline limestones which weather cinnamon brown. An unusual and characteristic feature of these limestones is large bowling-ball-size chert nodules, and one-meter-long chert lenses found in the middle and basal parts of the section.

Only carbonates from Armstrong's Cycle 1 are preserved in the area. Armstrong (1955) measured and described a 28 meter section above the Ponce de Leon Springs. He found 25 meters of subtidal to intertidal limestones with numerous algal stromatolites. Macroscopic sedimentary features include laminations and intraformational conglomerates. Microscopic studies of the chert nodules, which preserve the original depositional fabric, reveal that the limestone was a calcisphere, ostracod, foraminiferan, pelletoid, lime mudstone, possibly deposited in a shallow restricted marine environment, or a calcisphere, sponge spicule, algal, stromatolitic lime mudstone deposited in an intertidal environment.

The Arroyo Peñasco limestones are easily distinguished from the overlying Pennsylvanian limestones. They are characterized by their steel blue color, lack of mud and clay, rarity of fossils, and presence of large chert nodules. The Pennsylvanian limestones, on the other hand, are generally muddy, dark gray or black, fossiliferous, and generally thin bedded. The Upper Desmoinesian limestones, which are steel gray and much thicker, contain abundant crinoid fossils.

Outcrops and Mapping Criteria in the Study Area

The Arroyo Penasco Formation crops out as a double spine along the crest of Fort Burgwin Ridge. It trends generally north-south, forming a gentle arc with dips of up to 80°E at the northern end of the hill, but with progressively smaller dips towards the south. Dips average 48°

at the southern end of the ridge. The formation has been displaced approximately 1,000' perpendicular to bedding by a north-south trending normal fault (see fault #1 on Plate 1). One half mile south of the Ponce de Leon Springs, the Arroyo Penasco can be seen in contact with the fault. Further to the south the fault appears to form a series of en echelon anticlines which expose Mississippian rocks along the eastern flanks of Fort Burgwin Ridge (see Fig. 8). At the southernmost end of the hill, the Arroyo Peñasco is much thinner and the limestone members disappear. A fault offset is evident at this southernmost end.

Just south of U.S. Hill a second occurrence of Arroyo Penasco dips 72° northwest and is overturned. These are interpreted to be on the east side of a Pennsylvanian north-trending fault which has displaced the outcrop approximately one half mile to the north.

The Arroyo Peñasco nonconformably overlies the Embudo Granite. What may be an 8 meter grus layer is in direct contact with the Mississippian rocks. That is, this feature may be a grus layer or a tectonic effect induced by the tilting of the overlying strata. The contact of the Arroyo Peñasco and aplitic granite in the north end forms a talus of 3" to 4" diameter aplite blocks.

The Arroyo Peñasco is easily recognized. The basal quartz member is a fine-grained, pale tan sandstone which forms laminar beds 2" to several feet thick. It is distinguished from the Pennsylvanian sandstones above it by its laminar nature, fine to medium grained texture, mineralogical purity, and intensive cementation which gives the rock its smooth texture and "ring" when struck by a hammer. The Pennsylvanian sandstones, on the other hand, are poorly sorted, dirty, and diagenetically "immature" with much higher porosities.

Sutherland (1963) mapped the Mississippian sandstones in the southern end of the field area. His map shows Mississippian sediments surrounding a Precambrian outcrop of Ortega Quartzite. Intensive exploration indicates that there are no Mississippian sediments on the north side of this hill. On the south side of the hill a brown platy sandstone is present, capped by limestone and sandstone beds. Though the sandstone is in many ways similar to the basal Arroyo Peñasco quartzites, the cyclic nature of this sequence suggests a Middle Desmoinesian age for these sediments.

Sutherland's mapping of Mississippian sediments along the southernmost flank of Picuris Mountain is also of questionable validity. The contact between the Embudo Granite and the younger sediments is easily distinguishable, but where the Ortega quartzite and the sandstones are in contact, they show very little contrast. In this study the contact is mapped on the basis of the appearance of north-south jointing in the Ortega which is absent in the overlying sedimentary rocks (see Plate 1).

The sandstones are determined to be Middle Desmoinesian age in accordance with the following observations. The sandstones along the contact contain very coarse sand with immature feldspar pebbles up to one centimeter long. The beds show lenticular scalloped breakage which may be cross-trough bedding. They contain some quartz cement but retain most of their porosity and have a rough texture. The sandstone beds are laterally extensive sheets 1 to 3 meters thick. As many as ten coarse sand cycles are mappable. These probably represent the wet alluvial fan, braided stream facies of the Early to Middle Desmoinesian described by Casey (1980), who feels that the beds located above U.S. Hill represent the margin of a large fan delta lobe whose center lies between Rock Wall and Tres Ritos (personal comm.). Certainly these rocks bear little resemblance to the clastic Mississippian rocks seen elsewhere in the map area.

Armstrong (1967) states that Sutherland mapped 750' of Del Padre sandstone at Rio Chiquito which consist of massive quartzitic cross-bedded sandstones. This section appears to be a high-energy, nearshore deposit which grades upward into Pennsylvanian sediments. The Mississippian sandstones that he mapped south of Picuris Mountain bear a resemblance to the Rio Chiquito deposits and were probably mapped as Del Padre using the same criteria (see Fig. 7).

Armstrong (1967) objects to Sutherland's classification in light of the following interpretations:

(a) The Arroyo Peñasco was deposited on a relatively flat peneplain with subdued topography.

(b) None of the highlands in Colorado or New Mexico contributed much clastic material to the Mississippian seas.

(c) The sedimentary record of Colorado and New Mexico does not indicate any tectonic activity in Mississippian time prior to Late Meramec or Chester times.

(d) The Mississippian carbonate rocks were never thicker than 175' and were deposited in less than 50' of water. Most of them were deposited in intertidal or subtidal environments.

Therefore it is unlikely that 754' of high-energy shallow marine sandstones would be deposited during a quiescent period characterized by very moderate clastic influx and limestone deposition. On the other hand, in Pennsylvanian time, the tectonic and marine conditions were suitable to have created this kind of deposit. Moreover, the top of the "Del Padre" at Rio Chiquito grades conformably into the Pennsylvanian sediments.

Definitely, these coarse sandstones are dissimilar to the Mississippian section seen in the remainder of the field area. They are much too immature, coarse grained, and show no evidence of tidal working. Furthermore, they appear to be conformable to the overlying Middle Desmoinesian sediments. The basal contact contains clasts of Ortega quartzite up to 6 cm long, indicating a very proximal source and a high-energy environment created by tectonic movements. I have remapped these beds as Middle Desmoinesian sandstones, deposited during a time of active vertical movements.

Paleo-history and Tectonic Activity during the Mississippian

The Arroyo Penasco Formation preserves the record of the maximum advance of the Mississippian seas in New Mexico. The transgressions began in the early part of the period, as progressive flooding from the east extended the seaway from the Cordilleran miogeosyncline through to the Paradox Basin. An arm of the sea reached around the northern flank of the Zuni Defiance Highland and covered the present-day San Juan Basin.

During Osage time, a second phase of transgression completely covered the low saddle between the Pedernal Highlands and the Zuni Defiance Uplift (see Fig. 8). A transgressive sequence of conglomerates

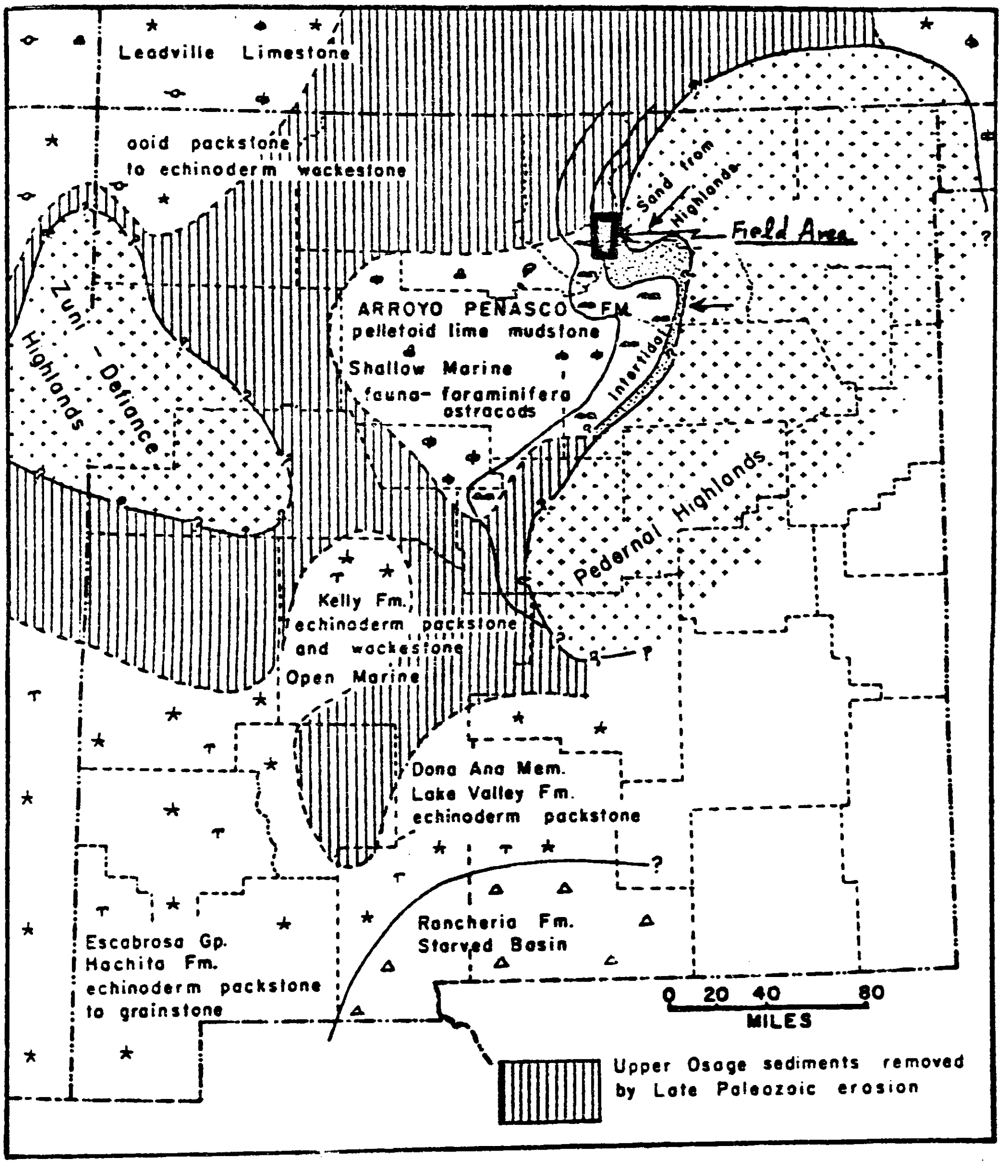


Figure 8. Restored diagrammatic representation of carbonate facies and depositional patterns in New Mexico at the initial stage of the Arroyo Peñasco carbonate sedimentation. From Armstrong (1955).

and sandstones was transported westward from the Pedernal Highland and deposited by braided streams onto a marginal system of tidal flats and beaches which fringed the shallow seas. This period was ended by a short pulse of subsidence followed by a period of stability. A system of carbonate offlaps covered the sands. These consisted of supratidal carbonate facies overlying intertidal stromatolitic algal mounds which in turn overlay a subtidal marine limestone facies. This system was reactivated at least three times by renewed episodes of subsidence, forming the Cycle 1, Cycle 2 and Cycle 3 units of the Arroyo Penasco (Armstrong, 1967).

During Late Meramec time, uplift began in northern New Mexico, exposing the Mississippian beds to subaerial erosion. At least one major uplift occurred during the Late Mississippian and Early Pennsylvanian which folded and truncated the Arroyo Penasco Formation. This activity is probably related to Ancestral Rockies Time deformation and occurred along the Uncompahgre Lineament and the Pecos Picuris Fault, as part of the development of the Taos Trough (see Fig. 9).

PENNSYLVANIAN DEPOSITS IN THE TAOS TROUGH

Paleohistory

The Taos Trough became a tectonically active cratonic basin by Morrowan Time. Its formation is closely associated with Late Paleozoic Ancestral Rockies uplift. It is bounded to the west by the Pecos-Picuris Fault whose north-south trace can be seen running the length of the Sangre de Cristo Mountains. The fault acted as a steep reverse fault which uplifted the Uncompahgre Highlands to the west. To the north, the Taos Trough is bounded by the Cimarron Arch. The southernmost boundary of the arch lies along the Uncompahgre Lineament and the intersection of the lineament and the Pecos-Picuris Fault may have acted as the focus of maximum subsidence in the Taos Trough (Casey, 1980) (see Fig. 9). The trough becomes shallower in a southeasterly direction, emerging onto the Sierra Grande Arch to the east and onto the Pecos Shelf to the south.

Tectonic activity in the terrain surrounding the trough has affected sedimentation within the basin. The stability, sediment input and water depth are dynamically linked to the structural evolution of the basin. Casey (1979, 1980) exhaustively studied the sediments within the region. He measured numerous sections whose facies enabled him to interpret the depositional systems active in the area. In his study, six time-stratigraphic divisions were made of the Pennsylvanian sediments on the basis of facies changes recognizable within the study

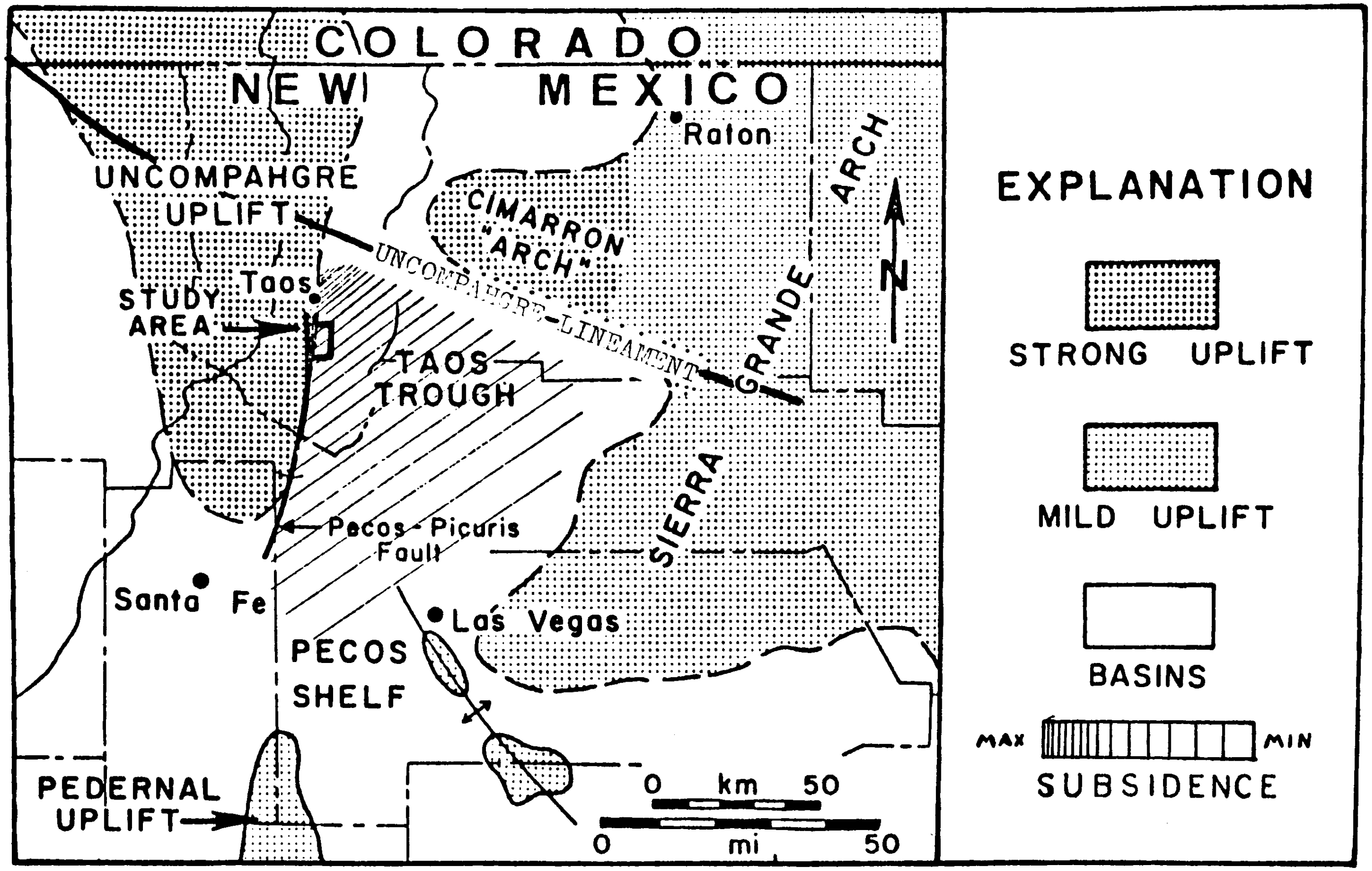


Figure 9. Pennsylvanian tectonic setting of northeastern New Mexico. The diagram shows the relationship between the Taos Trough and the Uncompahgre Uplift, the Pecos-Picuris Fault, the Cimarron Arch and the Sierra Grande Arch. Adapted from Casey (1980).

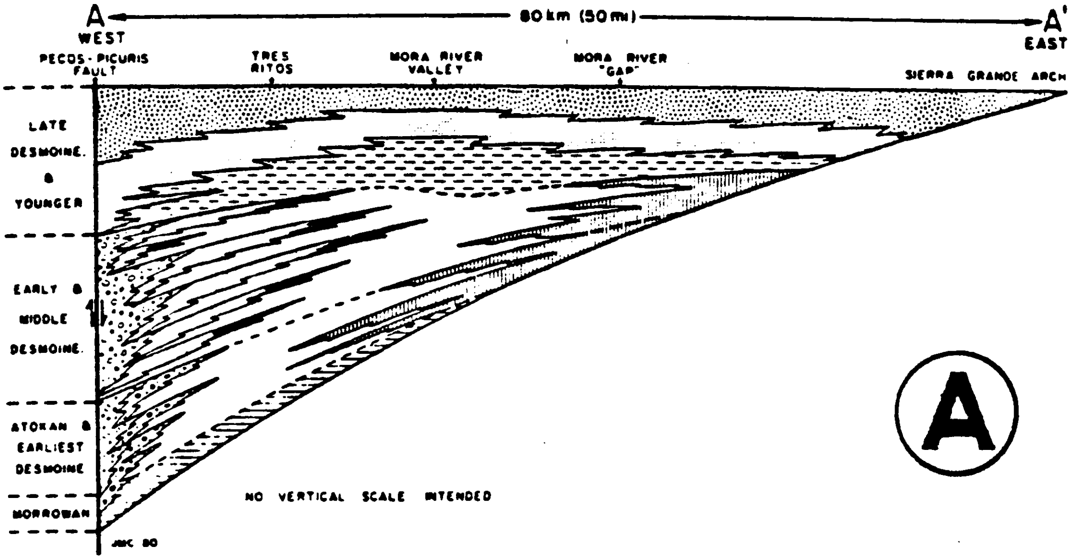
area.

Pennsylvanian Stratigraphy

Several methods of describing Pennsylvanian rocks in New Mexico have been used. The lithostratigraphic method, used for mapping purposes, recognizes large units of similar rock type. The biostratigraphic method subdivides the sediments on the basis of fusulinid zones. These two prominent methods have resulted in a wide range of stratigraphic names, of which only two are widely utilized in the northern Sangre de Cristo Mountains.

There are three thick, generalized lithologic units (see Fig. 10): (1) a lower, predominantly clastic rock unit called the Sandia Formation, (2) a middle, predominantly limestone unit, called the Gray Limestone Member of the Madera Formation, Read and Andrews (1944); Baltz and Bachman (1956), and (3) an upper clastic rock and limestone unit called the Upper Arkosic Member of the Madera Formation.

Sutherland (1963) proposed a new set of lithostratigraphic names tied to fusulinid zones (see Fig. 11). Extensive fossil collection has enabled the time-stratigraphic relations to be worked out in detail. He recognized that the Sandia Formation and the Gray Limestone Member of the Madera Formation are actually the proximal and distal facies of a single depositional system. He divided these two facies into the Flechado Formation in the north - primarily coarse clastics and La Pasada Formation to the south - mainly limestones and shales deposited on a stable platform. He demonstrated that these formations are time equivalent. Both are capped by the Alamitos Formation which



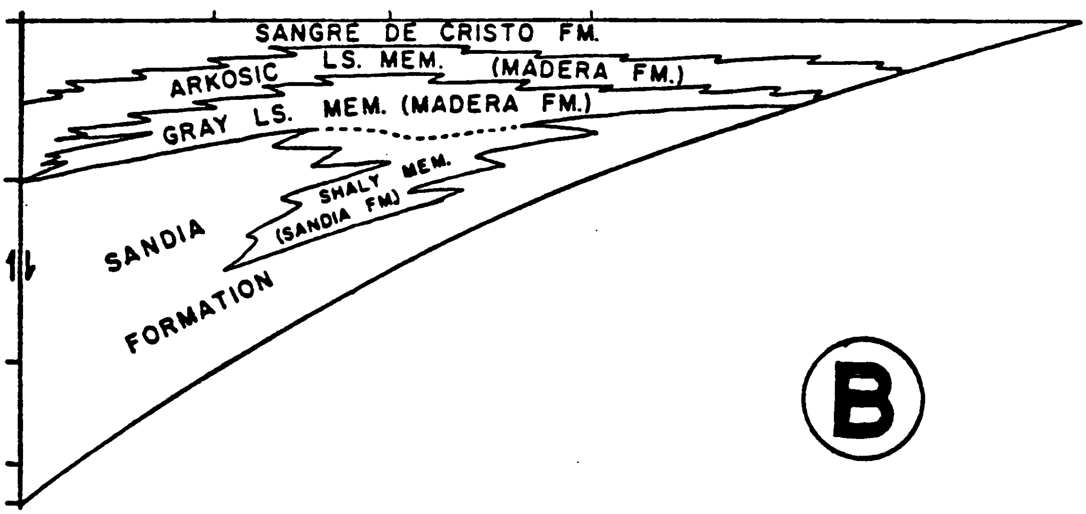


Figure 10. A: Biostratigraphic intervals demonstrate highly varied lithology. Casey (1980).

B: Lithostratigraphic units. Baltz and Bachman (1956).

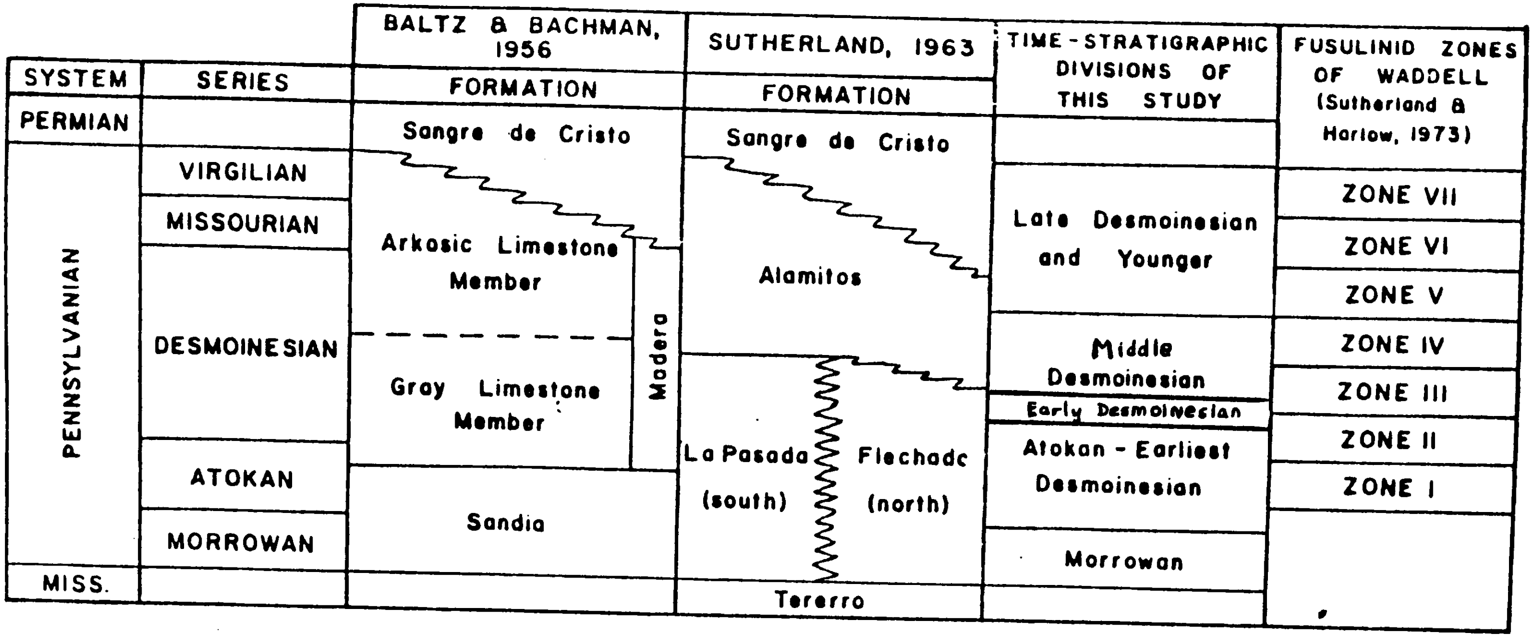


Figure 11. Stratigraphic nomenclature of the Taos-Mora area and informal time-stratigraphic divisions used in this study. Adapted from Casey (1980).

marks a major influx of feldspars into the basin and indicates an abrupt change in the style of sedimentation. The Alamitos, of Middle to Late Desmoinesian age, grades upward into the overlying Sangre de Cristo Formation.

Casey (1980) adopted the biostratigraphic framework established by Sutherland and Harlow in 1973 (see Fig. 11). Their zonation is based on fusulinid and brachiopod fauna which provides the primary basis of correlation. Casey divided the Pennsylvanian into four informal time-stratigraphic units in order to make his divisions consistent with recognizable changes in depositional style which in turn reflect the tectonic evolution of the Taos Trough.

I have modified Casey's subdivisions to correspond better to mappable lithologic changes. The subdivisions are made on the basis of facies changes which reflect the tectonic activity in the area. Consequently, the addition of an Early Desmoinesian division has been added to Casey's subdivisions to facilitate division of the Pennsylvanian sediments into alternating packages of rapidly deposited sediment and shallow marine sediment.

Morrowan through Late Desmoinesian Sediments

Morrowan

Morrowan sediments are found on the west side of Rio Grande del Rancho between Fort Burgwin and Talpa (see map). They are in contact with the Mississippian limestones and form the eastern flanks of Fort Burgwin Ridge. The sediments are approximately 800 ft thick. They were dated by Sutherland (1963), Sutherland and Harlow (1973), on the basis of brachiopod fossils.

The rocks unconformably overlie Mississippian karst limestones (Casey, 1980). The old erosional surface is infilled in my area by a laterally extensive, fine red sandstone which is covered by coarse clastic units. These coarse sandstones are considered to be the first Pennsylvanian sediments. They grade upward into fine sandstones, succeeded by burrowed and rippled coarse siltstones with carbonate flasers. This sequence was deposited by an early Pennsylvanian transgression which first eroded the Arroyo Penasco, and then deposited fluvial sands over the area. Estuarine conditions followed the fluvial episode.

Above the estuarine rocks, nearly 500 ft of fine-grained brown sandstones and siltstones with thin lignite beds were deposited. These consist of coarsening-upward cycles ranging from 3-10 m thick, which coarsen upward from shales and muds to trough cross-bedded sandstones. The sandstones are laterally discontinuous and probably represent stream channels. Near the top of the Morrowan system, coarse sandstone beds are more prevalent. The boundary between the Morrowan and the Atokan appears gradational in the field but it has been mapped at the appearance of the first continuous sheet sand beds (see Plate 1).

The source of the Morrowan rocks appears to be largely the Precambrian Ortega Quartzite and younger sediments because the strata do not contain much feldspar. Sutherland (1963) proposes that the Ortega-Embudo Granite contact had a shallow dip and that Pennsylvanian erosion of the Precambrian rocks exposed increasing amounts of granite. Later

Pennsylvanian deposits are much more arkosic. Furthermore, fragile, embayed quartz grains are found in the lower sandstone unit. The fragility of these clasts implies that the source of the sediments was proximal.

Atokan to Earliest Desmoinesian

Approximately 2,000 ft of Atokan strata are exposed in the northern part of the Taos Trough, forming many of the low-lying wooded hills directly south of Taos. In the study area, they form a low shoulder on the eastward flank of Fort Burgwin Ridge and are exposed on both sides of the Rio Grande del Rancho gorge, just south of Talpa. They are truncated to the north by Rio Grande graben border faults and disappear southward under Early Desmoinesian shales and alluvial cover as a consequence of a regional southerly dip (see Plate 1).

In the Atokan rocks, an abundance of very coarse clastic units records the first major upward movement of the Uncompahgre Highlands along the Pecos-Picuris Fault. The base of the section is mapped at the appearance of a sequence of broad, ribbon-shaped sandstone bodies (see Plate 1). Since these ribbons commonly occur as complex, overlapping layers, the base of the unit cannot be delineated by any one sandstone bed. The boundary is marked by a change in depositional style. Mapping was further complicated because these sands, which acted as a competent member within incompetent shales, have been strongly folded, and many beds are upside down. The most continuous beds at the base of the section show few sedimentary structures; consequently their attitudes can only be extrapolated from the surrounding folds.

The westernmost and lowermost section of the Atokan is characterized by sheet-like sands and conglomerates which contain clasts up to 10 cm across. These are interpreted by Casey (1980) to be distal alluvial fan and braided stream deposits. The ribbon-like shape of the sands suggests that river discharge was characterized by major avulsions rather than by lateral migration of the stream bed. These avulsions were probably caused by sharp changes in the rate of activity along the Pecos-Picuris fault three and one-half miles to the west.

The ribbon sands grade laterally to the east into coarsening-upward fan-delta sequences which are beautifully exposed on the eastern side of Route 3 near Talpa. The typical fan delta cycle evolves upward from dark gray siltstones to interbedded siltstones and thin sandstone beds with occasional fossiliferous limestones. These coarsening-upward cycles are capped by coarse sandstone and conglomerate units with channel-like geometry. Occasionally the channels are replaced by foreset-bedded sandstones, usually less than 4 meters thick.

Capping the delta-front system are braided-stream deposits with typical laterally continuous sands and conglomerates. These become very coarse with mud clasts up to 30 cm long and Pennsylvanian sandstone and limestone clasts up to 12 cm in size. These clasts are an indication that by earliest Desmoinesian time, older Pennsylvanian sediments were being uplifted and reworked. Their size and fragility implies that the source was only a few miles west of Route 3 near the Pecos-Picuris Fault (Casey, personal communication).

In response to activity along the Pecos-Picuris Fault, sharp pulses of coarse conglomerates and very coarse sandstones were deposited by braided streams from the north. Apparently the earliest Desmoinesian was a period of episodic movement along the boundary fault between the Taos Trough and the Cimarron Arch (see Fig. 9).

Early Desmoinesian

Directly above the braided stream sediments of the Atokan lie 240 ft of shales and phylloid algal limestones. They mark a distinct hiatus in coarse sediment deposition in the northern part of the Taos Trough. These rocks were considered by Casey (1980) to be the top of his Atokan-Earliest Desmoinesian unit. The limestones have been traced from the top of the hill containing Fault #19 (see Plate 1) northward across the ridges overlying Rio Chiquito to a quarry next to Highway 64. Farther south, the shales are found at the stream level of Rio Grande del Rancho and apparently they underlie fluvial and fluvial deltaic sequences of the early-middle Desmoinesian section along Rio Pueblo. Because of their thickness, these rocks make a convenient division between the units above and below, and so are mapped as a separate unit.

Early-Middle Desmoinesian

An early to middle Desmoinesian sequence of laterally extensive, coarse sandstone and granule conglomerates, interfingered with fine sandstones and siltstones further up the section, are found in the southern part of the map area (see map). Extensive high elevation weathering and ground cover make a detailed analysis of these rocks impossible. Outcrops consist of linear exposures of jumbled blocks with sigmoidal dish shapes approximately one meter in diameter. When these blocks are found in place, they appear to represent confused sigmoidal cross-bedding and trough cross-bedding similar to the alluvial fan, braided stream facies described by Casey six miles south of U.S. Hill.

The sandstones lie directly on the Precambrian basement and appear to thicken markedly across a normal fault which juxtaposes Precambrian granites with Pennsylvanian sediments. On the east side of the fault, the sandstones are almost paraconformable to Morrowan limestones and shales (see Plate 1 and Fig. 13). Conformably overlying the Middle Desmoinesian rocks are Late Desmoinesian limestones, sandstones and shales. This relationship is evidence of a Middle Desmoinesian age for the alluvial fan deposits.

Early to Middle Desmoinesian sedimentation in the northern part of the field area is envisioned as distal alluvial sheet sands which prograded over an exposed Precambrian basement. The sediments crossed a normal fault which was reactivated during or after Early Desmoinesian sedimentation. Subsequently, the area was submerged and Late Desmoinesian limestones and shales were deposited over the sands.

Late Desmoinesian

The Late Desmoinesian was a period of transgression in the northern part of the Taos Trough. The low-lying delta platform formed during the Middle Desmoinesian was quickly inundated and shallow shelf conditions prevailed during Late Desmoinesian time. Most of this section is missing in the area. However, it is characterized by a thick section of calcareous siltstones and very fine sandstones interbedded with fossiliferous limestones which are generally referred to as the "Gray Limestone Member" of the Madera Formation (Sutherland, 1963) (see Fig. 10).

In the study area, a thick sequence of limestones and shales are found. These overlie the largely sandstone and conglomerate Middle Desmoinesian strata. The section is most continuous south of the Mica Mine and one mile west of U.S. Hill. Exposures are poor and a critical part of the section is obscured in a thickly overgrown canyon. If the sequence is uninterrupted, it consists of several cycles of sandstone and conglomerates which alternate with blue fossiliferous limestone units up to four meters thick. Limestone content increases rapdily up-section and sandstone becomes less abundant. The top of the section is capped by 320 ' of calcareous siltstones and mudstones which surround a low hill and probably cover the valley on top of U.S. Hill.

For mapping purposes, the base of the section is marked at the appearance of the first limestone. This evaluation is supported by the influx of extremely micaceous Ortega clasts in conglomerate beds which overlie the outcrop of Ortega at the mica mine. A similar conglomerate was found cropping out along the roadside at U.S. Hill. This conglomerate divides the predominantly coarse clastic Middle Desmoinesian sandstones from the shale section above it.

MIOCENE AND YOUNGER SEDIMENTS

Geologic History

By the end of Eocene time, the physiographic relief of the region was low and sediments had lapped back onto the worn down Laramide geanticline. In Oligocene time and continuing into Miocene time, large scale volcanic activity began in the San Juan Mountains and spread into the Sangre de Cristo Mountains. Lower Miocene volcanics are exposed northwest of Taos. The Oligocene and Lower Miocene volcanics are generally intermediate in composition.

In the Sangre de Cristo Mountains, volcanic activity was preceded by an orogenic pulse which is recorded as coarse conglomerates at the base of the volcanic deposits. The Carson Conglomerate of Just (1937), Picuris Tuff of Cabot (1938) and Abiquiu Tuff of Smith (1938) are aprons of volcaniclastic rocks with basal conglomerates containing Precambrian and Paleozoic clasts. The geographic distribution of these lower Miocene rocks suggests that they are parts of a sedimentary apron which spread southwards from volcanic highlands in northern New Mexico and southern Colorado.

The size of the basin is not known. The basins of the Rio Grande Depression (Espanola and San Luis Basin) did not exist inasmuch as Abiquiu and Picuris Tuffs are found in fault slices high on either flank of the present basins.

About 20 m.y. ago, the basins began to form in which the Tesuque

Formation of the Santa Fe Group was deposited. These basins were more extensive than the present Rio Grande Rift valley and the Tesuque Formation is tilted and faulted toward the basin.

In the late Miocene and early Pliocene, the grabens of the Rio Grande Depression first began to be defined. At the south end of the San Luis Basin, a sharp anuglar late Pliocene unconformity exists between the Santa Fe Group and the overlying Servilleta Formation. Modern active movement along the Embudo Fault indicates continuing activity in the area.

My field area contains Picuris Tuff. The younger members of the Santa Fe Group, the Tesuque and Ancha Formations are absent, and they apparently exist only to the west of the Taos Fault where they are buried under the Servilleta Formation. Younger sediments in my area are Pleistocene glacial deposits and recent stream deposits (see Plate 3).

Picuris Tuff

The Picuris Tuff was first described by Cabot (1938). Most authors consider the formation to be equivalent to the Abiquiu Formation of Smith (1938). Although these rocks are separated by the Rio Grande Rift, they are thought to be part of a continuous apron which is interrupted by the graben. In the graben the tuffs are overlain by the Tesuque Formation of the Santa Fee Group. The base of the Picuris Tuff is deposited on an erosional surface, or it is found in fault contact with older rocks. The top of the formation may interfinger with the Tesuque alluvial fan deposits (Baldwin, 1956).

In my field area, the base of the formation crops out along New

Mexico Highway #3, just north of Fort Burgwin. Here, coarse conglomerates containing Precambrian and Paleozoic clasts in a tuffaceous matrix are interfingered by yellow, green or white clay layers, and sandy layers. One mile north of U.S. Hill, thin beds of shale and green-white clay beds crop out along the road. Some marl beds and immature, calcite-cemented sandstone beds are found next to Fault #16. These beds are similar to Pennsylvanian sediments except for the presence of clayey and calcite cements.

The Miranda Graben has exposures of pink, volcanic, water-laid tuffs which floor most of the valley. A long spine of volcanic breccia containing angular clasts up to 3 cm in size of rhyolitic material with phenocrysts of hornblende and quartz, has been rotated to nearly vertical and forms low hill crests. Contiguous with the breccia are banded, water-laid tuffs six feet thick. These pink volcanic rocks probably represent the upper portion of the Picuris Formation, whereas the coarse conglomerates represent the base, Montgomery (1953) concluded that the maximum thickness of the formation represented in the area is from 1,250 ft - 1,750 ft.

Glacial Deposits

Moraine and Talus Slope Features

Many cirques are recognized in this part of the Sangre de Cristo Range. They lie above 10,700 ft where well-defined moraines and glacial till are preserved in the upper part of most valleys. Morainal forms, however, are poorly preserved because they lie below the timber line and detailed mapping is not possible. On the basis of good exposures in the upper part of Rio Nambe, there appears to be evidence for only two substages of glaciation plus a protalus rampart composed of coarse blocks located well in front of the modern talus on the cliffs of the cirques (Miller, 1963). It seems likely that they are of Wisconsin age.

In the south-central part of the field area, at the intersection of cross-sections K-K' and I-I', is a formation marked Qp€o on the map. These rocks comprise a steep-sided ridge which trends north-south and are found underlying two other hills which flank Vallecitos. The lack of good outcrop on such steep-sided hills is surprising considering that the Ortega is an extremely resistant rock and less resistant Pennsylvanian sandstones are found in place. The Ortega rocks appear to be large oblong blocks 3 ft by 2 ft by 2 ft on a side. If they are not in place, then the geometry and size of the exposures suggests a protalus rampart or a moraine. Certainly, the blocky nature of these outcrops suggest a geomorphological explanation for their origin.

Glacial Outwash Benches

Remnants of terraces of glacial origin are noted throughout the region wherever the timber cover is thin or absent. Several terraces are present in the field area (mapped as Qg). They overlie all types of rock and form flat caps on the top of hills. The terrace gravels lie loosely on the surface; individual rocks are well rounded and range from 3 in to 10 in in diameter. Ortega clasts with good cross-bedding indicate that the source area lies outside my field area in the Picuris Mountains. The fact that Fault #21a cuts three of these benches indicates that movement along this fault is Holocene.

Quaternary Alluvium

Alluvial deposits in the field area were not studied. At least one terrace is evident in the Rio Grande del Rancho valley but it has not been differentiated. All alluvial rocks which are being deposited by ongoing processes have been mapped as Qal.

PENNSYLVANIAN STRUCTURE

Geological Setting

The field area and its immediate surroundings occupy the northwestern corner of the Taos Trough. The trough is formed by the uplift of the Uncompahgre Highlands to the west of the Pecos-Picuris Fault and downdrop of the area immediately to the east (see Fig. 9). There are 16,000-20,000 ft of Pennsylvanian and lowest Permian rocks preserved in the Sangre de Cristo Mountains (Baltz, 1965), indicating that the minimum throw of the fault was at least 16,000 ft. Because of the strong overprint of Laramide and younger deformations, very little is known of the deformational style associated with the Paleozoic Ancestral Rockies. However, it is generally believed that the deformation occurred along large-scale normal faults which uplifted huge horsts. Deformed zones adjacent to these horsts are generally very narrow (Harms, 1964; McKee, 1975).

Because of its location two miles east of the Pecos-Picuris Fault, my field area has unusually good exposures of the Pennsylvanian-basement contact. Laramide uplift has tilted the basement-sediment contact to 80°E in the northern and central parts of the field area (see cross-sections A-L inclusive) and to 45°S in the southern part of the field area (see cross-sections I-J). Therefore, it is necessary to rotate these contacts to horizontal as in Figure 12 to compensate for the effects of Laramide faulting and to show the style of faulting during

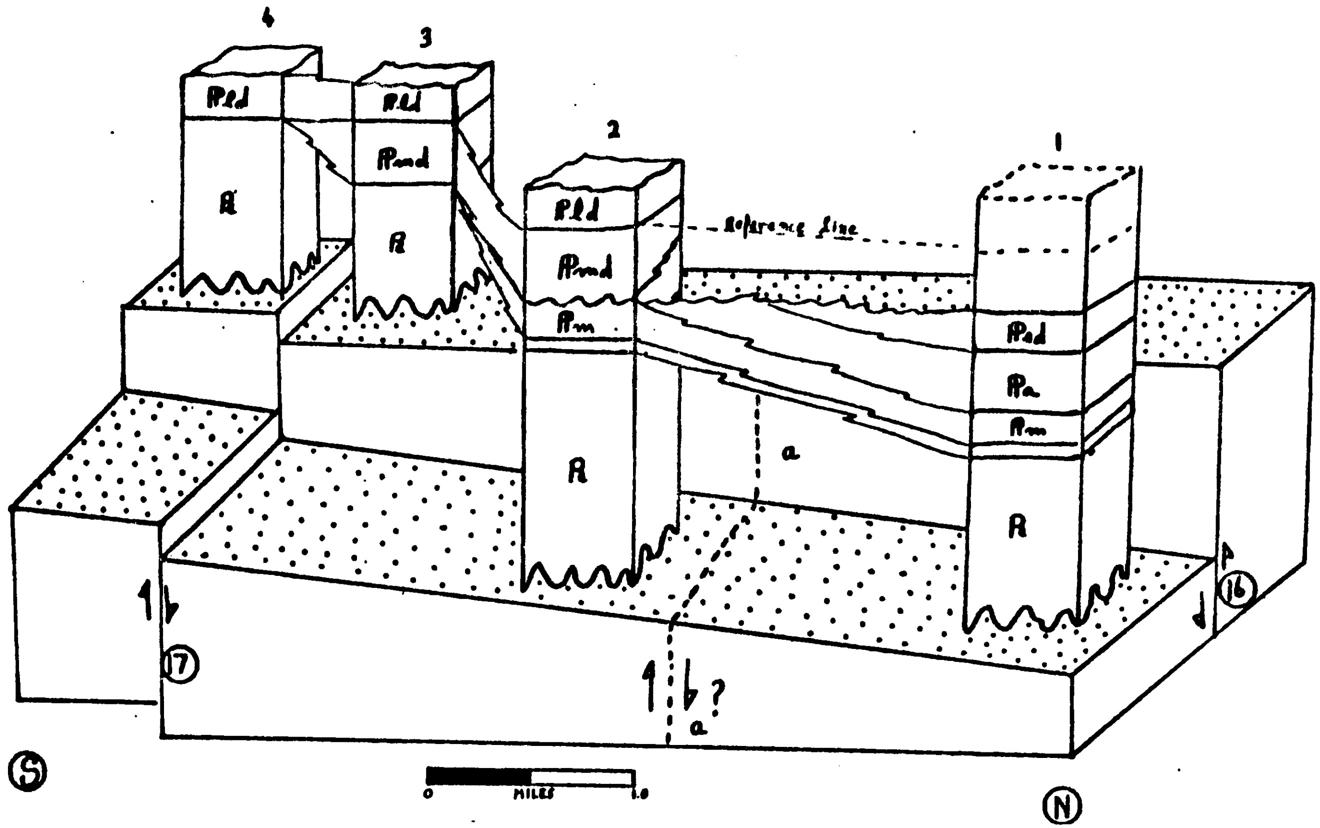


Figure 12. Schematic diagram showing hypothetical position of Pennsylvanian sediments during Late Desmoinesian. Beds have been rotated to horizontal. Stippled blocks emphasize movements of Faults #16 and #17. Columns show varying sedimentary thicknesses. Note unconformity between columns #1 & #2.

the Pennsylvanian.

The basal Pennsylvanian sedimentary sequence in the northern part of the field area (see map and cross-sections) consists of Morrowan shales capped sequentially by Atokan sands and earliest Desmoinesian shales. These latter shales are exposed in the mountaintops to the east and south of my field area and south they are covered by Middle Desmoinesian sands and Latest Desmoinesian shales and limestones. In the areas directly east of my field area, the rocks are flat-lying and, with the exception of some normal faulting, they are undisturbed. Workers in the area feel that a complete section is preserved.

However, as a result of Pennsylvanian tectonics, the Pennsylvanian sedimentary sequence is incomplete near the Pecos-Picuris border fault. For example, in the south of my field area, Middle and Late Desmoinesian rocks were deposited directly on the basement. Figure 12 takes the base of the Late Desmoinesian (the most recent Paleozoic rocks in my field area) as a reference plane. The varying thicknesses of section from this plane to the basement shows the basement topography during the Pennsylvanian.

Discussion of Fault #16*

The sedimentary record indicates that during pulses of uplift along the Pecos-Picuris Fault, Pennsylvanian rocks were exposed and eroded. The Atokan fluvial sands in the north, for example, contain

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*Fault numbers refer to numbering system on Plate 1.

Pennsylvanian cobbles and rip-up clasts. It is my hypothesis that during the Morrowan through Early Desmoinesian the area to the east of Fault 16 (see Plate 1 and Fig. 13a) and perhaps the entire area contained a complete sedimentary sequence. Just prior to the deposition of the Middle Desmoinesian rocks a major uplift occurred along the Pecos-Picuris Fault. This movement resulted in the formation of a set of faults which paralleled the Pecos-Picuris Fault and lowered a series of blocks into the basin (see Fig. 13). Fault #16 is represented as a normal fault which dropped the eastern block into the basin as seen in Figures 12 and 13. The throw along Fault #16 increases to the north creating a ramp. When the Pennsylvanian sediments were exposed and eroded during the Middle Desmoinesian, a northward-thickening wedge was preserved which is represented by an unconformity between column 1 and column 2 in Figure 12. Sediments to the west of the ramp were entirely stripped from the basement (see Fig. 13b).

In support of my theory, Casey (1979) has shown that a huge alluvial fan delta was formed ten miles south of my area during the Middle Desmoinesian in response to a sharp pulse of uplift at that time. The delta sheet sands radiated south, east and west from the fault scarp and inundated the southern portion of my field area. Since the Middle Desmoinesian sands thicken greatly across Fault #16, the fault was active during sedimentation. If one compares cross-section I with J, one can see the elements of this transition. In I the Middle Desmoinesian rocks are deposited on granite and the entire sedimentary package has a uniform thickness (see Plate 1). In cross-section J these

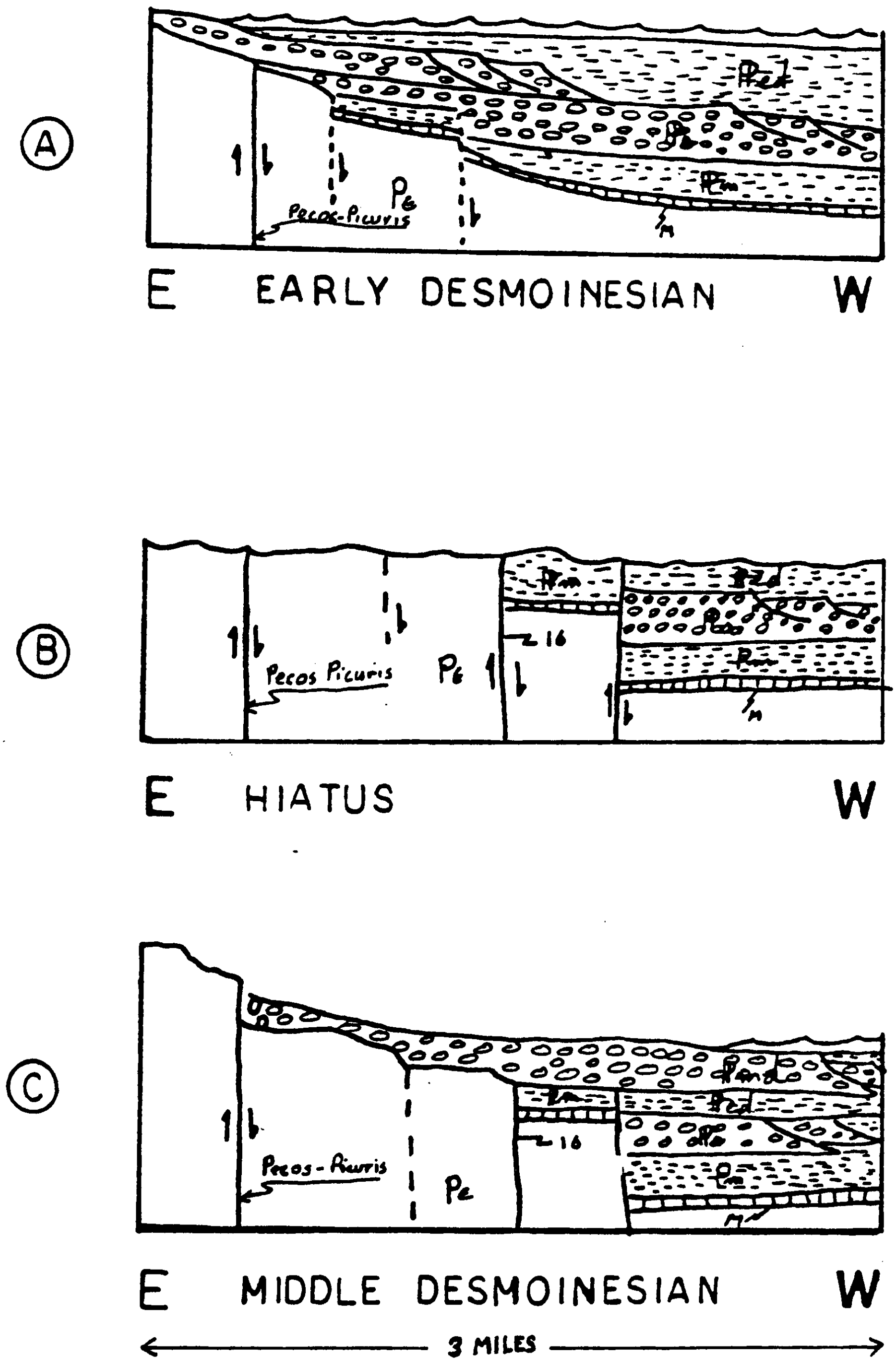


Figure 13. Time sequential block diagrams show how faulting and erosion influence the sedimentary cycles preserved on either side of Fault #16.

sediments are deposited on shales which overlie the Mississippian rocks, and they appear to thicken away from the fault.

This interpretation is not without problems: Problem 1) The sediments near Fault #16 which directly overlie the Mississippian rocks, mapped as Morrowan, consist of shales and muddy limestones with abundant fossil debris. Fossiliferous limestones exist in the northern exposures of the Morrowan; however, these limestones are not abundant. I determined the stratigraphic position of these sediments by the deduction that the Mississippian rocks which they cover were not exposed during Pennsylvanian time (across the fault they have been eroded away) and, since the other exposures of Mississippian are covered by Morrowan shales, then these sediments are most likely Morrowan. As a corollary to this idea, the shales near Fault #14 are probably the same age. Problem 2) The block diagram in Figure 12 could be totally remodified by drawing a fault separating column 1 from column 2 at (a) (dotted line in Fig. 12). This fault might then parallel Fault #17 and operate in a similar manner.

Though there is the possibility that either error exists, (1) the sediments in column 2, Figure 12 are not Morrowan, (2) there is an east-west fault at (a)), I believe that the integrity of diagram 13 has not been violated. However, the tie-lines in Figure 12 would have to be redrawn either as a sedimentary onlap (not changing the structure) or by introducing a fault. I do not favor a major 1,500 ft east-west fault perpendicular to the Pecos-Picuris Fault because there is no direct evidence of its presence, and because most of the uplift and subsidence occurs along the north-south Pecos-Picuris Fault.

Discussion of Fault #17 and Fold I*

Fault #17 separates the Precambrian rocks which contain the mica mine (see discussion of Precambrian rocks) from the Late Desmoinesian sediments. It is a normal fault which strikes east-west and has the north side down. Drag along the normal fault has formed the shallow syncline (I) which parallels the fault on the north side (see cross-section I and Plate 1).

The Late Desmoinesian sediments exposed on the north side of the fault are probably younger than those on the south side. I reach this conclusion because micaceous clasts from the exposed Ortega mica schist are not present on the north side of the fault. On the south side of the uplifted and exposed Precambrian block, the outcrops of Late Desmoinesian sediments appear to have been deposited on the basement and contain clasts of Precambrian schist in the basal sandstone. Upsection, mica is absent. However, a 10 ft zone of muscovite-bearing sandstones and shales are exposed along the roadcut at U.S. Hill. This zone marks the transition from arkosic Middle Desmoinesian sandstone to Late Desmoinesian shale. Since the area to the north of the fault is covered by over 1,000 ft of Middle Desmoinesian sandstone, I assume that there is no northerly source for the Precambrian clasts and that the block to the south was an uplifted, exposed horst at the end of the

*Folds are identified on Plate 1 by upper case letters.

Middle Desmoinesian (see Fig. 14a).

I envision a mechanism where the Middle Desmoinesian horst is uplifted over 1,000 ft and tilted to the southeast. The Middle Desmoinesian sediments were eroded from the horst and subsequent inundation by Late Desmoinesian seas created a gradually encroaching marine sequence of shale and limestone which contained clasts of mica (see Fig. 14b). Eventually, the horst was covered over and no more mica was included in the sediment. This theory explains why there is 40 ft of mica-bearing sediment one mile to the east of the horst, why contiguous with the horst there are only inclusions at the very base of the sediment, and why there are no inclusions across the fault.

The Late Desmoinesian was supposed to be a time of stable shelf conditions and no large-scale vertical movements. Therefore, the fault activity which folded the syncline, fold I, probably occurred after Late Desmoinesian sedimentation (see Fig. 14c). The most likely time for such activity would be the latest Desmoinesian, a time of prograding fluvial deltaic systems which presumably occurred in response to uplift to the west (Casey, 1979).

A second hypothesis for the timing and nature of Fault #17 might also explain the evidence. If, after the lowest Late Desmoinesian sediments were deposited, the fault became active and dropped the north side of the fault, then these sediments would contain no mica. The sediments would be preserved due to the faulting while the block to the south was being eroded. Subsequent deposition of mica-bearing Late Desmoinesian sediment is now preserved on the south side as a conse-

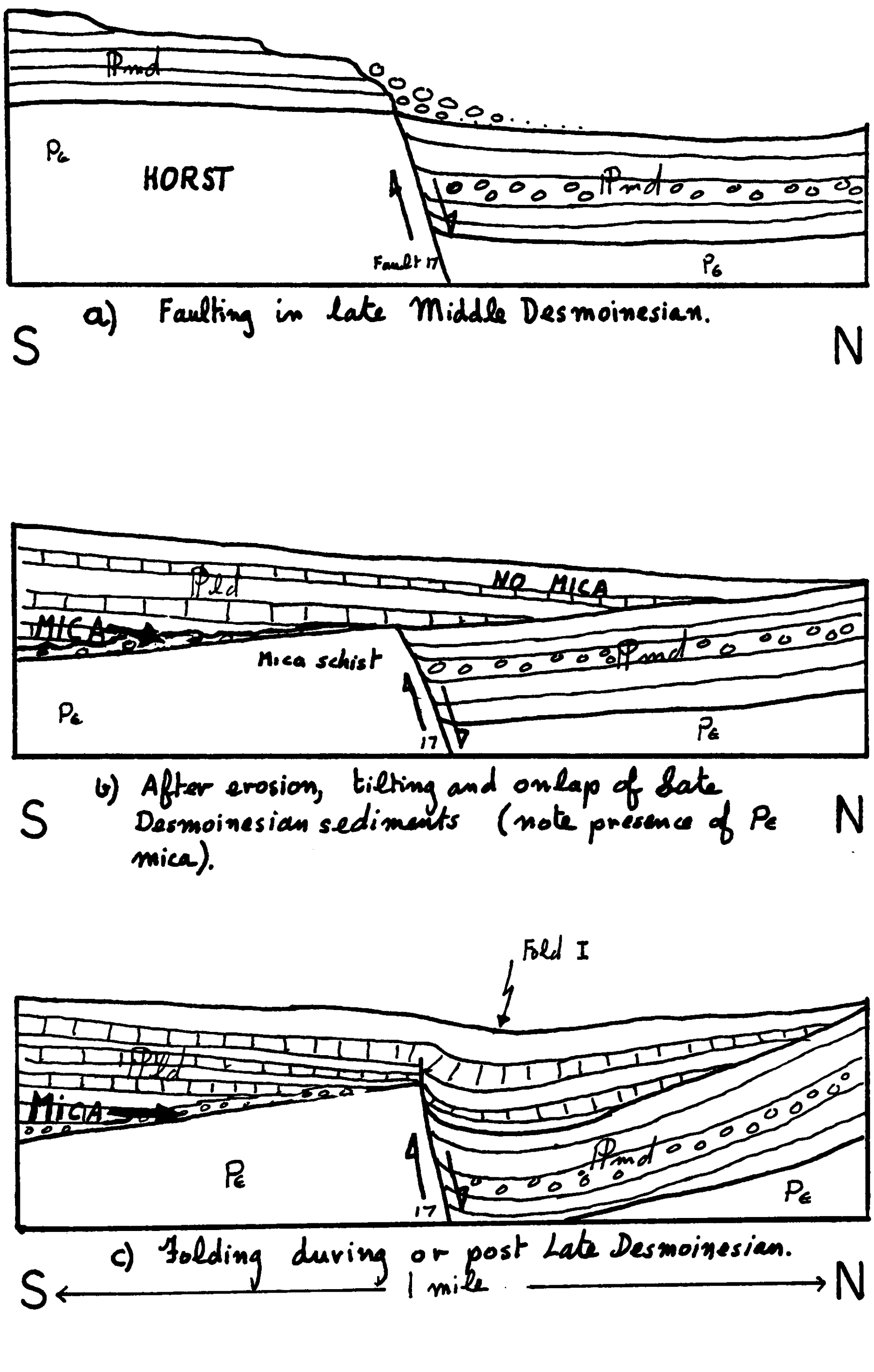


Figure 14 . Time sequential block diagrams demonstrate activity along Fault #17 and Fold I west of U.S. Hill.

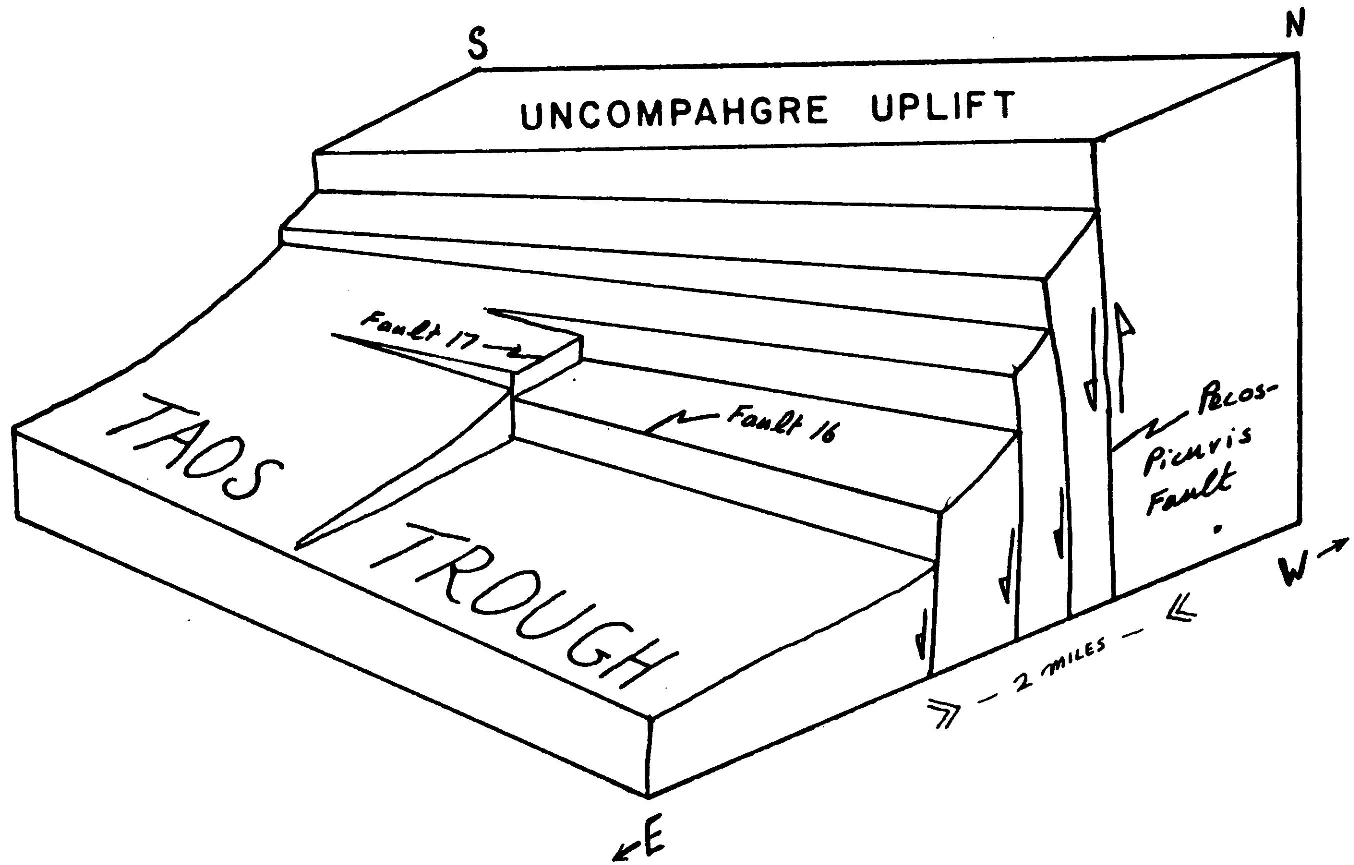


Figure 15. Block diagram shows how uplift along the Pecos-Picuris Fault results in fault slivers such as Fault #16 which produce ramps which are subsiding into the Taos Trough. Fault #17 lies at right angles to the master fault.

quence of present-day erosion of the surface which was tilted during the Laramide, exposing older rocks in the north.

The main weakness of this idea is that the Late Desmoinesian was supposed to be a time of gentle subsidence and marine high-stand. The throw of the fault is approximately 1,500 ft and the exposure and erosion of a horst is not suggestive of quiet shelf conditions.

It is my belief that Fault #17 occurred in response to stresses caused by the renewed uplift of the Uncompahgre Highlands to the west. As I have already stated, Fault #16 and other (hypothetical) faults to the west that are parallel to the Pecos-Picuris Fault may have acted as ramps which accommodated the subsidence into the basin (see Figs. 12 and 15). Maximum subsidence in the trough was in the northwest. Fault #17, therefore, may act as a crosscutting fault which dissects a long, north-south block, allowing the north end of the ramp to subside relative to the southern end as in Figure 15.

To summarize, it is my interpretation that during the Middle Desmoinesian Fault #17 became active in response to uplifts in the west. The load of the ramp to the north was relieved by creating an independent block to the north, allowing the area to the south to uplift and erode (see Fig. 14a). The sediments on the north side of the block were preserved. A gentle southerly slope to the region enabled onlapping Late Desmoinesian sediments to cover the horst and leave no mica clasts to the north of the fault (see Fig. 14b). Renewed activity on the fault in Latest Desmoinesian folded the Late Desmoinesian sediments (see Fig. 14c).

Summary of Pennsylvanian Tectonics

The Pennsylvanian style of deformation is characterized by a series of normal faults which parallel the Pecos-Picuris Fault (see Fig. 15). Fault #16 is probably a north-south normal fault and it was active during the Middle Desmoinesian causing a thickening of the sediments across the fault (see Fig. 13c and cross-section J). Fault #17 is an east-west fault and was active in the late Middle Desmoinesian and later it was reactivated to form fold I. Fault #17 might have occurred to relieve stresses caused by the greater subsidence rate of the northern part of the area (see Fig. 9).

LARAMIDE STRUCTURE

Timing of the Laramide Orogeny

The term "Laramide" is used to describe the tectonic period from Late Cretaceous to Late Eocene. The Laramide Orogeny commenced at different times in different parts of the Southern Rocky Mountain province. This fact probably reflects the advance of an irregular mountain front rather than discontinuous orogenic pulses, because the sedimentary history of Laramide rocks is both continuous and conformable (Tweto, 1975). During the span from Late Cretaceous to Late Eocene, at least 25 my, my area was affected by two of the phases he recognized: greater than 70 my ago and Paleocene (see Fig. 16).

Orogeny in the Sangre de Cristo Mountains

The Sangre de Cristos are bounded to the west by mid-Tertiary normal faults that separate them from the deeply subsided San Luis and Rio Grande grabens. The range can be divided into two structural regions. In the north, the mountains are essentially the continuation of the Mosquito Range and are part of the east flank of the Sawatch Anticline. To the south, they form the eastern flank of the rejuvenated Uncompahgre Highland.

In the south, Precambrian rocks of the western part of the range are separated from the Pennsylvanian rocks of the eastern part by the Pecos-Picuris Fault (Miller et al., 1963); a Precambrian geofracture. Large dip-slip movements occurred both in the Pennsylvanian and

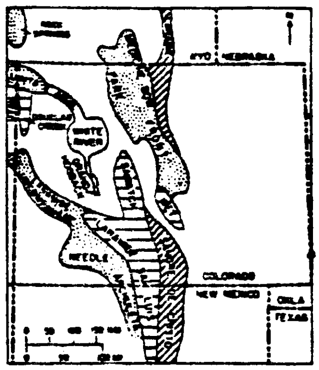


Figure 16. Major uplifts in Southern Rocky Mountain province classified according to time of first rise. Horizontal rules: Cretaceous, over 70 m.y.; coarse dots: Cretaceous, 70 to 65 m.y.; diagonal rules, Paleocene; fine dots: Eocene. From Tweto (1975).

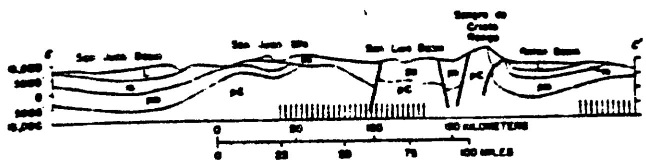


Figure 17. Diagrammatic cross-section across San Juan Basin and Sangre de Cristo Range. From Prucha et al (1964).

Late Cretaceous times; the fault became the eastern margin of a San Luis (Uncompahgre) Highland on each occasion.

The main body of the Sangre de Cristo Range is characterized by a system of Laramide thrust faults and folds in a sequence of Paleozoic rocks, and by a younger block fault system that has elevated the Precambrian rocks to great topographic relief (see Fig. 17). Laramide thrusting and folding did not extend to the flank of the San Luis Highlands until Paleocene time.

The main Laramide structural elements in the Sangre de Cristo Mountains are the vertically uplifted Precambrian block to the west of the Pecos-Picuris Fault, splay faults which dip west towards the Pecos-Picuris Fault, and synclines which parallel the east side of the fault. The Jicarilla Fault is the largest splay fault in the area (Sutherland, 1966). It juxtaposes Precambrian rocks against Pennyslvanian sediments in the Truchas Peaks area. The fault dies out northward and may be expressed as a north-south trending anticline at Rio Pueblo near Tres Ritos (see Figs. 18 and 19). The trend of the fold points directly into my field area.

The Jicarilla Fault is a reverse fault which dips westward 65° to 70° beneath the Precambrian. Near the fault, Pennsylvanian sediments are turned up steeply and may be overturned (see Fig. 20), but they revert to horizontal strata within one quarter mile of the fault. Similarly, the Holy Ghost Syncline, which parallels the Pecos-Picuris Fault, is probably caused by drag along the fault.

A minimum of 1,500 ft of vertical movement is recorded on the

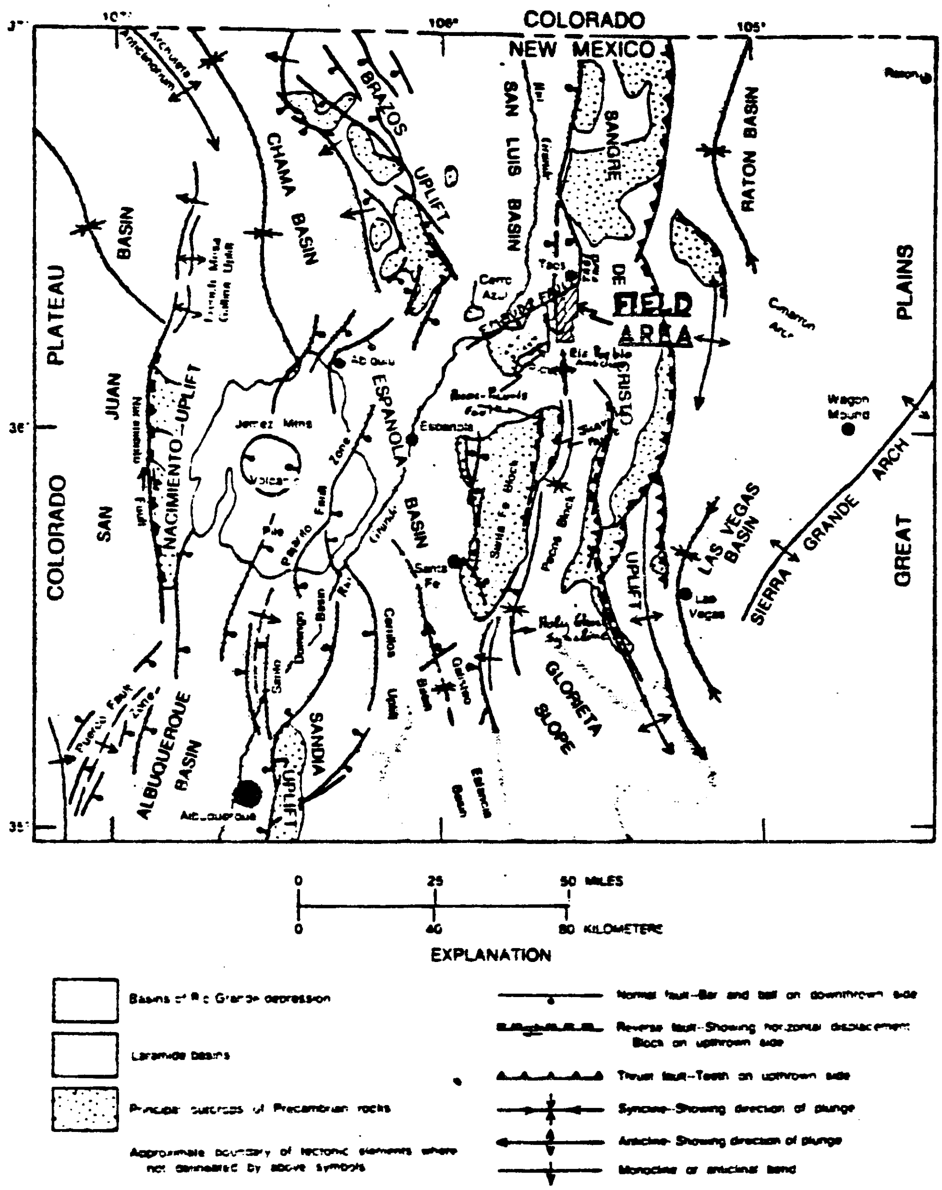


Figure 18. Major tectonic elements of north-central New Mexico showing the position of the field area. Adapted from Baltz (1978).

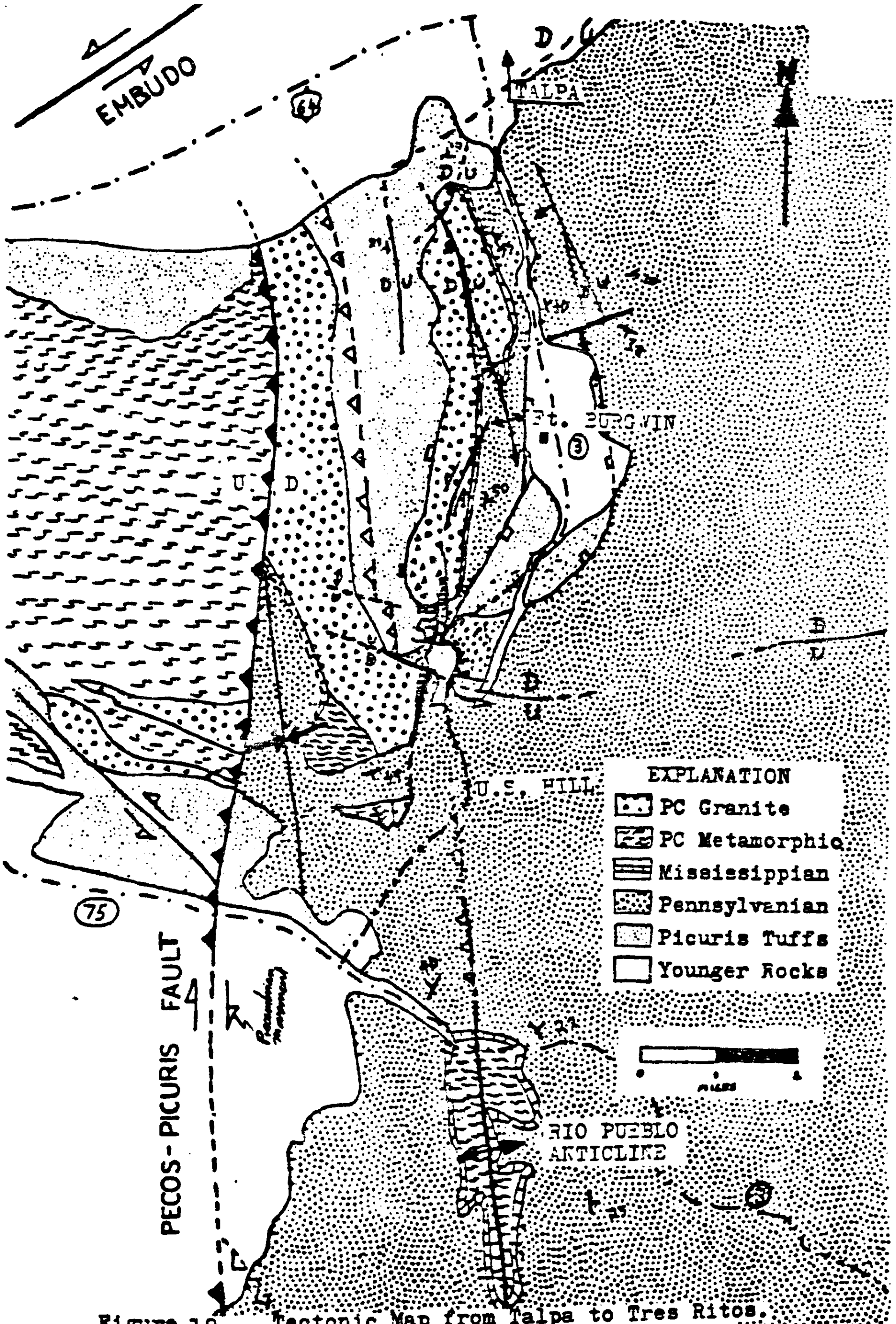


Figure 19. Tectonic Map from Talpa to Tres Ritos.
For structural symbols see Plate 1.

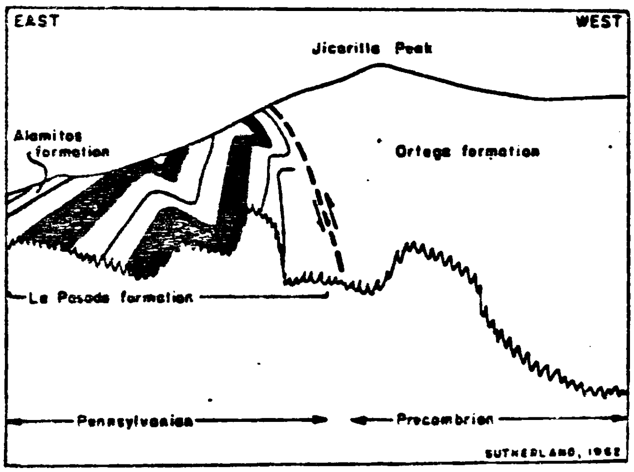


Figure 20. South view of Jicarilla Fault, east of the Pecos-Picuris Fault at latitude 36°(see fig. 18). From Sutherland (1963).

Pecos-Picuris Fault. However, the movement on the Jicarilla Fault may indicate the true stress field which indicates transport to the east (Woodward, 1979). Horizontal shortening during the Laramide was about equal to the vertical uplift. In the southeastern Sangre de Cristo uplift north of Las Vegas, the total horizontal shortening across several zones of thrusting is less than 2 miles and the vertical structural relief relative to the Las Vegas basin is about the same (Baltz, 1972, 1978).

My field area is located 2 mi east of the Pecos-Picuris Fault which acts as the master fault in the region. My area contains numerous faults which parallel the master fault, and the basement-Mississippian contact which dips steeply to the east. Further to the east, the Pennsylvanian sediments are deformed into recumbent folds which abruptly die out into flat-lying, undeformed strata. It appears that the folding and tilting is a reaction to subsidiary faults which are similar to the Jicarilla Fault (see Fig. 20). The deformation in my area does not involve any apparent cross-cutting dislocations along the sediment-basement contact. Apparently such faults are located further to the west, and we are looking at the lower levels of the uplift.

The style of deformation in the Sangre de Cristo Mountains and in the study area is very different from that of the rest of the Rocky Mountain Front Range. The Front Range is characterized by blocky uplifts which forced very large, simple drape folds in thick competent units which, in many cases, overlie an incompetent Cambrian shale. Exposures in Colorado and Wyoming are concentrated at higher structural levels of the uplifts and generally only the Block I, Block II and Block III parts of the fold (Stearns, 1978, p. 11) are exposed for study (see Fig. 21). What appears at lower levels, Block IV and Block V, and particularly how the space problem below Block IV is resolved is largely a matter of conjecture.

Because the sediments in the Sangre de Cristo Mountain block were continually stripped from the basement during the Laramide Orogeny (Tweto, 1975, Osterwald, 1961) exposures along the Pecos-Picuris Fault involve much lower structural levels (see Fig. 22). Exposures in my field area involve the master fault and a series of subsidiary faults which are exposed at the basement contact, and also the region which would correspond to Stearns' Blocks IV and V. As a consequence of the deeper structural level, the dips of the faults are steep (see the discussion on the controls on Laramide Deformation).

Furthermore, the style of deformation in the region is considerably affected by the lithology. In Colorado and Wyoming, the sediments are dominated by thick, homogeneous limestone and sandstone units which form the "strut" of the drape folds. The Pennsylvanian sediments in the Sangre de Cristo Mountains are predominantly thin-bedded sandstone and shale. The Mississippian sediments are attached to the basement and act as part of it (see Fig. 22). The Pennsylvanian sediments, on the other hand, because of their heterogeneity, act as independent units whose folded wavelength is dictated by parameters such as bedding thickness and the distance of separation between beds. These deform as asymmetrical, overturned, flexural-slip, cylindrical folds whose axes

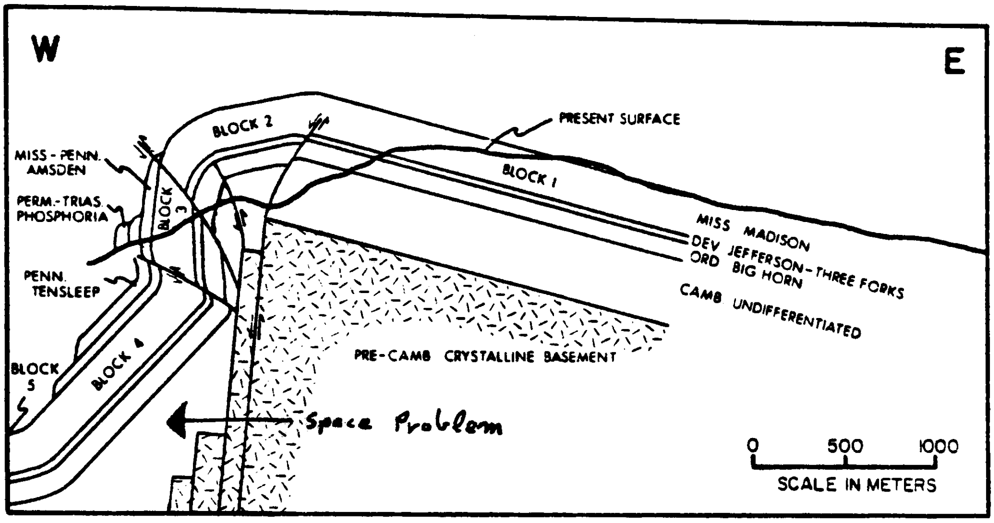


Figure 21. Cross-section of Rattlesnake Mountain west of Cody, Wyoming. The figure demonstrates a typical forced fold and Stearns (1971) block designations. From Stearns (1978).

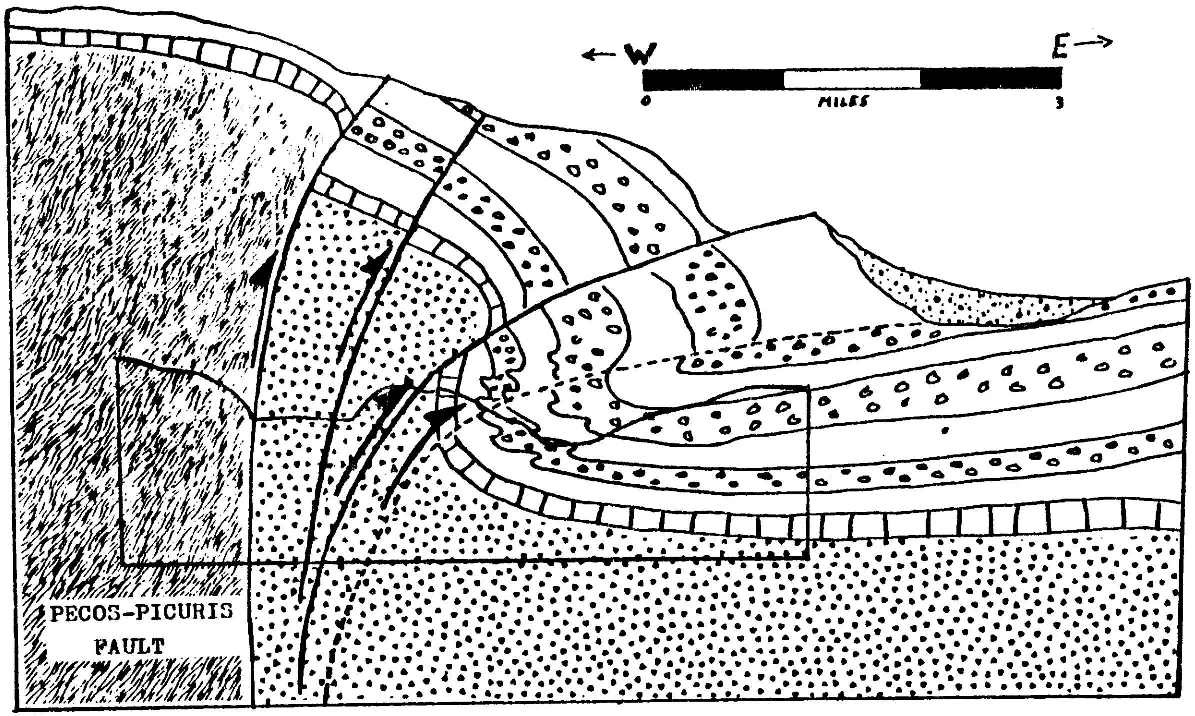


Figure 22. Schematic cross-section demonstrating the style of faulting along the Pecos-Picuris Fault. Note that the splay faults propagate upwards - each fault becoming less steep near the surface. The Pennsylvanian sediments are shown folded into small folds at the base and at the top they act as a drape fold. Insert represents the present erosion surface taken from figure 30.

parallel the Pecos-Picuris Fault.

The lithology has allowed the rocks to conform to the uplift of the basement by interstratal slip and lateral mass transport into the hinges of the fold (see Fig. 22). A high degree of granulation near the basement-Mississippian contact suggests that bedding slip may be a contributing factor which enabled the Mississippian rocks to conform to the shape of the uplift.

In summation, I feel that my field area offers an excellent opportunity to explore the lower structural levels of the Laramide uplifts. In particular, the style of folding in the Pennsylvanian sediments might have application to the study of the deformation of the Cambrian rocks which underlie much of the Front Range.

Controls on the Style of Laramide Deformation

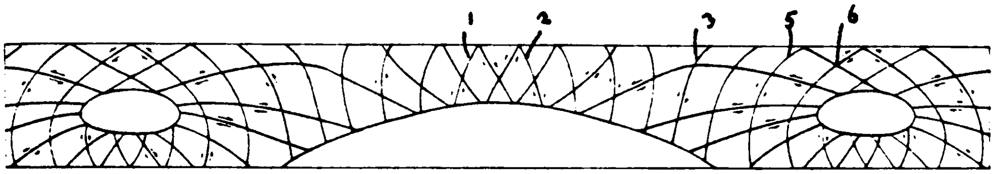
Laramide structures have been studied extensively in Colorado and Wyoming. Several parameters appear to be important in controlling the type of deformation which occurs. The most important parameters are (A) basement control, (B) depth of sedimentary cover and (C) the rock types.

(A) Chamberlin, 1945, Osterwald, 1961, and others have noted that the Rocky Mountain Foreland is characterized by vertical zones of faulting which trend both north-south and east-west. These blocks act as individuals and are controlled in form by the basement. The Precambrian rocks have acted as a rigid basement. Segmentation of the basement has resulted in recurrent vertical movements of basement blocks which have produced folds and faults in the overlying sediments.

The basement uplifts caused narrow zones of deformation which demonstrate simple drape folds, steep normal faults, reverse faults and shallow reverse faults. Are both normal and reverse fault movements consistent with a model that invokes purely vertical movements within the basement? Hafner (1951) solved a boundary value problem which fits the general requirements of basement deformation during the Laramide. Figure 23 represents some of the faults which would be generated when a brittle material is subjected at the depth of the Moho to absolute upward and downward movements. Faults ranging from high angle normal (1 and 2), high angle reverse (3) and low angle reverse (5 and 6) can be produced by a consistent mechanical system.

(B) There is a distinct change in fault character from east to west. In the Rocky Mountain Foreland to the east, deformation is blocklike along nearly vertical faults. To the west, shallow thrust faults predominate with low dips. This phenomenon is not attributed to a change in the stress regime (from vertical to horizontal), rather it is attributed to the increasingly thick sedimentary section as one moves from east to west.

Several models have been proposed to explain the change from high angle to low angle faults. The simplest model is described by Prucha et al. (1964) (see Fig. 24). This model states that the angle of the fault is a function of depth. The initially vertical faults rotate away from the uplift as they reach the surface. Therefore, thrust faults would form only in the Rocky Mountain Geosyncline where the sed-



.. Diagram of potential faults that would be produced in a continuous, elastic block by differential vertical loads at the bottom (after Hafner, 1951). The lower boundary loads are in the form of a smooth sinusoid, the end loads are linear increases in the horizontal component of stress (normal burial), and the upper boundary represents the air-rock interface. Arrows indicate sense of shear on potential faults.

Figure 23 . A spectrum of fault types can be generated by simple vertical movements. Faults #1 and #2 = high angle normal; #3 = high angle reverse; #5 and #6 = low angle reverse.

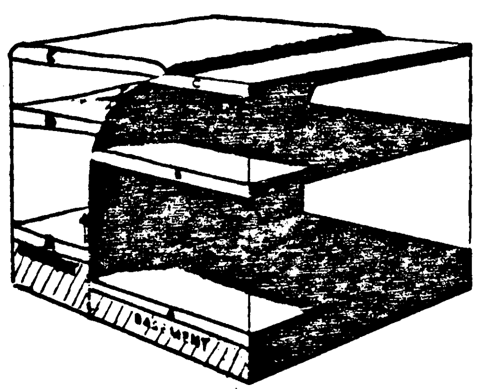


Figure 24 . The diagram shows that the dip of upthrust is a function of depth. For example, the dip at A is greater than the dip at C. From Prucha et al (1964).

iments are very thick. Fault planes in the Front Range, on the other hand would be nearly vertical.

Sanford (1959) presented still another set of geologically feasible conditions which would produce curved faults in a brittle medium (see Fig. 25). These faults were produced in a sand box experiment with a hydraulic piston uplifting the sediments. The faults are high angle at depth and curve to low angle thrusts at the surface. As the piston moves upward, reverse faults form sequentially. As each fault dies out, it is replaced by new reverse faults which begin where previous faults curve away from the vertical. Furthermore, steep normal faults may form behind the thrust faults to relieve the space problem created by the rotation of the fault plane.

Stearns (1975) demonstrated the space problem involved with the uplift in a plexigalss cutout experiment (see Fig. 26). Clearly, to compensate for the space created in Figure 26b a normal fault must form as in Figure 26c. This observation compares well with the sandbox experiment and the structure of the Owl Creek Mountains (see Fig. 27). Caution must be used however because Figure 28 from the same paper indicates that the same space problem can be solved by reverse fault slivers.

In my field area, numerous faults are found parallel the Pecos-Picuris Fault. Their present sense of slip is normal and their position is in between the master fault and the upturned limb of the fold (see Fig. 22). A normal regime, such as that presented by Osterwald (1961) satisfies the strikes and dips within the field area with the

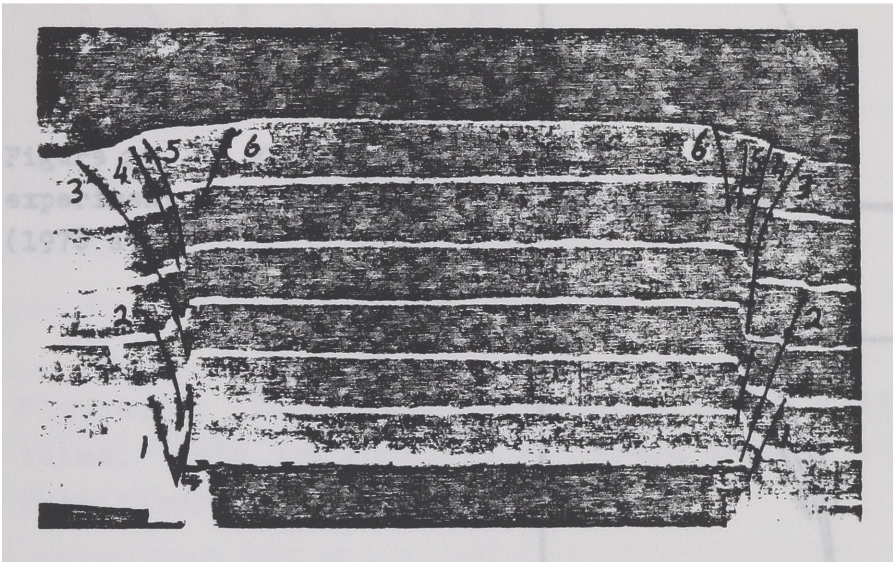
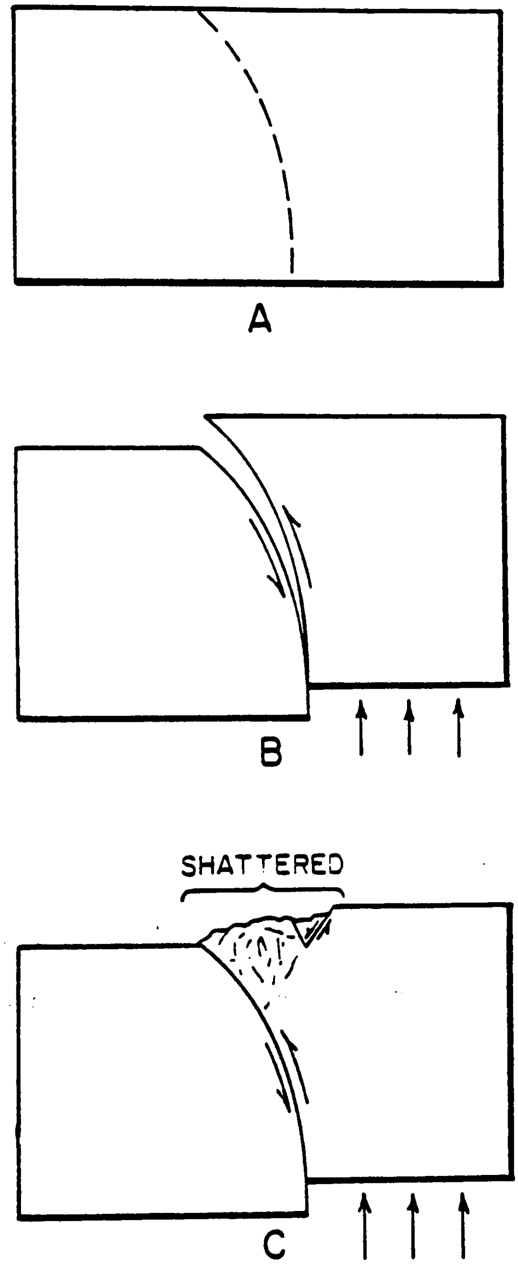


Figure 25 . Photograph of the front view of a sandbox experiment. The lower piston has been displaced about 3 cm. Faults are traced on front: numbers refer to the order of their formation. Therefore faults # 1 & 2 formed before faults # 3, #4 & #5. Fault #6 is a normal fault (see fig. 26). From Stearns (1978 a).

Figure 26 . Plexyglass experiment. From Stearns (1978 a).



— A) Cross section of a glass block with curved cut (dashed line) through it. B) Configuration of the glass block (brittle) if vertical displacement is created along the right hand side. If material is brittle a gap is created. C) If the glass block is confined and gaps are not permitted the lobe of the uplifted block must break off along normal faults and shatter to conform to the left side block.

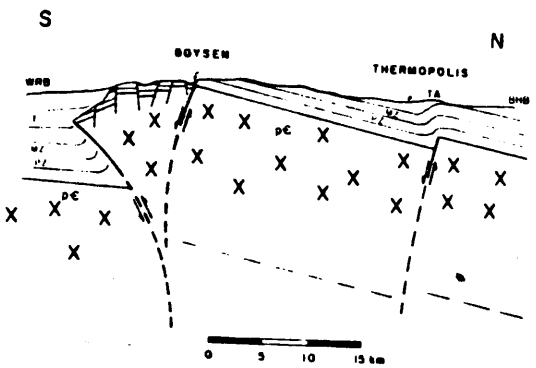


Figure 27 . Scematic cross-section of Owl Creek Mountains. Note the location of Boysen normal fault. From Stearns (1978b).

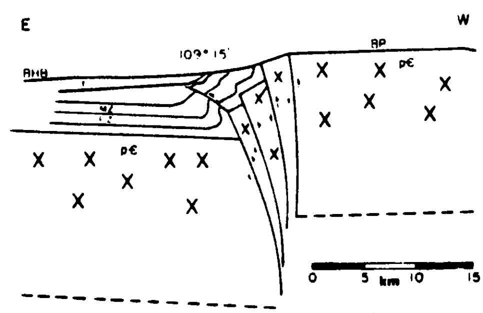


Figure 28. Schematic cross-section of the eastern flank of Beartooth Mountains. From Stearns (1978b).

exception that there is no evidence for the main reverse splay fault. To solve this problem I have located the fault to the west of the upturned sediments (a solution consistent with the data). Therefore, the region to the east of this fault must have been under compression at the time. Since the predominant post-Miocene deformation in the region is rifting and normal faulting, the normal component of these faults can be ascribed to the reactivation of older Laramide reverse faults. Woodward (1979) describes similar phenomena in the south. Therefore, I believe that reverse fault splays such as those described in Figure 28 best solve the space problem in my field area and that the normal sense seen on many faults is a younger motion.

(C) The third parameter to be discussed is the effect of the sediment type. The sedimentary cover in the Sangre de Cristos is considerably thinner than that of the Rocky Mountain Foreland in Colorado and Wyoming. Those areas are covered by thick Paleozoic carbonate sequences which act as a "strut" in Stearns' nomenclature (1971, 1975, 1978). In Colorado the strut is free to assume monoclinal folds without faulting because the incompetent Cambrian shale moved by interstratal slip and mass transport into the hinges of the folds.

Pennsylvanian shale and sandstone in the main part of the Sangre de Cristos deformed on a much smaller wavelength than counterparts did in the Rocky Mountain Front Range. The wavelength of the folds is a function of the thickness and spacing of the more competent members as in the model by Currie et al. (1962). The thicker, more dominantly clastic units such as the Middle Desmoinesian sand probably copied the simple forced-fold morphology, but the thinner interspersed Atokan and Morrowan sand and shale units in my field area have acted with much shorter wavelengths in response to the compression caused by the rotation of the basement-Mississippian contact.

In conclusion, Laramide structures in my field area are controlled by a major north-south geofracture, the Pecos-Picuris Fault. This fault has numerous faults and folds that are parallel to it within the field area. The faults are steeply dipping which is consistent with the depth of present exposure. The type of rocks involved in the area is not conducive to the formation of simple forced folds. Rather, the forms observed are more consistent with interstratal slip and mass transport into the hinge caused by the rotation of the basement-Mississippian contact to nearly vertical.

LARAMIDE STRUCTURES OF FORT BURGWIN RIDGE AND U.S. HILL

Introduction

The ensuing discussion has been divided into two sections because the northern two-thirds of the field area is distinctly different from the southern part (refer to Plate 1 and Fig. 29). Fort Burgwin Ridge, in the north (from line K-K' to line A-A' on Plate 1), has a pronounced north-south alignment of folds, faults and morphological features and involves different Pennsylvanian rocks from those in the south (see discussion of Pennsylvanian structure). In the U.S. Hill region (from line K-K' south), the strata, faults and morphological features strike east-west. I will discuss the Fort Burgwin area first, then the U.S. Hill region and then their interrelation.

Fort Burgwin Ridge

Morphology and Geology

Fort Burgwin Ridge occupies the northern two-thirds of the field area (see location map, Fig. 29). The ridge is the dominant topographic feature in the region and its double crest, which is formed by the Arroyo Penasco Formation (Mississippian), is clearly visible in aerial photographs. The ridge separates a Precambrian terrain to the west from Pennsylvanian rocks to the east.

The block diagram in Figure 30 (see Fig. 29) is a north view of

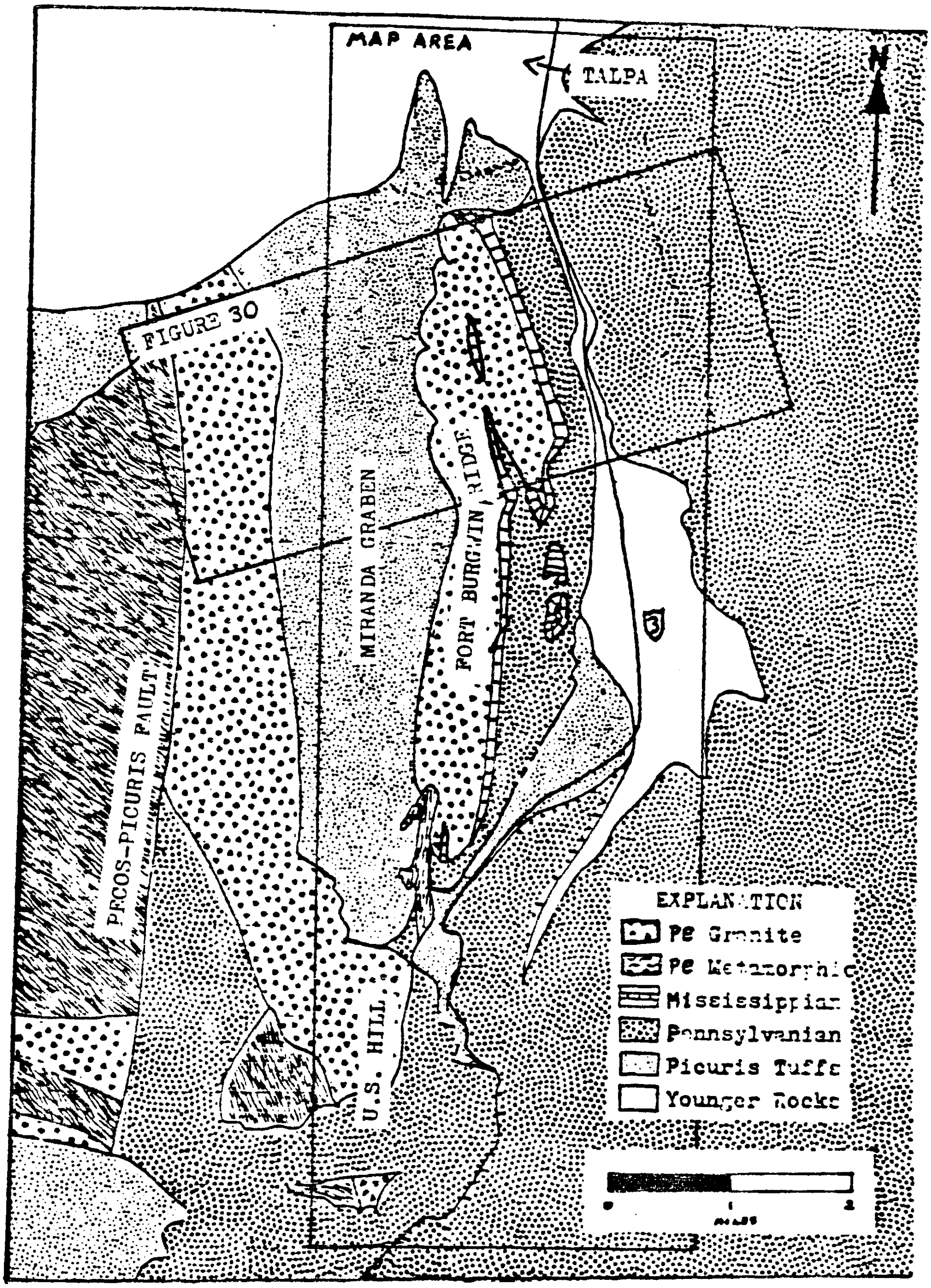


Figure 29. Location Map.

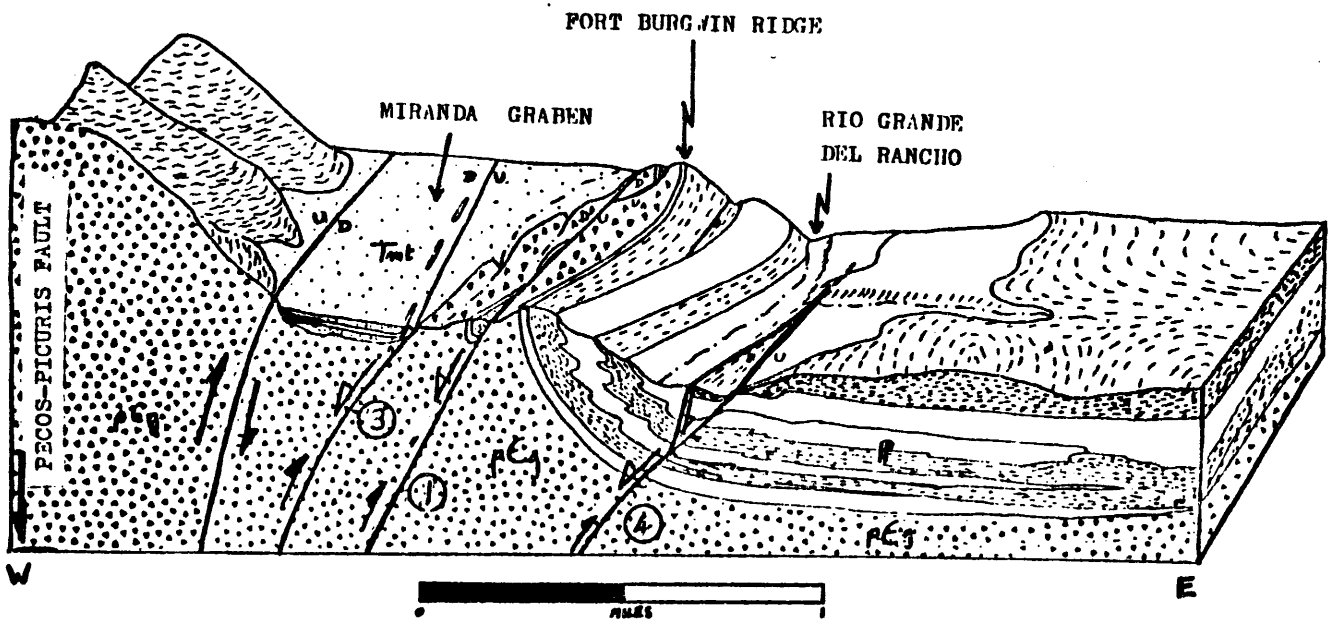


Figure 30. North view of Miranda Graben, Fort Burgwin Ridge, Rio Grande del Rancho valley showing structure. Black arrows are Laramide movements; Clear arrows indicate Miocene and younger movements. pCo = Ortega Metaquartzite; pCg = Granite; M = Mississippian; P = Pennsylvanian; Tmt = Picuris Tuff. Numbers 1, 3, 4 refer to plate 1.

a six by three mile section lying athwart the Pecos-Picuris Fault, across the field area and into the flat-lying sediments of the Taos Trough. It includes the mapped area from cross-section lines E-E' to A-A'.

The Pecos-Picuris Fault forms the westernmost edge of the diagram. Eastward from the fault, the Picuris Mountains stop abruptly and form a straight scarp along the westernmost edge of a post-Miocene graben. I interpret the scarp to be fault-controlled. The valley is filled with Miocene Picuris tuffs. To the east one climbs up the western flank of Fort Burgwin Ridge and crosses Fault #1 which separates the double crests of the ridge. The ridge is held up by Mississippian limestone. The eastern flank of the ridge has exposures of Morrowan and Atokan shale and sandstone, and these sediments are recumbently folded. Continuing east across the Rio Grande del Rancho, the Atokan rocks become flat-lying. They are cut by a small graben (Faults #4 and #5) which lowers a sliver of Middle Desmoinesian shale into the sandstone section. Further east, horizontal strata predominate.

Identification of Laramide Structures

The rocks involved in the Laramide Orogeny are pre-Late Eocene. Consequently, only the Paleozoic and Precambrian rocks within the field area could have been involved in Laramide tectonics. It is difficult to separate Laramide structures from the effects of post-Miocene vertical movements. For example, dips within the Picuris tuffs indicate 20° of tilting to the north between lines H and L on the map and on the northern flank of Fort Burgwin Ridge. Also, near Fault #3, the strata have rolled over steeply to the west. These movements are evidence that post-Miocene structures must be present in the older rocks.

Several features, however, are probably of Laramide origin. These are: (1) the very steeply dipping Mississippian-basement contact, (2) the strongly folded Pennsylvanian strata between the contact and Fault #5, including fold C (but not folds D, E, F or G; see Plate 1), (3) fold B and related deformations and (4) the strong north-south trend of Fort Burgwin Ridge which is paralleled by the canyons in the granite on the western flank of the ridge and by Fault #8. Also, some other features may be related to Laramide tectonics. These are: (1) all faults which trend north-south (most notably Faults #1 and #3 and possibly even Faults #4, #5, #12, and #13) may be old Laramide trends which have been reactivated during younger deformations, and (2) some of the jointing and crushing of the granite may be Laramide, though it has already been noted that much of it is Precambrian in origin.

Folding in the Pennsylvanian Sediments

Even before this investigation it was presumed that the rollover of the Mississippian-basement contact was a Laramide feature. How this movement was accomplished, however, was conjectural. I began my work with the premise that the basement-sediment contact was rotated by faulting. Furthermore, examination of the contact itself and the uniform attitude of the Mississippian rocks indicated that the Mississippian acted essentially as part of the basement. The Pennsylvanian, on the other hand, acted independently. The key to the nature of the Laramide structure of the area was the unravelling of the fold style of the Pennsylvanian rocks between the Mississippian outcrops and Fault #4. The trend of these folds indicated that the local maximum compression had a minimum plunge of 35° to the east (see Fig. 31). This fact indicated that the fault which rotated the Mississippian and folded the Pennsylvanian must have been located to the west and passed above the outcrops exposed on the ridge.

Mapping within the Pennsylvanian strata is made more complicated by two factors. The first factor is the deltaic nature of the sediments. Frequent pinchouts, caused by lateral and vertical changes in sedimentary depositional styles, make bed-walking difficult. Secondly, due to the overturned nature of the folds, the hinges of the folds were preferentially preserved due to their steep dips. This factor resulted in a jumble of dips and common overturned beds. The overturned beds are particularly confusing due to the lack of sedimentary clues to identify the tops of the beds. On the field map some beds, which have been interpreted to be overturned in the cross-sections, do not have overturned symbols because of this uncertainty.

A further effect of the overturned folds is that, with the exception of fold C, it is impossible to place the axial traces of the folds on the map. For example, follow the contact between the Morrowan and Atokan rocks from north to south. The dips in the north are around 55° east. As one crosses cross-section line C, they change to 80° east to overturned 72° west. At line D they are 39° to 21° west but by by cross-section line E they have rolled back to vertical, dip 78^{o} east and quickly become less steep. These dips delineate two folds, but their axial traces would follow the contours of the arroyos because the axial plane is nearly horizontal. Therefore, to simplify the map, these traces were omitted.

To unscramble the geometry of these folds I used the "down structure viewing method" (Mackin, 1950) by stacking simplified cross-section lines on a light table (see Fig. 31). I used a uniform spacing which reflected the north-south separation of my cross-section lines and assumed a uniform dip of 5.5^{o} (see Table 2). The east-west orientation of the cross-sections was controlled by the position of the Mississippian limestone. The result of this method was that the roll-over of the Morrowan-Atokan contact described in the paragraph above was explained by recumbent folding. This pattern fit well with the style of fold C.

Figure 32 is a dramatization of the structure which emphasizes the interaction of the basement with the various folded units in the Pennsylvanian. The Mississippian and granite are rotated to vertical as a unit. The Pennsylvanian sandstones, because of the thick Morrowan shale, are free to act independently and they form compressional folds to compensate for the shortening caused by the curling up of the basement.

The diagram clarifies several confusing aspects of the geology. It is apparent that minor fluctuations of the erosional level cut through by Morrowan shale of dramatically different dip. The geometry

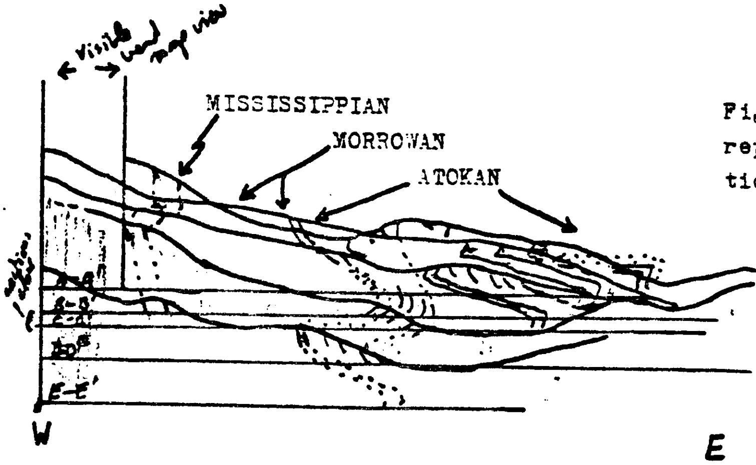
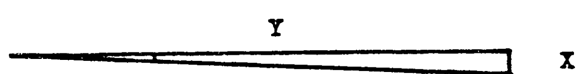


Figure 31. Worksheet representing cross-sections on a light table.

TABLE 2

DISTANCE BETWEEN CROSS-SECTIONS		ELEVATION DIFFERENCE	X/Y = tan DIP ANGLE	
A-B	2,000'	180'	180/2,000	.09*
B-C	1,000'	90'	90/1,200	.075
C-D	2,600'	230'	230/2,600	.088
D-E	2,500'	300'	300/2500	.12
			AVERAGE	5.5°



Measured Angle = 5.5°

* radians

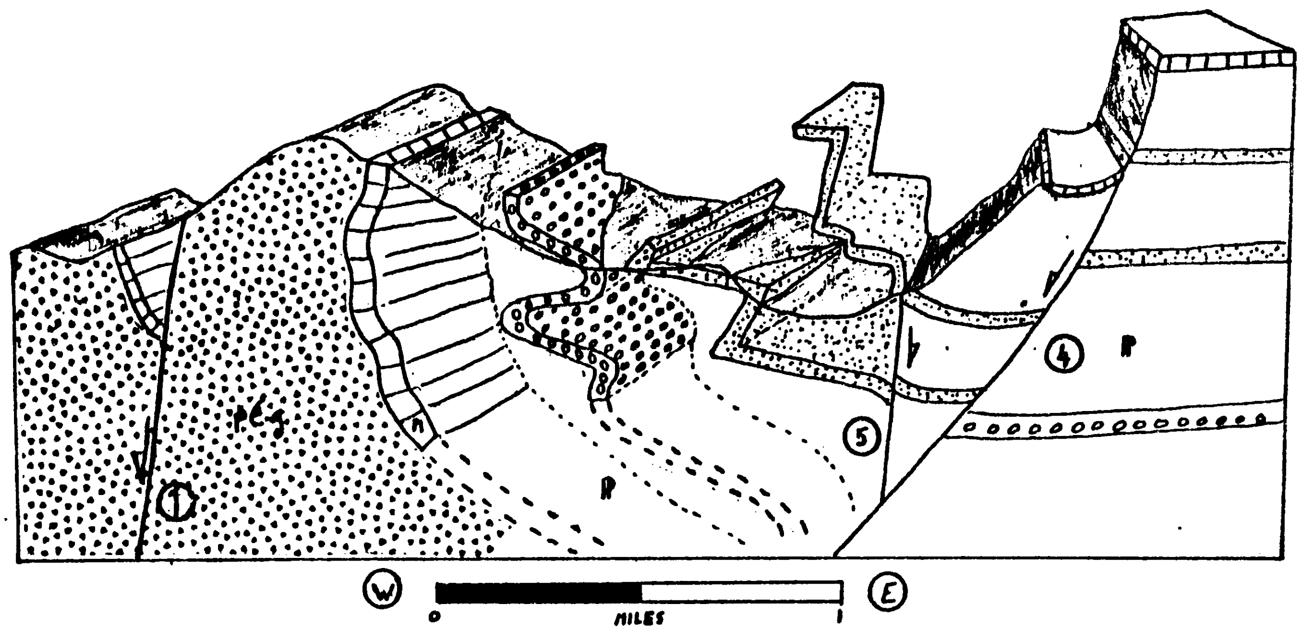


Figure 32. Schematic diagram showing fold style in the Pennsylvanian rocks. Diagram was made from composite of cross-sections A through E.

requires that the Morrowan section is thin under the synclines and thick in the anticlines - a situation typical of decollement folding. Figure 32 is the basis for the interpretation of cross-section lines A through F. Vestiges of the folds in the Pennsylvanian are seen in cross-sections G and H; however, the Laramide structure in these cross-sections has been heavily overprinted by post-Miocene deformation.

Across the Rio Grande del Rancho, the folds in the Pennsylvanian die out. Fold B, however, probably is related to Laramide tectonics because the shales that separate individual monoclinally-folded conglomerate units show the effects of compression. Figures 33 and 34 are sketches of selected folds. Clearly, some of the smaller beds in the shale have been shortened, implying that monocline B should be considered as a compressional kink and not as a monocline related to the formation of the graben to the west (Faults #4 and #5).

The deformed zone is one half mile wide. The folds in the Pennsylvanian strike north and are asymmetrical with the short limb to the east. This fact alone suggests that the stress that formed the folds was from the west and the plunge of the stress vector is greater than 35° to the east. Figure 32 clearly demonstrates that the folds are part of a fold train which indicates that they are a feature of compression parallel to the bedding plane and not drag features along a fault.

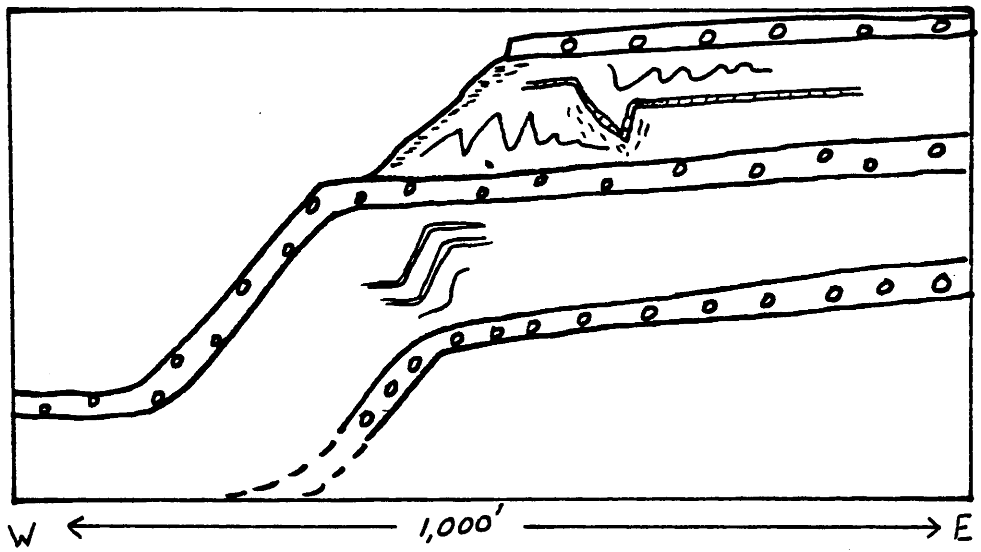


Figure 33. Detail of fold B looking south. Small folds in shale have different wave-length to large fold in the conglomerates.

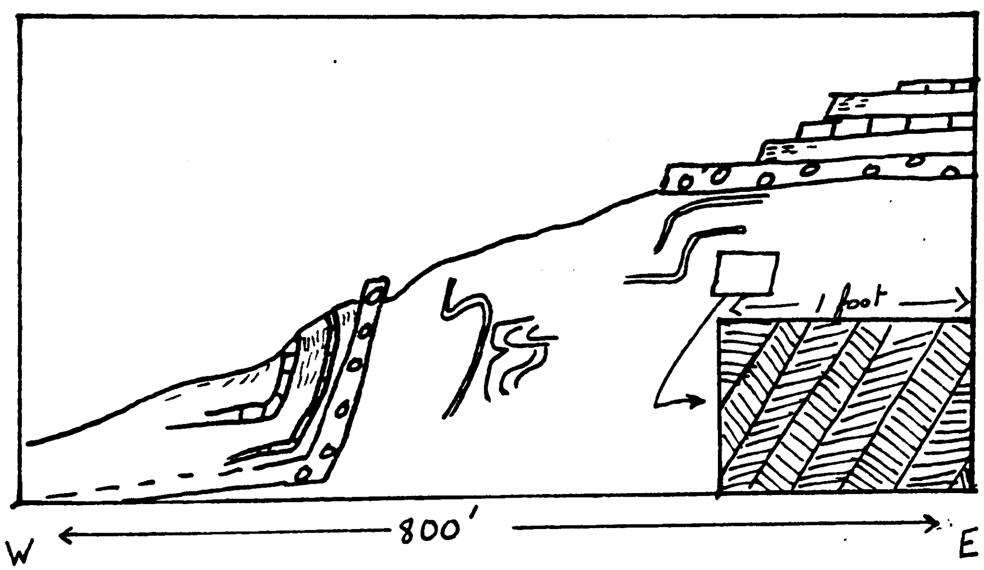


Figure 34. Detail of small "z" fold east of fold B. Insert shows small scale deformation along slip planes within the shale.

Laramide Thrust Faults

Two types of Laramide thrust faults could cause the rotation of a basement block. Stearns (1978b) has presented two styles of faulting along the margins of a basement uplift. One model involves shallow thrusting which is due, in part, to mushrooming of the uplift (see Fig. 27). The other style of uplift involves slivers of basement which are dragged up with the main uplift along steep reverse faults (see Fig. 28). The Jicarilla Fault, to the south of the field area, is of this latter type. The ensuing discussion models both types of faulting and compares them to details within the region.

Figure 35a shows a shallow thrust fault which originates at the Pecos-Picuris Fault (see insert), and emerges along the Rio Grande del Rancho. The basement-sediment contact is uplifted and rotated as it is carried over a curved fault plane. Even though the model shows that normal Fault #1 is consistent with the tensional stress generated behind curved uplifts (Stearns, 1979a, Fig. 26), three problems arise from this interpretation.

First, to achieve an 80^{o} rotation of the sediments, the fault plane must rotate 80^{o} from the transection of the sediments to their present location. Figure 35a clearly shows that this rotation must occur over 800 ft of vertical elevation of the fault; a rotation which is incompatible with the nature of thrust faults. The insert in Figure 35a shows that even a more moderate curvature of the fault is inconsistent with the location of the Pecos-Picuris Fault. It is practically impossible for the vertical Pecos-Picuris Fault plane to become horizontal over such a short distance.

Secondly, this model is compatible only with a fault having a very low dip angle. Therefore, the fault barely passes over the sediments on the east side of the Rio Grande del Rancho. These rocks form the footwall and it is highly unlikely that there would be no signs of internal deformation. Furthermore, the model does not explain why the Pennsylvanian sediments on the headwall form a fold train since fold trains are not drag features.

Finally, there is no evidence for the fault along the stream valley.

The second model postulates the fault to lie farther to the west. The model explains the upturn of the Mississippian-basement contact as a drag feature. Brecciation of the basement rocks, induced by the movement of the fault, and perhaps by weakening by deeper-seated splay faults, allows the contact to rotate as a hinge. Compression caused by this rotation, or by the weight of the headwall, folds the Pennsylvanian sediments. Fault #1, however, is unexplained by this model and presumed to be a younger feature.

Figure 35b shows that Fault #3 could have been the main fault. The next fault to the west is seen in Figure 19 (white triangles). These faults are modeled as splay faults of the Pecos-Picuris Fault. They show up well in aerial photographs (Skylab #SL2-10-019, 020, 021 and U.S.G.S. VAGK 1-127, 1-2, 1-131). Fault #3 continues beyond the mapped region as a pronounced lineament of hills. The fault at the base of the Picuris Mountains parallels the Pecos-Picuris Fault and

MODELS FOR GENERATING STEEP MISS.-BASEMENT CONTACT

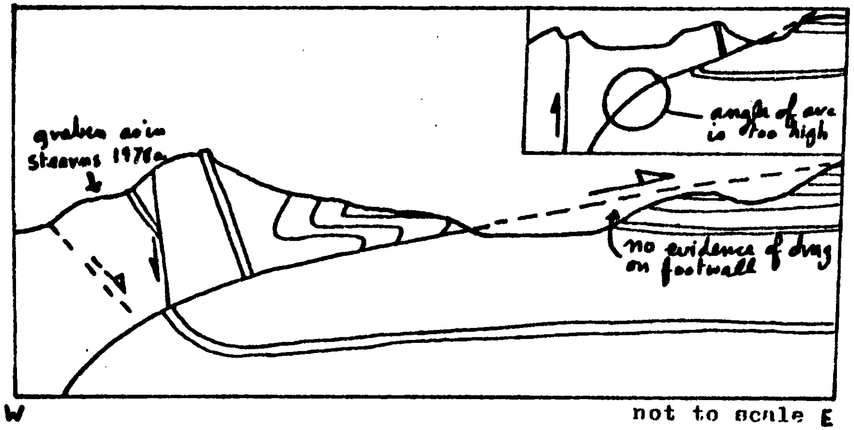


Figure 35a. . A shallow thrust fault rotates the contact along the plane of the fault.

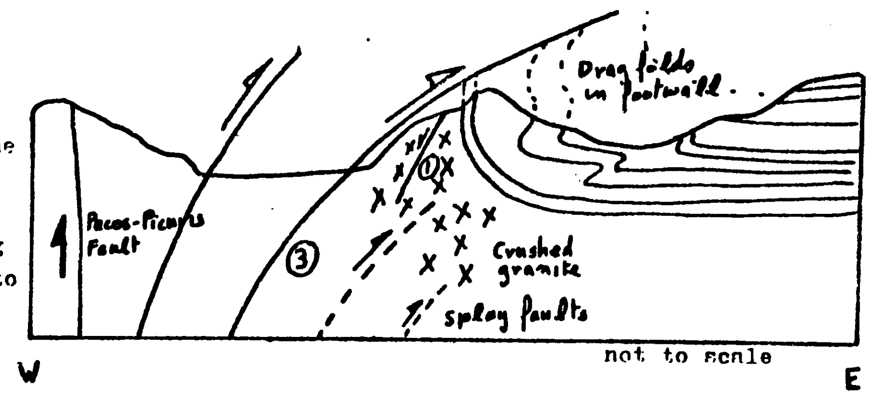


Figure 35b. Steep reverse faults form drag folds in the footwall. Brecciation of the granite by small splay faults which are propagating upward permits the granite to conform to the fold shape.

acts as their eastern margin. Though the fault has not been mapped, the straight outcrop pattern of the fault indicates its presence.

Turning back to the models presented by Couples and Stearns (1979b), I envisage that the rotation of the basement was caused by basement slivers, much like those in Figure 28. This rotation is documented in their experiment 302 (Fig. 37) where a dip angle of 45° was caused by steep reverse faults. Even though Fault #3 and the fault at the base of the Picuris Mountains have a normal sense of motion at present, it seems likely that these were thrust faults during the Laramide because the folding of the Pennsylvanian indicates that the stress field was compression from the west. The lack of a large stratigraphic offset in the Pennsylvanian can only be accomplished by drag along faults further to the west. Furthermore, this model is consistent with the behavior of the Jicarilla Fault (see Fig. 20), which also caused folding in the Pennsylvanian sediments. In conclusion, the second model (Fig. 35b) best fits the geological evidence and is the model favored here.

Behavior of the Basement

The next question is the curvature of the Mississippian-basement contact (see cross-sections A through H). Several interpretations are possible. Figure 35b shows that the rotation could be implemented by hidden faults which either die out or have remained undiscovered. A postulated hidden fault would enable the Mississippian-basement contact to rotate about its axis with the result that the basement would

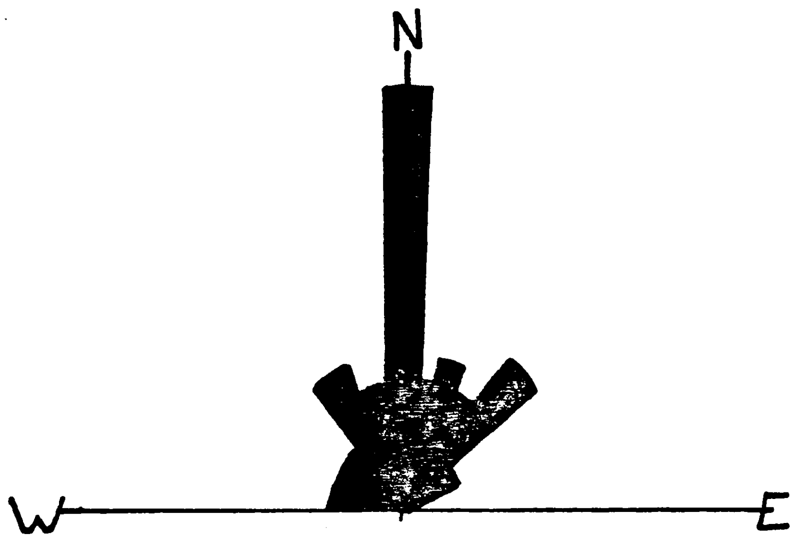


Figure 36. Joint sets in the Precambrian rocks of Fort Burgwin Ridge.

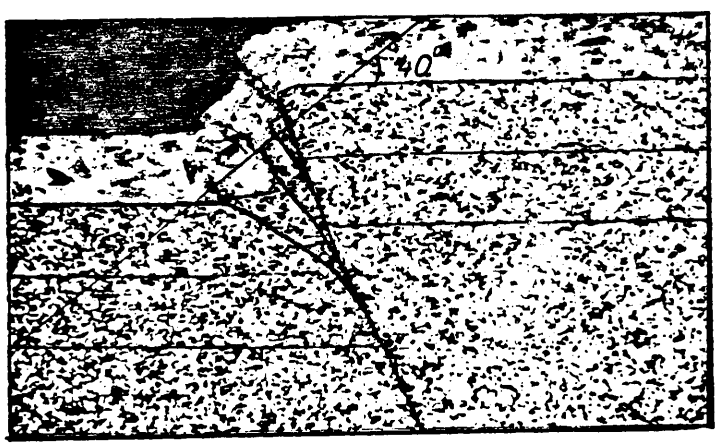


Figure 37 . Photomicrograph of experiment 302 showing splinter blocks developed in brittle sandstone forcing member. Note rigid body rotations and zones of cataclasis. From Friedman et al. (1976).

not have to assume a curved shape. The only drawback to this model is that the faults are conjectural.

The second model allows the basement, in this case granite, to conform to the curvature of the fold by small brittle failures. An article by Hudson (1955) describes the Fountain Creek Flexure. He observes that the basement was able to conform to the fold form of the overlying beds. The prime prerequisite for this motion is the weakening of the basement rocks prior to the deposition of the strata through the formation of zones of crushing, zones of fractures or minor faults. Stearns (1971) also describes a similar phenomenon along the southern corner of the Beartooth Mountains, Wyoming. The basement appears to follow the form of the fold, but where the basement-sediment contact is exposed, the basement is so completely shattered that it is capable of acting like loose sand. He observes that locally the basement does deform by cataclastic flow. He describes such a fold in Manitou Springs, Colorado, where the subadjacent granite conforms exactly to a monocline in Cambrian sandstone which has 100 ft of structural relief. The granite, however, is fragmented into 1 in pieces that can be removed by hand.

The granite in the field area satisfies these conditions. The granite has a cataclastic texture and a distinct foliation, both created during Precambrian deformations (see chapter on granite). Because of its cataclastic nature, I feel that the granite responded to the Laramide uplift by deforming on several scales. Small-scale offsets of quartz dikelets and joint sets with larger offsets are frequent up to one half mile from the contact (see discussion of Precambrian rocks). The rose diagram of joint sets, Figure 36, shows that there is a predominance of joints which trend north-south, parallel to the trend of the fold and the uplift in general. Secondly, the granite appears to have been deeply weathered prior to the deposition of Mississippian sediments. Actually, this phenomenon may not be weathering, but cataclasis. Next to the contact the granite was observed to be broken down to 50% of the average phenocryst size observed elsewhere. This feature was not observed everywhere along the contact and may have some explanation other than physical crushing, such as the vertical contact acting as a conduit for water.

In the northern part of Fort Burgwin Ridge, the porphyritic granite is replaced by aplitic granite whose fine-grained nature and lack of microdeformational and mesodeformational features contrast with the rest of the region. This granite has behaved in a distinctly different manner. The hillsides, from line B-B' north, are covered by a talus of 4 in blocks of aplite. Near the contact, the rock is even further shattered. Major joints are nearly parallel or are perpendicular to the strike of the contact.

Therefore, the texture and general condition of the granite indicate that it has responded to stress in a brittle manner. If this stress is Laramide in age, then the curved Mississippian-basement contact seen in cross-sections A through H and in Figure 35b can be explained by small-scale deformations in the basement. The Mississippian rocks would be able to deform, within certain limits, as though they were not attached to the basement. If this is the case, some interstratal slip must have occurred between the two members to allow for the change in radius of the fold.

As a final note to this argument, probably several factors contributed to the rotation of the Mississippian-basement contact. In my thinking, small-scale faulting, granulation, and jointing helped accommodate the rotation. However, the behavior of the Mississippian should not be considered as independent of the basement. Essentially, it is attached to the basement and both members have acted as a hinge which has tilted up in response to reverse faults in the west (see Fig. 35b).

Fold Style

The Pennsylvanian sandstones, unlike the Mississippian rocks, act independently from the basement. This is primarily due to the Morrowan shales low in the section which act as the decollement zone and flow into the anticlinal hinges of the folds. Sandstone units, where separated by thick shale units, behave as similar folds, maintaining the fold form but not themselves thickening or thinning (see Fig. 38). Up-section, the Atokan strata contain less shale, and the folds behave as flexural slip, parallel folds and, as a consequence the folds become tighter. This effect can be seen in cross-section C-C' and in Figure 38. The lower fold forms are more open, but the anticlines become more pointed higher in the Atokan section. Therefore, the folds are a continuum of styles from similar at one extreme to parallel at the other.

The folds are overturned folds which strike parallel to sub-

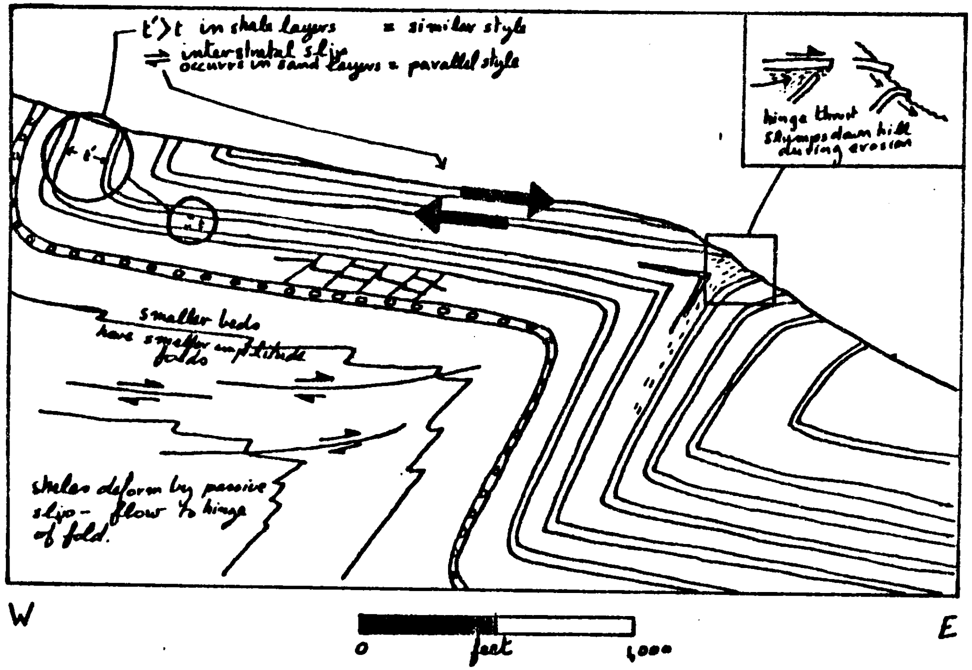


Figure 38. North-looking detail of fold C in Pennsylvanian rocks located on line C-C'. Note movement of shale into anticlinal hinges and insert of hinge thrust which explains the dip of the lower limb as a function of downslope creep.

parallel to the axis of the uplift. Table 2 shows that the lowermost folds plunge 5° to the south, but Figure 31 and Figure 32 indicate that the upper folds do not have the same trend. I attribute this to younger rifting of the northernmost end of the field area which opened the folds like a fan by drag across the northern tip of Fort Burgwin Ridge. Manifestations of this drag are the curvature of both the basement contact and fold C, and the formation of Fault #18 and perhaps Fault #2.

The western limbs of the folds are approximately 1,200 ft long and the eastern limbs are roughly 500 ft long. This latter figure is conjectural because no eastern limb is exposed in its entirety. Therefore, the fold train is asymmetric and verges to the east which is consistent with a model of thrust from the west. The amplitude of the folds is probably in the order of 700 ft. However, the fold train itself is probably short, existing only in the area of compression in the syncline formed by thurst. The deformation caused by the larger fold form probably dies out rapidly above the areas shown in the cross-sections (see Fig. 22).

Because the form of the folds changes from the bottom upward, the dips and angle of closure vary significantly. The folds directly overlying the shale appear to have a more gentle radius of curvature: dips along the Morrowan-Atokan contact sweep a cylindrical arc from 54° east to 42° west overturned, indicating approximately 90° of closure. Fold C, which represents the hinge line of the upper strata, has dips between 16°-22° for the upper limb and from 78° east to 35° west for the lower limb. This range of closure, from 125° minimum to 60° maximum, is probably due to north-south variation in lithology. The more open fold in the north is in a thin sandstone member surrounded by shale. A similar fold angle was noted in thin sands near cross-section C. However, the thicker sand units with less shale surrounding them has more closure. The thicker beds do not show hinges in this case. The hinges are absent and I suspect that hinge-line thrusting parallel to the long limb has occurred to allow the fold to assume the tighter form. In Figure 38 the dip of the overturned beds probably decreased due to down-slope slumping during erosion which made the angle of closure seem tighter than it really is (see insert in Fig. 38).

There is evidence that folding exists at a smaller amplitude and wavelength than the folds discussed above. These folds are seen in individual limestone and sandstone units within the shale. In the area near fold B, several small-scale folds were seen (see Fig. 34). These folds have limb lengths less than 12 ft and show considerable shortening. In the Morrowan shale, cross-section line D-D' crosses a small synform, cross-section line F-F' crosses a small anticline, cross-section C-C' crosses a small antiform in the western Atokan outcrops. I do not think that these folds are due to soft-sediment slumping because of their association with other folds. Rather, the folds reflect the importance of stratification in determining the geometry of their form. Where folds involve thin-bedded strata in an incompetent medium, small wavelength folds will form. Thicker strata or packages of strata will assume longer wavelengths. This fact has been demonstrated experimentally by Currie et al. (1962).

The folding in the shale is an indication that the shale has migrated into the hinges of the folds (see Fig. 38). In order for the Atokan sandstones to assume their shape, migration of the Morrowan shale from the synclinal axes to the anticlinal hinges must have occurred. This transport occurred by interlayer slip within the shale. As a consequence, the Atokan sandstone and shale sequence was able to act independently of the basement.

In conclusion, three scales of folding are found. The largest scale is the large synclinal bend formed by the tilted-up Mississippian-basement contact. This folding forced the Morrowan shales to migrate to the hinge of the fold. The Atokan sandstone reacted to the shortening by folds which were independent of the lower units because of the intervening thick shale. Intervening beds of smaller thickness acted with amplitudes and wavelengths which are a function of their bedding thickness and the amount of interference of intervening beds. The thicker beds reacted by flexural slip, forming parallel folds; as a consequence, their form became constricted upward developing cusped folds and finally small thrusts breaking the anticlinal hinges.

U.S. Hill

Setting

Sutherland et al. (1963) made a reconnaissance map of U.S. Hill as well as regionally. Their data are included in Figure 39. The region to the west of U.S. Hill is dominated by Picuris Mountain which is part of a Precambrian uplift extending from the Rio Grande Graben

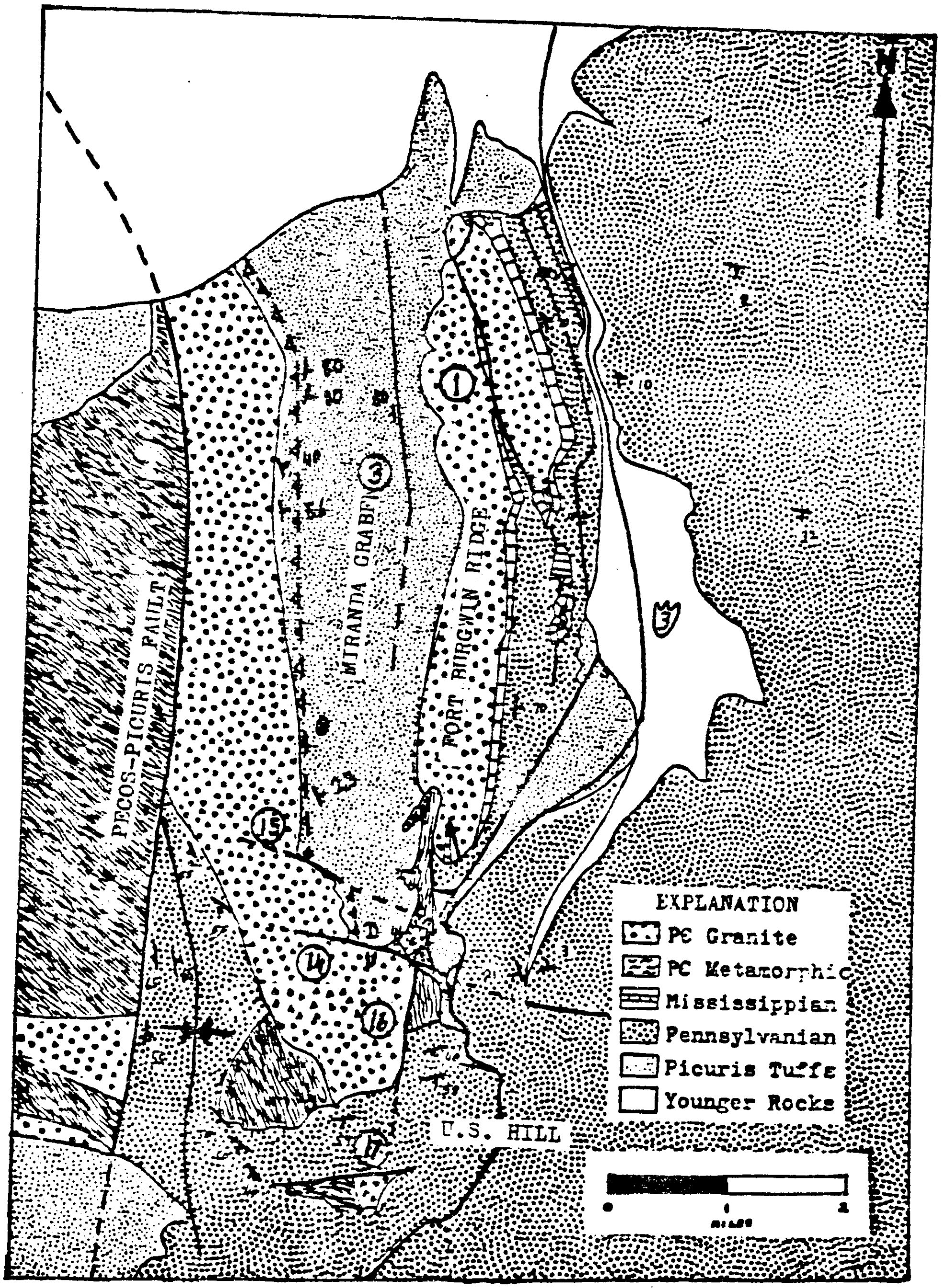


Figure 39. Sketch map from Talpa to U.S. Hill including data from Montgomery (1953).

to the north, to U.S. Hill in the south. It is bordered on the west by the Pecos-Picuris Fault and on the east by a post-Miocene graben (see Fig. 40). Only the southeastern part of the mountains is on my field map.

Lithology

In the field area, the mountain is made of Precambrian granite and Ortega metaquartzite. These are exposed from an altitude of 9,740 ft to become buried below 8,800 ft by Middle Desmoinesian strata to the south and southwest. The eastern and northern edges of the massif are fault boundaries. Middle Desmoinesian sediments were deposited on the Precambrian except in the area directly east of Fault #16. The Middle Desmoinesian rocks dip between 45° to 60° south and strike nearly east-west. Sediments to the north of Fault #14 have jumbled attitudes, and, probably, they were rotated along the fault. Good exposures along Route 3, between where Fault #14 crosses the road and the curve next to J' (see Plate 1), show that the Pennsylvanian strata strike nearly north, parallel to similar strata along Fort Burgwin Ridge. Preliminary exploration below these outcrops, along the Rio Grande del Rancho, show that the rocks across the river have gentle dips.

The region between Fault #14 and cross-section K-K' is very poorly exposed due to vegetation and a cover of Miocene tuff. This cover is unfortunate because it is the area which links the north-trending structures of Fort Burgwin Ridge with the east-west structures of U.S. Hill. Roadcuts along Route 3 three-quarters of a mile north of

U.S. Hill show that the Pennsylvanian strata strike north and dip to the west. These sediments look like Morrowan shales and their similar trend implies that they may be a continuation of Fort Burgwin Ridge structures. The westerly dips could represent Laramide folding due to uplift from the west.

Faults

Faults #16 and #17 and Fold I are Pennsylvanian structures. Faults #14 and #15 and the southerly dip of the Middle Desmoinesian sediments are Laramide structures.

Fault #14 shows stratigraphic offset: Morrowan rocks are juxtaposed against Precambrian rocks. This fault has an oblique sense of motion which produces both strike-slip and normal components of offset. It is considered to be a tear fault which dies out in the Pennsylvanian strata. A remarkably straight gash down the crest of a ridge on the eastern side of Rio Grande del Rancho is also part of this fault.

Fault #15 has been mapped in the basis of aerial photographs (U.S.G.S. VAGK 1-130 and 1-131). It separates a step-like offset in the Precambrian outcrops which is shown exaggerated in Figure 39. It probably dies out in my field area. The apparent offset of Precambrian Ortega and Pennsylvanian sediments along this fault trend may be purely coincidental. The presence of Ortega metaquartzite and Pennsylvanian shale was determined on the evidence of float down the eastern flanks of the ridge. This evidence is tenuous because the ridge is capped by Quaternary gravels which contain clasts of both rocks. However, there is a definite transition from Ortega to Pennsylvanian along the ridge. It has been suggested (Muehlberger, personal communication, 1980) that the Ortega outcrops are part of a terminal moraine or part of a talus slope because similar blocky outcrops have been seen along the Pilar-Vadito Fault line near Pilar.

Model of the U.S. Hill Region

The structure of the U.S. Hill area cannot be fully understood in isolation. It is a continuation of trends in the north and west. The transition from north-trending structures to east-trending structures occurs over a distance of one mile. Unfortunately, this mile is obscured by cover between Fault #14 and cross-section K-K'. Nevertheless, drape-folded structures in Colorado and Wyoming change strike over similar distances, and I feel that the structures of U.S. Hill and Fort Burgwin Ridge are part of a coherent whole.

Figure 39 summarizes Sutherland et al. (1963) and my data. The mountains to the east of Pecos-Picuris Fault, become separated from the fault and trend east (see Fig. 40). In the inverted V formed by the fault and the Precambrian rocks is a wedge of Pennsylvanian sediments which dip towards the fault. Drag along the fault has curled up the sediments, forming a syncline. The Pennsylvanian sediments appear to dip off the flanks of the massif and swing through an arc of 90^{o} to parallel the strata at U.S. Hill. The structure of the northern part of these mountains, therefore, appears to be the nose of an anticline which plunges into the Pennsylvanian sediments of the Taos Trough. The anticline is the southern continuation of the monocline formed by uplift on splay faults parallel to the Pecos-Picuris Fault.

Block diagram 40 is a model which explains and dramatizes the structure. It represents a northward view of the region, with U.S. Hill in the foreground. Fort Burgwin Ridge is in the northeast and the Picuris Mountains are represented as a block to the west. The splay faults flanking the Miocene graben are exaggerated. The post-Miocene graben and Miocene Picuris Tuff are not Laramide features, and they obscure the geology over the critical interval from Fort Burgwin Ridge to U.S. Hill.

The diagram shows the Pennsylvanian sediments which are deposited directly on the massif. They dip west and are curled up by drag along the Pecos-Picuris Fault. They swing around the nose of the fold to become offset by the Pennsylvanian Fault #16. The block diagram then strikes obliquely across the structures to show that Morrowan sediments on the south side of Fault #14 strike north-south and have steep dips; here interpreted to be folded like the Morrowan sediments at Fort Burgwin Ridge. Across the fault, the Pennsylvanian sediments are offset to the west and they show a variety of dips and strikes. There is a minimum of 2,000 ft of strike-slip offset on Fault #14. This offset is great enough that all folding, such as that at Fort Burgwin Ridge, could occur within the interval covered by the Picuris Tuff seen in Figure 40. East of Rio Grande del Rancho and under Highway 3, only nearly horizontal strata are found.

What is the nature of Faults #14 and #15 and the movement of

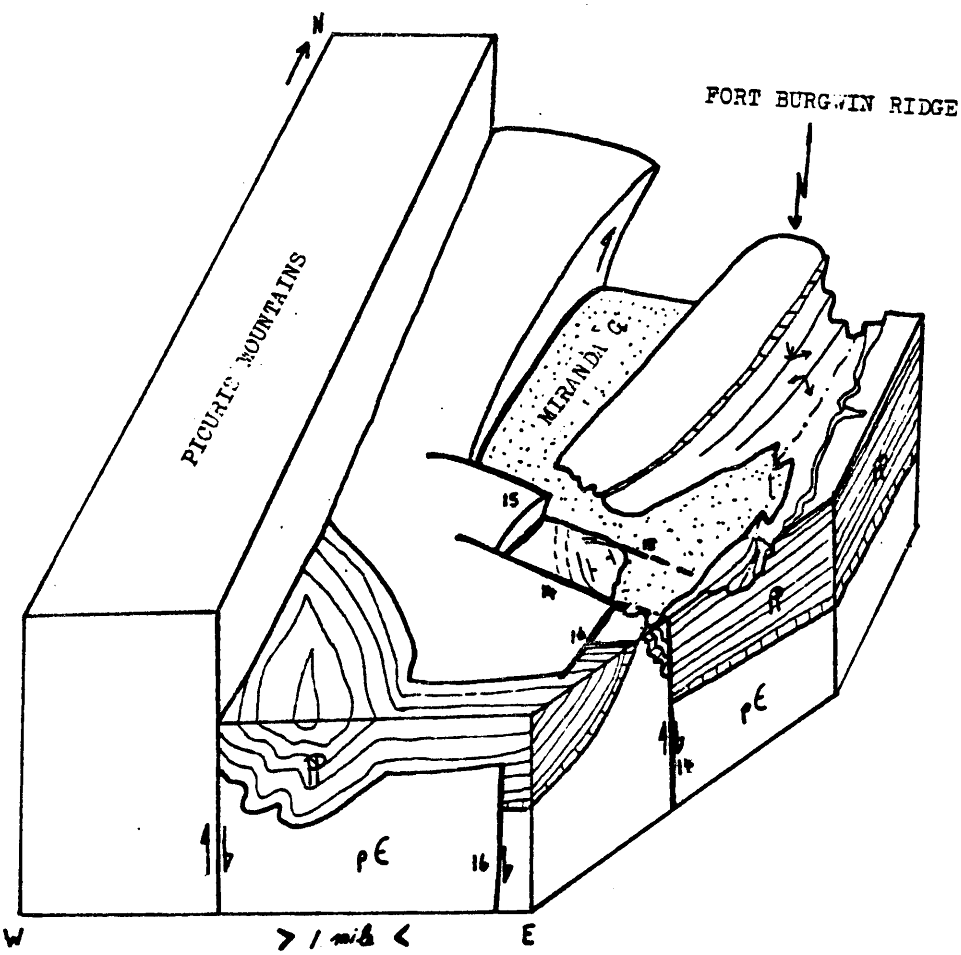
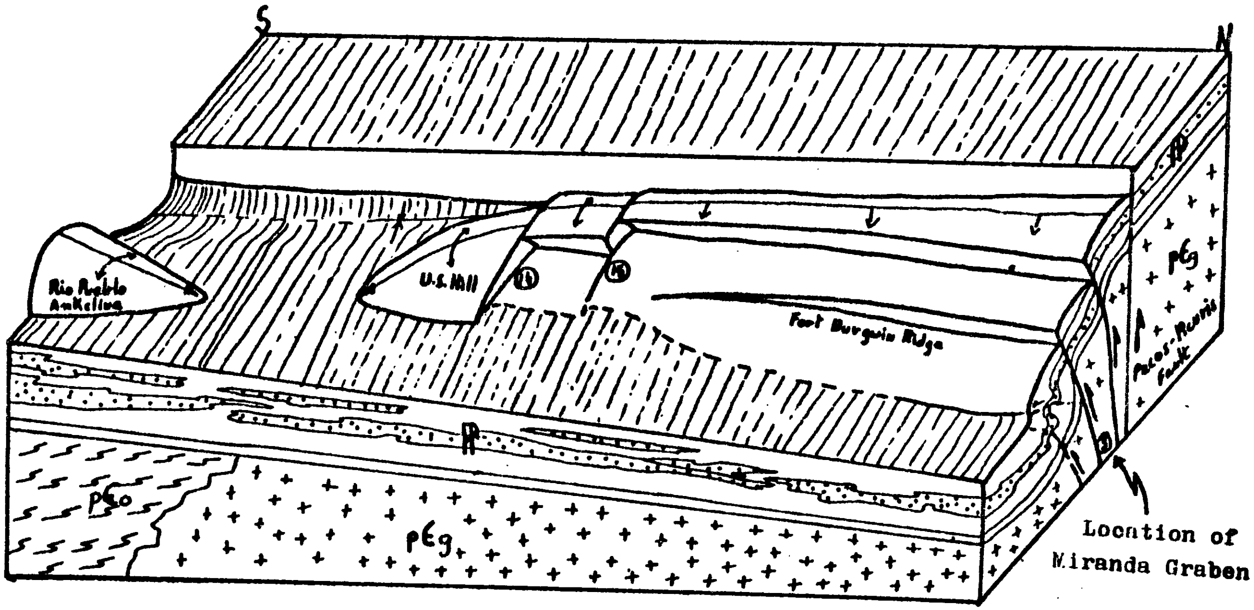


Figure 40. North looking view of field area with oblique corner. Laramide thrust faults are emphasized. Miranda Graben is a post Miocene feature which utilizes Laramide faults.

the massif away from the Pecos-Picuris Fault? Figure 41 shows the entire structure of the region during the Laramide. On the northern end of the block, the structure is just like that modeled in Figure 22. A monocline is formed by uplift of Pennsylvanian sediments along splay faults in the granite. Southward, this uplift appears less extreme and the monocline is offset away from the Pecos-Picuris Fault by Faults #14 and #15 and forms an anticline. Faults #14 and #15 in this model, are tear faults. I feel that their location is influenced by the change in basement lithology. Figure 19 shows that the basement in the northern region is granite. In the southern region - particularly in the west - the massif is made of Ortega metaquartzite. The Ortega is a layered competent unit. The granite, as has already been discussed, is a less competent rock which faulted easily. Therefore, the multiple splay faults in Figure 41 probably confronted a caprock of Ortega which acted as a barrier to their passage. This effect caused the monoclinal structure to die out. Whether this inhomogeneity forced the fold away from the Pecos-Picuris Fault, is a matter of conjecture, but certainly Faults #14 and #15 assisted in this motion.

Figure 41 shows the nose of the Rio Pueblo fold which plunges north. This structure was mapped by Sutherland et al. (1963). It has a core of Ortega metaquartzite. I feel that it is significant that the Rio Pueblo fold (which lies on the trend of the Jicarilla Fault to the south) should express Laramide deformation by fold structures rather than faults. Two factors may have contributed to this phenomenon. The first is the more competent nature of the Ortega. The second factor is

Figure 41. Block diagram showing differential uplift along the Pecos-Picuris Fault during the Laramide. Splay faults die out towards the south. Both U.S. Hill and Rio Pueblo anticlines are cored by Ortega metaquartzite. The transition from a monocline in the north to an anticline in the south is accomplished along tear faults #14 and #15.



that the Pecos-Picuris Fault had a much smaller throw between the nose of the U.S. Hill anticline and the Jicarilla Fault because only over this interval are Pennsylvanian sediments found on the west side of the fault. Figure 41 demonstrates the change in relief across the fault from north to south. This factor alone could explain the transition from an anticline to a faulted monoclinal structure. The anticline may be the first-formed structure which, with increased uplift, grew progressively westward along splay faults which migrated towards the master fault.

Summary of Laramide Structure

Exploration in the northern part of the field area revealed that the Mississippian-basement contact has a north strike and dips 80° to the east. Recumbent folds in the Pennsylvanian rocks to the east indicate that the deformation was caused by splay faults to the west. Figure 41 shows how a monocline was formed during the Laramide along a series of splay faults which were generated by the uplift of the Pecos-Picuris Fault and which cut the granites to the east of the fault.

Southward, the monoclinal structure is detached from the master fault, forming an anticline. On the westward side of the anticline, a syncline was formed with its western limb dragged up by the Pecos-Picuris Fault. Variable dips along the fault are explained as drag folding such as those seen along the Jicarilla Fault (see Fig. 20). The sediments dip approximately 55° off the western and southern flanks of the fold. Joints in the Precambrian rocks trend just west of north and may indicate the fold axis. It is presumed that the sediments turn around the nose of the structure and continue north along the eastern flank of the anticline, but this fact is not well substantiated.

Three and one half miles southward, a similar anticlinal structure has been formed by Laramide deformation. Its axis trends into my field area. The Pecos-Picuris Fault disappears from the nose of the U.S. Hill anticline to the Jicarilla Fault and Pennsylvanian sediments are found on the western side of the fault. Therefore, uplift along the fault was not uniform and anticlines appear to have formed where the uplift was least.

This fact suggests that anticlines were the first expression of the uplift along the Pecos-Picuris Fault. As the fault grew, splay faults uplifted slivers of basement and the initial anticline became part of a monocline. This idea is consistent with the behavior of splay faults formed by a piston in sand (Stearns, 1979a; see Fig. 25). The lowermost splay faults form first. As they migrate upward, each fault curves away from the main fault. Therefore the first surficial expression of these splay faults would be away from the master fault. Continuing deformation results in younger and younger faults that migrate towards the master fault and which have increasingly large surface dislocations.

Furthermore, the change in lithology at the nose of U.S. Hill anticline suggests that the competence of the basement rocks could influence the style of deformation. If the deformation pathway were deflected by a competent caprock, it seems likely that it would be expressed by a lateral component which would move the nose of the anticline eastward along Faults #14 and #15. The fact that the Tres Ritos anticline has an Ortega core supports the conclusion that lithology as well as a change in the amount of uplift may have influenced the formation of anticlinal structures.

LATE MIOCENE AND YOUNGER TECTONICS

Introduction: The Rio Grande Graben

Location

The Rio Grande Graben commences as a narrow horst in Colorado, and extends to form a broad collapsed vault in southern New Mexico. The graben is a series of right lateral, en echelon, linked basins. In north-central New Mexico, these basins are the Albuquerque Basin in the south, the Española Basin and the San Luis Basin which extends northward into Colorado (see Fig. 18). My field area is located on the southeasternmost margin of the San Luis Basin, near the junction of the Taos Fault and the Embudo Fault.

Timing of Rifting

The downwarp of the Laramide structural high postdated a Late Eocene erosion surface which now forms a structural datum by which to measure later events (Epis and Chapin, 1975). The calcalkalic volcanism which started approximately 26 m.y. ago was probably the first expression of the rift. The Abiquiu Formation and the Picuris Formation - both Oligocene to Miocene tuffs (and probably part of the same formation) - are preserved on the western and eastern margins of the San Luis Basin respectively. Presently, the Picuris Tuff is found at high elevations within my field area which indicates that there is a large structural throw along the Taos and Rio Grande Border faults.

In the early Miocene, a broad, 40-mile-wide depression formed. It was filled by bolson sediments which are dated at approximately 24-26 m.y. b.p. (Chapin and Seager, 1975). Baltz (1978) places the beginning of downwarp in the north at approximately 20 m.y. ago, at the base of the Tesuque Formation. The downwarp was accompanied by a change in volcanism from calcalkalic to the basaltic andesites which interfinger with the Tesuque Formation.

The actual beginning of the Rio Grande Graben rift valley at Taos postdates the Santa Fe Formation and predates the Servilleta olivine tholeiites. This places the period of widespread block faulting in the north from about 9-11 m.y. ago (late Miocene, early Pliocene) (Chapin and Seager, 1975; Christiansen and Lipman, 1972). The rate of extension across the rift has been on the order of 0.3 mm/yr over the past 26 m.y. (Woodward, 1977), and over the last 1.1 m.y. the rate may have slowed to 0.05 mm/yr (Golombek, 1980). Geophysical data (summarized by Cordell, 1978) show that the rift follows the margin of older Laramde and Pennsylvanian uplifts. Gravity surveys suggest that the faults follow a right lateral, en echelon grid which aeromagnetic data show to be part of the Precambrian structural grain. High heat flow and high electrical conductivity along the rift imply that there is an anomalous crustal and upper mantle structure under the axis of the rift.

Various models have been suggested to explain the rift. Certainly the uparched mountains with a central graben filled with rift valley sediments and a volcanic sequence from calcalkalic to olivine tholeiite suggests a continental rifting model. A concept of east-west regional extension and possible mantle upwelling is favored by many authors (Chapin, 1971; Chapin and Seager, 1975; Lipman, 1969; Lipman and Mehnert, 1975; Bridwell, 1976). Evidence also exists for transform faulting and left lateral movements. The amount of extension is approximately the same as the amount of compression that occurred during the Laramide and, in 1978, Cordell reasoned that the northern region (across the San Luis Basin) is a collapsed structural vault. Furthermore, tholeiitic basalts which were thought to be restricted to the northern part of the Rio Grande depression are now known to occur in the eastern Colorado Plateau and east of the Sangre de Cristo uplift (Aoki and Kudo, 1976). Therefore, further evidence is needed to substantiate a continental rift model and the term "rift" is used loosely in this paper.

San Luis Basin

The San Luis Basin extends from the Picuris Mountains northward into Colorado. In New Mexico, the basin appears to be a half-graben with a hinge line along the Brazos uplift to the west (see Fig. 18). The eastern border is downdropped along the Rio Grande Graben border fault which may be the northward continuation of the Pecos-Picuris Fault. Eastward from this line is the Taos Reentrant (a cusp-like bite in the Sangre de Cristo mountain front) which is outlined by the Taos Fault. This fault and its subsidiaries truncate the northern end of my field area (see Fig. 42) and Precambrian, Mississippian, Pennsylvanian and Miocene rocks are juxtaposed against recent alluvial fill. The southern boundary of the San Luis Basin is marked by the Embudo Fault which runs from Taos to the Jemez Mountains. The throw of this fault reverses along strike as the fault separates the down-to-the-west Espanola Basin from the down-to-the-east San Luis Basin. The Embudo Fault may merge with the Taos Fault and the Rio Grande Graben border fault as it passes near my field area (see Fig. 43).

Chapin (1971) postulates 36,000 ft of basement relief within the basin and approximately 30,000 ft of alluvial fill. Muehlberger (1979) indicates that there is 10,000 ft of structural relief across the Embudo Fault itself. Most of the alluvial sediment in the San Luis Basin has been covered by the Servilleta Basalts which range from 4.5 m.y. before present (Ozima et al., 1967) to 2.8 m.y. ago (Manley, 1976). Since their extrusion, the basalts have been cut by the Rio Grande which flows southward down the San Luis Basin and turns southeast around the Picuris Mountains to follow the Embudo Fault into the Española Basin.

Embudo Fault

The Embudo Fault runs from the Taos Fault to the Jemez Caldera and it is part of the Jemez Lineament which extends from Springerville, Arizona to the Raton-Clayton-Capulin volcanic field in eastern New Mexico. It trends N60^{o}E, it displays as much as 10,000 ft of structural relief, and it is older than 7 m.y. because Precambrian rocks are in-

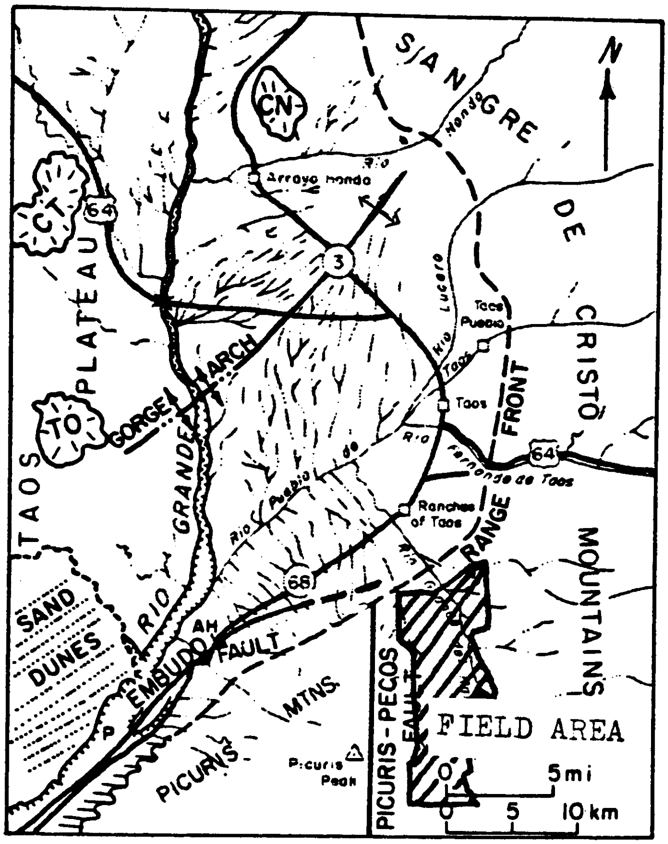


Figure 42 . Location of field area in relation to the Embudo Fault, the Taos Fault, which is part of the Range Front, and the Picuris Mountains. Adapted from Muehlberger (1979).

cluded in the Ojo Caliente sandstone member. Muehlberger (1979) has mapped reverse faults along road cuts across New Mexico Highway 68, ten miles south of Ranchos di Taos (see Fig. 42). The Embudo Fault, therefore, consists of several reverse splay faults and, to the east, they curve toward the Pecos-Picuris Fault and my field area. The Embudo Fault probably merges with the Taos Fault and the border fault, but the Taos Reentrant is full of recent alluvium making the postulated relationship of faults unclear.

The Embudo Fault follows an earlier Precambrian geofracture which is now acting as the boundary fault between the San Luis Basin and the Española Basin. It has been suggested that the fault is a left-lateral transform fault which offsets the two basins to the right. Lisenbee and Woodward (1979) suggest that the Tijeras-Cañoncito Fault which offsets the Española Basin from the Albuquerque Basin is also a transcurrent fault. Both faults show later Tertiary left-lateral movement with large components of dip slip.

If the Embudo Fault was a typical transform fault due to normal extension across the rift, then the fault would be a pull apart zone. Compression across the fault is indicated by reverse faults in the alluvium and by the Gorge Arch which parallels the fault in the middle of the San Luis Basin (see Fig. 42). These structures show that there must be north-south, left-lateral slip along the rift itself (Kelley, 1977). Muehlberger (1979) points out that the intersection of northeast-trending faults (Embudo and Tijeras) with north-trending faults forms a series of diamond-shaped basins. Compression along the Embudo Fault implies that the diamonds are being shortened along their short diagonals (rotated counterclockwise). Though left-lateral slip has not been demonstrated along the Embudo Fault, the right-lateral offset of the two basins implies that there is a strike-slip component along the fault commensurate with the offset of the basins.

It is my feeling that the Embudo Fault is a transpressive fault which expresses both north-south left-lateral movements along the rift and east-west extension across the rift. This compression would make the Picuris Mountains rotate in a counterclockwise manner forming a graben to the west of Fort Burgwin Ridge (see Fig. 43). A series of faults within the area parallel the Pecos-Picuris Fault and their normal sense of slip accommodates the opening of the graben. Several of these faults (Faults #1 and #6) show horizontal slickensides. It is probable that these horizontal movements are minor and that they are a direct result of rotations in the Fort Burgwin Ridge block as it rotated into the opening graben to the west.

Post-Miocene Structural Features in the Fort Burgwin Ridge Area

The post-Miocene structures in the Fort Burgwin Ridge area are summarized in Figure 1. Essentially, two major events have occurred since the deposition of the Miocene Picuris Tuff. There are (a) the truncation of the northern end of the field area by subsidiary faults of the Taos Fault and (b) the opening of two grabens, one on each side of Fort Burgwin Ridge. At present, Fort Burgwin Ridge is a horst with grabens to the west and southeast and an incipient graben to the north-

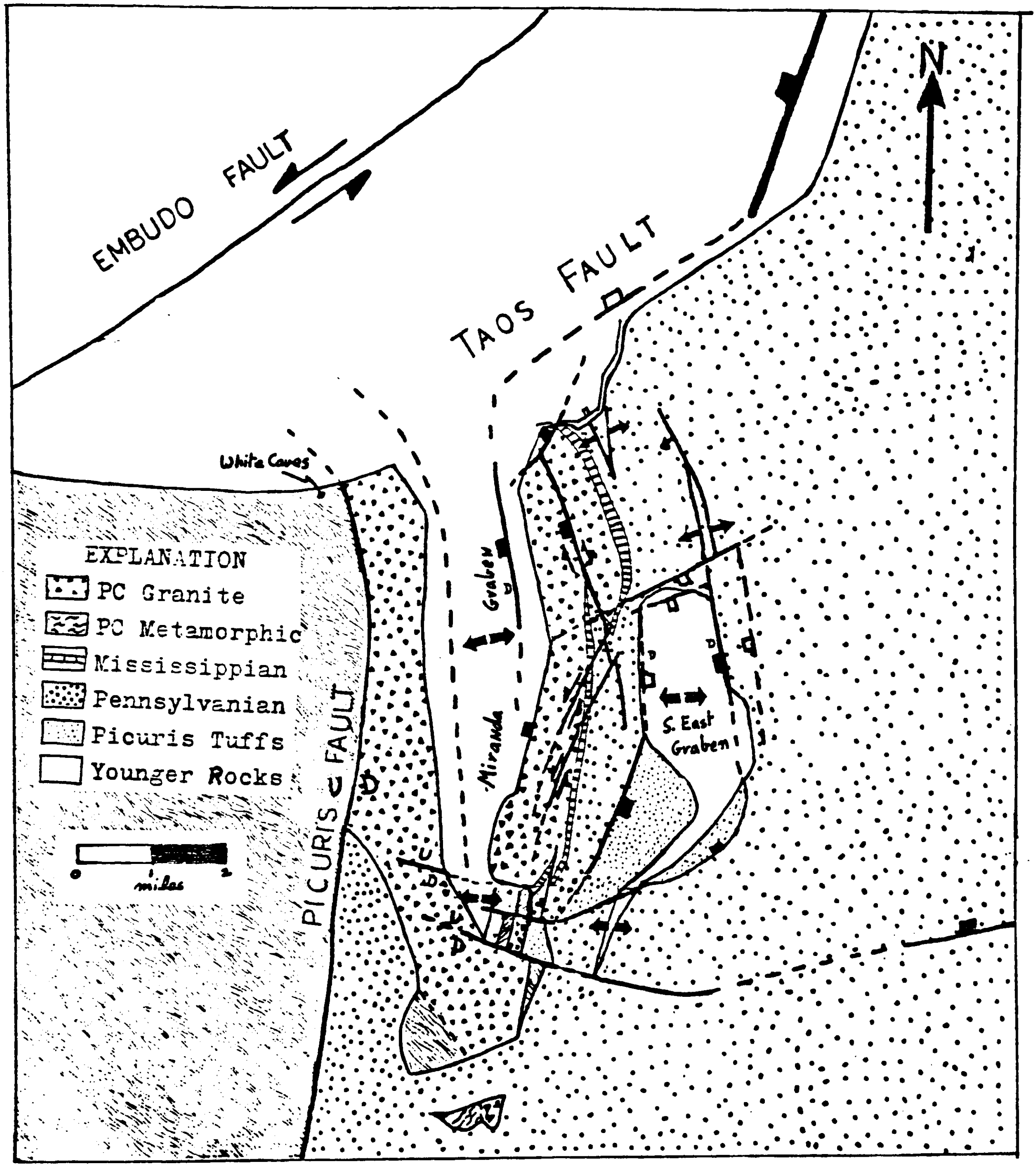


Figure 43. Generalized map of the region showing Miocene and younger structural elements. Divergent arrows emphasize rifting within the field area.

east. The block-like nature of the horst is emphasized in Figure 45. It is the thesis of this paper that transpressive (strike-slip) forces along the Embudo Fault have rotated the Picuris Mountains counterclockwise, opening the wedge-shaped Miranda Valley to the east of the Pecos-Picuris Fault. Normal faulting has dissected the region into blocks which also have undergone rotational movements to accommodate the rifting of the area.

The Northern Boundary

The Taos Fault lies approximately one mile north of my field area (see Fig. 43). Faults #21, #21a, #21b, #21c truncate the northern edge of Fort Burgwin Ridge and are subsidiaries of the Taos Fault which forms the range front scarp (see Plate 1). For example, Fault #21 lies northeast of the steep slope which divides Pennsylvanian sediments to the south from alluvial sediments to the north. The fault has been inferred to be out in the alluvium but close under the Pennsylvanian bluffs.

Fault #21a juxtaposes Picuris Tuff with dips of 25° to the north against Pennsylvanian and Mississippian sediments. Young (1945) described this fault as an unconformity, but Montgomery (1953) observed dips of 50° south towards the older sediments. The faulted nature of this contact can clearly be seen from Route 3 at the entrance of the Rio Grande del Rancho canyon near Talpa. Pennsylvanian sediments which strike north along the hinge line of fold C are abruptly truncated by Picuris Tuff beds which dip 25° to the north. The tuff beds are cut off by Pennsylvanian sediments and their dip at this location may be caused by drag along the fault.

Fault #21b was mapped by Montgomery (1953) who observed that both depositional and fault relationships were present along the contact between the Precambrian granite and the Picuris Tuff. It is my feeling that Arroyo Miranda follows this fault for approximately one mile. Fault #2, therefore, may be subsidiary to Fault #21b.

There is a distinct change in elevation which strikes east-northeast across the north end of Miranda Graben. This lineament probably marks the continuation of Taos Fault trends across the valley. It has been mapped tentatively as Fault #21c (see Plate 1).

These faults (numbers 21, a,b,c) are fairly unusual in that they do not follow the general north-south structural grain. Because they clearly related to Rio Grande Graben border faulting, they will not be discussed further as most of the pertinent structure exists outside the scope of this study.

The Western (Arroyo Miranda) Graben

The Miranda Graben is flanked on both the east and west sides by fault-bounded mountains of Precambrian rocks. The western margin of the graben is formed by a steep splay fault of the Pecos-Picuris Fault. This splay has reversed its Laramide motion and now acts as a normal fault in response to post-Miocene tensional stress in the region.

The fault can be seen on aerial photographs GS VAGK 1-130 and Skylab SL2-10-019, 020, 021. In the field, the contact between the

Picuris Tuff and the Precambrian granite shows both sedimentary and fault relations. For example, one mile southeast of the White Caves (see Fig. 43) Montgomery (1953) observed that beds of Picuris Tuff which dipped nearly vertically were found near a granite-tuff contact which was depositional. Just south of this area the fault is clearly visible and Picuris Tuffs are tilted east at an angle of 50^{o}. Further south, these beds dip less steeply and the graben is probably shallower.

The eastern flank of the graben is marked by Fault #3 and #3a (see Fig. 44). Fault #3 is defined by a north-trending line of hills which are held up by more resistant beds of Picuris Tuff. Picuris pyroclastics are seen dipping gently to the west on the eastern side of the fault and steeply to the west across the fault. Therefore, the Picuris Tuff appears to be curled up along both its eastern and western margins by drag along the faults and, consequently, the rocks form a syncline (see Fig. 44).

Fault #3a parallels Fault #3 and may be the continuation of Fault #21b. It follows Miranda Canyon south and curls around the southern end of the Fort Burgwin Horst block. Tuffs to the east of this trend appear to be in depositional contact with the granite. Therefore, Montgomery's (1953) map which shows the fault as a line which separates granite from tuff seems to be an oversimplification. Fault #3a lies in Picuris Tuff for much of its length and may join with Fault #13 at the south end of Fort Burgwin Ridge.

Fault #2 lies in the extreme north end of the field area. It is a normal fault which forms a prominent fault plane along a 200 ft

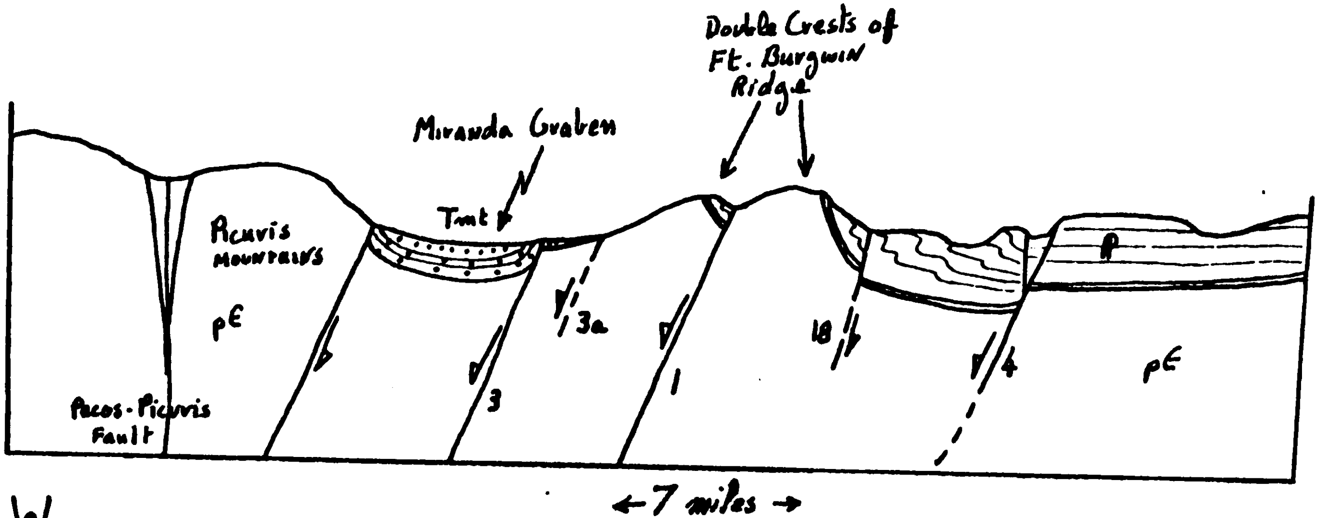


Figure 44a. Schematic cross-section across north end of the field area from Picuris Mountains to eastern edge of map.

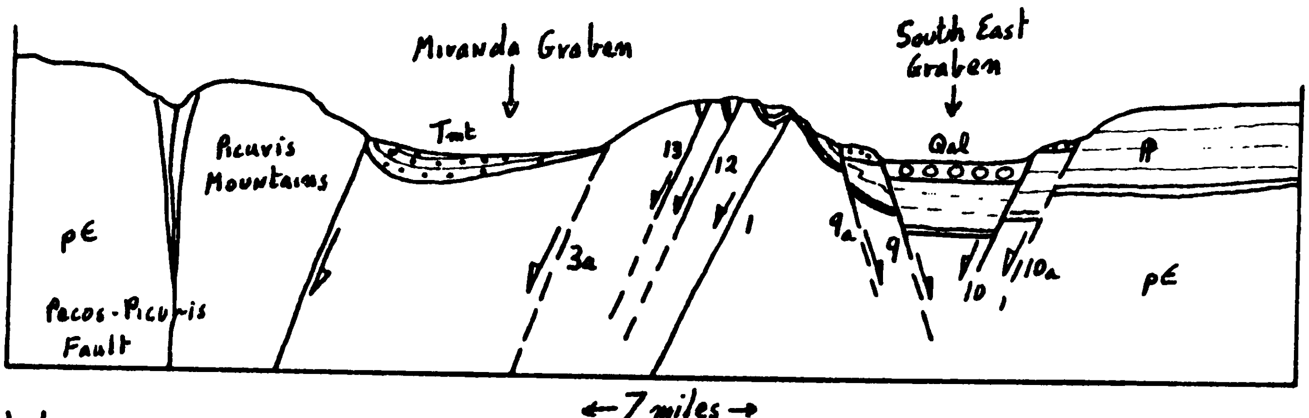


Figure 44b. Schematic cross-section across south end of field area.

long outcrop which strikes north-northeast from Ponce de Leon Hot Springs. The scarp has dip-slip grooves in it and a five-foot-wide breccia zone on the west side of the scarp. The scarp crops out northward and, at the northern end, the fault divides repeated sequences of Mississippian and Morrowan sediments. Along the fault plane, between the repeated Mississippian units, a test pit shows some manganese oxide (manganite or pyrolusite) mineralization along joint planes.

The junction of Fault #2 with Fault #1 is located at the hot springs of Ponce de Leon. The source of the springs lies directly on Fault #2. This probably attests to a recent age for the two faults because the fault is still sufficiently open for fluid migration.

Fault #1 extends south-southeast from the springs, and has a gouge zone of crushed granite approximately two feet wide. At the intersection of the fault and cross-section line A-A', the fault is five feet wide with low, vertically-dipping walls which show slickensides which plunge 10° north (see Plate 1). The area between the fault walls is crushed granite with epidote and calcite mineralization.

Slivers of Mississippian rocks appear along the fault at the intersection of cross-sections C-C' and D-D' with the fault line. The fault plane dips nearly vertically and slickensides and grooves plunge between 10° and 0° north. The throw of the fault at this location may be in excess of 3,000 ft (see cross-sections C-C' and D-D'). Just west of these outcrops there is a small fault with a dip of 28° west and strike due north which shows epidote mineralization. This fault is probably a subsidiary of Fault #1.

Following the fault further south, Paleozoic rocks form a north-striking outcrop wedge within the granite and the fault lies along the eastern edge of the contact. The wedge shape indicates that the throw of the fault is decreasing southward. At cross-section D-D' the throw of the fault is approximately 2,000 ft and south, at F-F' the throw is only 500 ft and becomes even less southward.

At the intersection of cross-section F-F' with the fault trend, a high ridge of Mississippian rocks forms a shoulder which diverts the westward-flowing channel at F-F' 500 ft due south. The fault plane probably lies at the bottom of this elbow. Mississippian rocks cross the stream, form an anticlinal nose pointing south, and reappear briefly on the western side of the fault. This structure is interpreted to be the breached nose of anticline D in cross-section F-F'. Anticline D reemerges 2,000 ft to the south and it is paralleled by a syncline to the west. The anticline-syncline pair terminates against Fault #6.

Both Fold E and Fold F are similar structures to Fold D. They have been left-laterally displaced along tear Faults #6 and #7. Fold E has a steep westerly dip along its western flank and the eastern flank forms a gently dipping hillside of limestone. The fold appears to plunge to the north, however, it is probably a doubly plunging feature like Fold F and the southern portion of the fold has been faulted out by Fault #7.

Fold F is the continuation of Fold E and it has an exposed granite core (see cross-sections G-G' and H-H'). The fold is paralleled by syncline G. Traces of these two folds continue southward into areas of heavy vegetation and poor exposures. Just west of Fold F are additional exposures of Mississippian rocks which may be the result of fault slivers or may be a function of the steep slope of the ridge at this point.

I feel that Fault #1 underlies Folds D, E, and F and that the fold forms are essentially drapes over the fault edge. Syncline G and related synclines are formed on the downdropped side of the fault. Because of the evidence presented by these folds, the north-pointing wedge of Paleozoic sediments, which appears to indicate a right-lateral fault of some magnitude, is interpreted to be a normal-faulted remnant sliver. The horizontal slickensides which show all along the fault are, therefore, considered to be minor features which probably occurred after normal faulting, at the same time as movements along Faults #6 and #7. Baltz (1978) and Lisenbee (1979) mention horizontal slickensides related to Rio Grande Rift movements. Both authors feel that very small scale movements are sufficient to form pervasive horizontal slickensides even though most of the movement along the faults was vertical. Moving southward, the southern edge of Fort Burgwin Ridge has been transected by Faults #12 and #13. Fault #12 is very similar to Fault #1, forming a north-pointing wedge of Paleozoic sediment in the surrounding granite (see cross-section L-L'). It is paralleled by a similar, but more minor, fault.

Fault #8, near cross-section H-H' parallels Faults #1, #3, #12 and #13 and it probably marks similar normal movements.

Fault #13 is more complex. Because of its shallower dip and its location along the steep western flank of Fort Burgwin Ridge, its outcrop pattern wanders around the stream beds which incise the fault plane. For example, adjacent to #13 on the map, there is a sliver of Paleozoic rocks which lies on top of the granite. The fault contact goes up the slope to the crest of the hill, crosses to the next arroyo and descends to the valley bottom to return to the crest of the hill on the other side of the stream. Fault #13 dies out to the north and it is possible that it is intercepted by Fault #3a. Southward, the fault follows a stream valley which separates granite to the north from Pennsylvanian sediments to the south. The fault truncates the Mississippian rocks and disappears southward under Pennsylvanian sediments. Fold H is really a faulted anticline (see cross-section K-K'). Southward, Fold H follows the base of a steep-sided, north-trending ridge until it is intersected by Fault #11. Unfortunately, the area to the south is heavily wooded.

The outcrops of Ortega boulders (see map and cross-section K-K'), are probably a periglacial feature which forms a resistant ridge to the west of Fault #13 because the rocks are extremely resistant to erosion and are either a moraine or part of a talus slope. In cross-section K-K', they are shown deposited unconformably on Pennsylvanian rocks. Faulting, prior to the glacial outwash terraces (Quaternary gravel on the map), lowered and preserved these outcrops.

Therefore, the western side of Fort Burgwin Ridge has been cut by a series of down-to-the-west normal faults. The block diagrams in

Figure 44 a and b show that the Miranda Graben is a down faulted syncline bordered on both flanks by west-dipping faults. It is my interpretation that Faults #1, #3, #3a, #12 and possibly Faults #18 and #4 are reactivated Laramide thrust faults whose sense of motion during post-Miocene rifting of the area have been reversed. A comparison of Figures 40 and 41 on pages 112 and 114 and Figure 44 demonstrates that the area is dissected by parallel reverse faults and that the Miranda Graben would sink down like the keystone in an arch if the northern end of the block diagram in Figure 41 were rifted open. Many authors feel that Laramide faults have reversed their motion with the relaxation of stress since Laramide time. Certainly, the strike and the abnormally steep dip toward the Pecos-Picuris Fault suggests that the Pecos-Picuris Fault is the master fault.

The Southeastern Graben

The southeastern graben is a standard graben flanked both east and west by progressivly downdropped slivers along inward dipping faults (see Fig. 44b). Fault #9 follows the western edge of the Rio Grande del Rancho valley forming a gentle scarp which hooks to the east at its southern end. Fault #9a parallels Fault #9 and lies to the west. It shows on aerial photographs and separates flat platforms of Picuris Tuff to the east from steep slopes of Pennsylvanian sediments to the west. Both Faults #9 and #9a intersect Faul #11 which trends south-southwest. This fault shows up as a slope break marked by a lithologic change from Pennsylvanian rocks to Miocene rocks. Montgomery (1953) shows a similar fault on his map.

The eastern part of the graben is flanked by Faults #10 and #10a. Fault #10a apparently separates Picuris Tuff from Paleozoic sediments and it may also trend into Ojo Sarco Canyon.

The northern flank of the graben is formed by monocline A which is marked by an abrupt change in dip of the Pennsylvanian strata from 8° southwest to a maximum of 54° south. Early Desmoinesian shales are thought to underlie the valley alluvium and probably contribute to the low elevation of the area.

The Northeastern Incipient Graben

The area northeast of Fort Burgwin Ridge also shows signs of rifting. Faults #4 and #5 form a small wedge-like graben (see cross-sections C-C' and D-D' and Figure 44a). Fault #4 appears to be the master fault with the greatest stratigraphic offset - approximately 1,000 ft. The Middle Desmoinesian sediments which are on the eastern side of the fault form a series of resistant synclinal outcrops along hill crests. The syncline has been made by drag along the fault.

Fault #5 is an antithetic fault to Fault #4. Displacement along the fault dies out northward as it changes from a fault to a monocline. Atokan sandstone layers dip nearly vertically along segments of the fault and, in some places, roll over without rupture.

Fault #19 is subparallel to Fault #4. It has a throw of only a few feet and its origin could be related to any period of faulting, including movements related to Fault #4.

Fault #18, in the northern end of the field area, is expressed by vertical beds along the fault and a deep stream bed. Furthermore, geometrical interpretation of the Laramide folds in the Pennsylvanian rocks shows that fanning out of the fold hinges has occurred at this location. Therefore, rifting on the northeastern flank of the ridge is probably occurring.

Figure 45 dramatizes how the soutern part of the Fort Burgwin Ridge is essentially isolated by flanking grabens. In the northeastern part of the area, the ridge is still attached to the flat-lying Pennsylvanian sediments to the east, but Faults #4, #5, and #18 indicate that the horst is pulling away in the north also.

Rotational Movements

The Mississippian outcrops along the Fort Burgwin Ridge crest form a gentle arc as the overall strike of the ridge changes from north 30^{o} west to south 55^{o} west - a total rotation of 55^{o}. Most of this rotation occurs at a pivotal point which is defined by a distinct kink in (1) the Mississippian outcrop, (2) the contact of the granite and tuff, and (3) in Fault #3a. This transition occurs between cross-section lines E-E' and F-F'. Fold A and Fault #20 both lie on a trend normal to the northern part of the ridge - north 60^{o} east - and this trend is dramatized in Figure 43.

Fault #20 has a minor offset of 20 ft. A small sliver of Mississippian rock appears 200 ft to the west of this fault and indicates either a complex movement along this fault or a splay fault parallel to

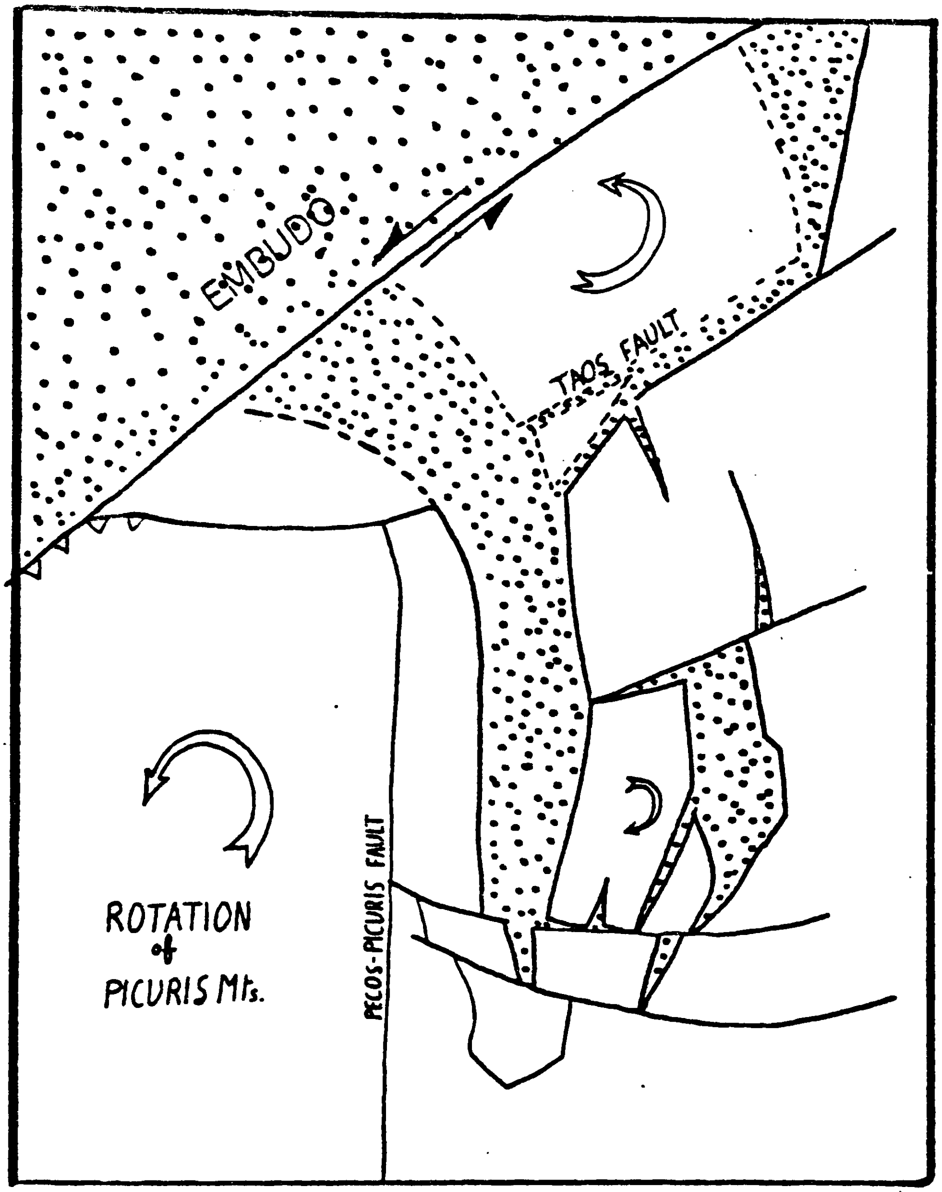


Figure 45. Diagram showing block-like nature of the field area. Stipled areas are low structural elements.

Fault #1. Across the Rio Grande del Rancho, monocline A truncates the small graben formed by Faults #4 and #5. It separates strata with gentle south-southeasterly dips from strata which dip up to 54° south. It is my feeling that these features (Fault #20 and Monocline A) are the consequence of the rotation of the southern part of the Fort Burgwin Horst clockwise away from the northern part (see Fig. 46). The rotation results in the opening of a widening wedge to the east along the northern boundary of the southeast graben.

Figures 45 and 46 demonstrate several consequences of this rotation. First, clockwise rotation of the southern block explains the opening of the southeast graben and the formation of Fault #11. Secondly, this rotation results in movement along the trend of Faults #14 and #15. In the diagram, these faults are shown extending eastward to connect with Rio Grande Graben border faults which show as deflected stream beds to the east. Thirdly, this rotation might explain normal faulting at #12 as a drag feature caused by strike slip movements which open Fault #12 like a fan. Finally, the clockwise rotation of the southern block explains the presence of Faults #6 and #7.

Figure 46 dramatizes Faults #6 and #7. The presence of these faults was first postulated on the evidence of the left-lateral offset of the Mississippian outcrops at folds D, E, and F (see map). However, only minor offsets in the Mississippian rocks at the ridge crest were found, but investigation along the trend of these minor breaks and the fold offsets uncovered Fault #6 in the granite. It has a gouge zone one foot wide of epidote altered crushed rock, a vertical dip, and

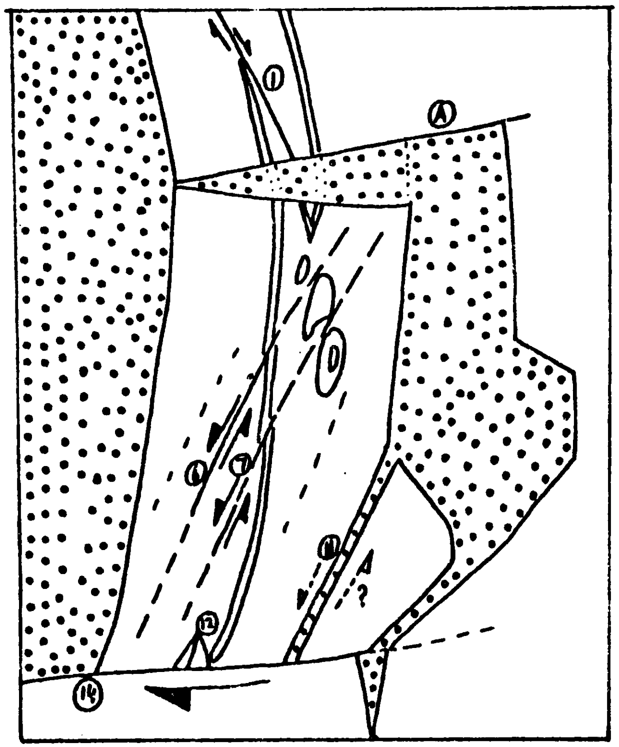


Figure 46a. Left-lateral movements on Faults #6 and #7 causes offsets in the Mississippian rocks.

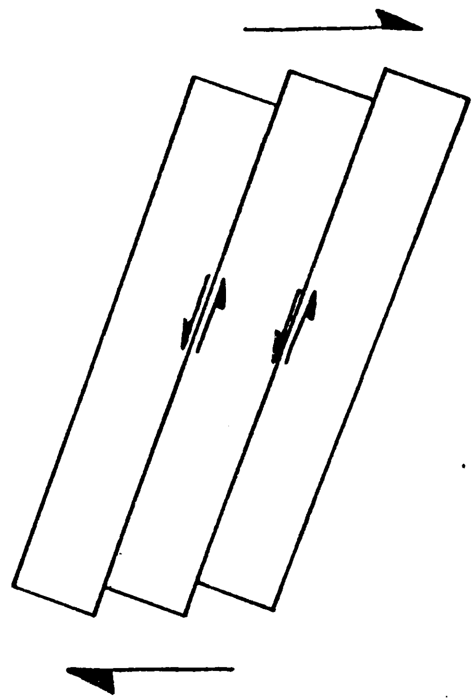


Figure 46b. Right-lateral shear upon a dissected block causes left-lateral strike-slip movements.

slickensides which plunge 22^{o} south-southeast. It is my interpretation that minor offsets along this fault, in combination with the steep eastern slope of the ridge, are sufficient to produce the left-lateral offsets seen on the map. As a further note, Fault #11 parallels Faults #6 and #7 and, consequently, may be considered as a similar fault sliver with strike slip components (see Fig. 46).

Probably a clockwise rotation of the southern block has occurred. This right-lateral shear couple on the rigid block has produced a set of north-northeast trending fault slivers within the block which have moved counterclockwise (left-laterally), in order to accommodate the rotation (see Fig. 46). Furthermore, if the southern block has rotated clockwise, then a counterclockwise stress must be induced in the northern block. Fault #1, as has already been noted, has prominent horizontal slickensides which indicate right-lateral movement along the fault.

Therefore, the left-lateral movement along Faults #6 and #7 (1) confirm that monocline A was formed by the wedge opened by the clockwise rotation of the southern block; (2) explain horizontal slickensides on Fault #1; (3) explain the greater opening of the southeastern graben than the northeastern graben due to the greater rotation of the southern block; (4) support the theory that rotational movements have been induced by strike slip movements along the Embudo Fault.

Conclusions

Left-lateral movement along the Embudo Fault has resulted in thrust faults in the alluvium to the northwest of my field area. It is my thesis that this compression has resulted in the counterclockwise rotation of the Picuris Mountain spur which creates a zone of extension to the east of the Pecos-Picuris Fault (see Fig. 43). Rifting of this area formed Miranda Graben on the eastern flank of Fort Burgwin Ridge (see Fig. 44 a and b). Miranda Graben is formed almost entirely by normal movements along reactivated Laramide thrust faults. Continued rifting has isolated Fort Burgwin Ridge and the area is now dissected into blocks (see Fig. 45). The Fort Burgwin Ridge horst is not entirely detached on its northeast side though graben formation has started in this area. In the southeast, a graben has formed as a consequence of the clockwise rotation of the southern end of the Fort Burgwin Horst which is pivoting around a point defined by the line formed by Fold A and Fault #20. Rotations in this block have formed a clockwise shear couple which, in turn, has caused counterclockwise, left-lateral, movements within the block itself. Similar but opposite movements have occurred along Fault #1.

The opening of the southeastern graben indicates that normal movements are probably concurrent with strike slip motions. It is probable that in general, with the exception of Faults #21a-d, normal movements have been propagating eastward in my field area. Strike-slip motions occurred generally later and are only minor features.

SUMMARY

This report has discussed over 1,800 million years of geologic history. In general, five major periods of deposition and deformation are important. These are briefly reviewed below.

(1) Precambrian: The Precambrian record begins with the deposition of the Ortega Group (ca 1,800 m.y.) - a series of metamorphosed sandstones and pelitic sediments - and the Vadito Group. The Embudo Granites were intruded from approximately 1,730 m.y. to 1,200 m.y. ago, during three metamorphic episodes and four deformational periods. The first recorded dextral strike-slip movement of the Pecos-Picuris Fault was 1,200 m.y. ago.

(2) Mississippian: The Mississippian was a time of tectonic quiescence dominated by shallow seas which deposited the sandstone and limestone of the Arroyo Penasco Formation.

(3) Pennsylvanian: Uplift of the Uncompahgre Highland to the west of the Pecos-Picuris Fault and subsidence in the adjacent Taos Trough resulted in the deposition of thick pulses of fan delta sediments from the highlands to the west into the trough to the east.

(4) Late Eocene: Due to east-west compression, the Laramide Orogeny uplifted the area to the west of the Pecos-Picuris Fault. Subsidiary thrust faults to the Pecos-Picuris Fault transported material eastward.

(5) Miocene and Younger: The Laramide uplift collapsed along old Pre-

cambrian and Laramide faults forming a central north-south depression through Colorado and New Mexico. The Rio Grande Graben formed approximately 26 m.y. ago. The Embudo Fault acts as a compressional transform fault offsetting the San Luis Basin from the Española Basin.

The mapping of the Fort Burgwin Ridge area has added data relevant to all aspects of the above history. These contributions are:

(1) The presentation of a detailed structural and geologic map of the Fort Burgwin Ridge and U.S. Hill area (see Plate 1).

(2) The drafting of twelve cross-sections of the area (see Plate 2).

(3) An examination of the Ortega rocks at U.S. Hill which points to similarities between these rocks and those of the Petaca District 23 miles to the northwest.

(4) On the basis of petrology anf field relations I suggest that the granite of Fort Burgwin Ridge is approximately 1,730 m.y. old.

(5) Syndepositional faulting during the Pennsylvanian along Faults #16 and #17 was observed. The fault style implies that uplift along the Pecos-Picuris Fault involved movement along parallel minor faults to the east which formed ramps into the Taos Trough (see Fig. 15).

(6) Many Laramide structures were discovered. Most notable of these is the upturned Mississippian-basement contact and the folded Pennsylvanian sediments to the east. A model involving a splay fault to the Pecos-Picuris Fault is proposed.

The splay fault (perhaps Fault #3) was to the west of Fort Burgwin Ridge. Drag along the fault curled the Mississippian rocks and formed folds in the Pennsylvanian rocks of the footwall (see Figs. 32 and 35). The examination of U.S. Hill suggests that the area directly east of the Pecos-Picuris Fault is a monocline which includes Fort Burgwin Ridge. The structure dies out southwards but reemerges as the Rio Pueblo Anticline (see Fig. 41).

(7) Indirect evidence has been presented which suggests that the Picuris Mountains are presently being rotated counterclockwise. The rotation explains such phenomena as the collapse of the Miranda Graben (see Fig. 40), the development of the southeast graben and an incipient graben to the northeast (see Fig. 45), and the occurrence of strike-slip movements along Faults #1, #6 and #7 (see Fig. 46).

The study of this area is incomplete. Among projects that would contribute important new data are:

(1) the granites to the east of the Pecos-Picuris Fault need to be dated.

(2) the Precambrian rocks on the east side of the Pecos-Picuris Fault (west side of Miranda Graben) should be examined to see how they relate to the granite of Fort Burgwin Ridge.

(3) the Pecos-Picuris Fault zone should be examined in detail from the Rio Grande Graben near Talpa south to Vadito.

(4) The Pennsylvanian sediments in Telephone Canyon, west of U.S. Hill and east of the Pecos-Picuris Fault should be mapped to determine the nature of the Pecos-Picuris Fault at that location.

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STRATIGRAPHY

			DESCRIPTION
	CENOZOIC		
	QUATERNARY		
Qal	Alluvium		Contemporary stream deposits
Qo	Outwash Gravel		glacial outwash terraces
QpЄo	Ortega Moraine		moraine or protalus deposits made exclusively of Ortega
	TERTIARY		
Tmt	Picuris Tuff	≈1,400′	miocene volcanic & fluvial deposits: coarse conglomerates; pink, white, green clay; water laid tuff; pink volcanic breccia.
	PALEOZOIC		
	PENNSYLVANIAN		
IPld	Late Desmoinesian	>1,000′	sandstone & limestone cycles which increase in limestone near the top.
IPmd	Middle Desmoinesian	≈2,200′	braided stream dominated alluvial fans coarse sandstone & conglomerate w/ shale intervals.
IPed	Early Desmoinesian	≈240′	Short sequence of shales capped by muddy limestone
IPa	Atokan	≈2,000′	coarse alluvial fan, braided stream & fan delta. Conglomerate, shale & muddy limestone.
IPm	Morrowan	≈800′	Muddy strand plain - coarsening upward cycles of coal, mud & shale to medium grained sandstone.
	MISSISSIPPIAN		
M	Arroyo Peñasco	≈130′	Carbonate rocks: blue-grey dedolomites and coarse crystalline limestone. Basal clastic rocks: honey yellow tabular sandstone.
	PRECAMBRIAN		
pЄg	Embudo Granite		Main granite: pink, foliated, porphyritic granite. Aplite dikes, amphibolite dikes, pegmatitic dikes & quartz veins.
pЄo	Ortega Metaquartzite		Blue grey coarse grained metaquartzite which increases in muscovite content near pegmatite dike & becomes fissile

Plate 3

PLATE 2

Cross-sections A-L refer to letters on plate 1.

EXPLANATION

GEOLOGY

QUATERNARY
Qal / Qg Alluvium / Outwash Gravels
QpЄo Moraine
TERTIARY
Tmt Miocene Picuris Tuffs

PENNSYLVANIAN
IPld Late Desmoinesian, ls+sh
IPmd Middle Desmoinesian, ss
IPed Early Desmoinesian, sh
IPa Atokan, fluvial sands
IPm Morrowan, shales

MISSISSIPPIAN
M Sandstone & lime stone

PRECAMBRIAN
PЄg Granites, Aplites, Pegmatites
PЄo Ortega Metaquartzites, Metaschists

STRUCTURE

horizontal strata
strike and dip of strata
strike and dip of joints
strike and dip of foliation
strike and plunge of lineation
strike and plunge of z fold
small syncline
Small anticline
H — H' Cross section line
8 fault ident. #
A fold ident. #
Roads
Stream
mine
inferred Anticline Syncline
large fold
inferred normal strike slip Thrust
U
D
large fault
contact
covered area
Boundary within IP

MAP OF THE GEOLOGY AND STRUCTURE BETWEEN TALPA AND RIO PUEBLO ALONG ROUTE 3 NEW MEXICO

Geology by Tom Chapin