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by William R. Seager and Greg H. Mack

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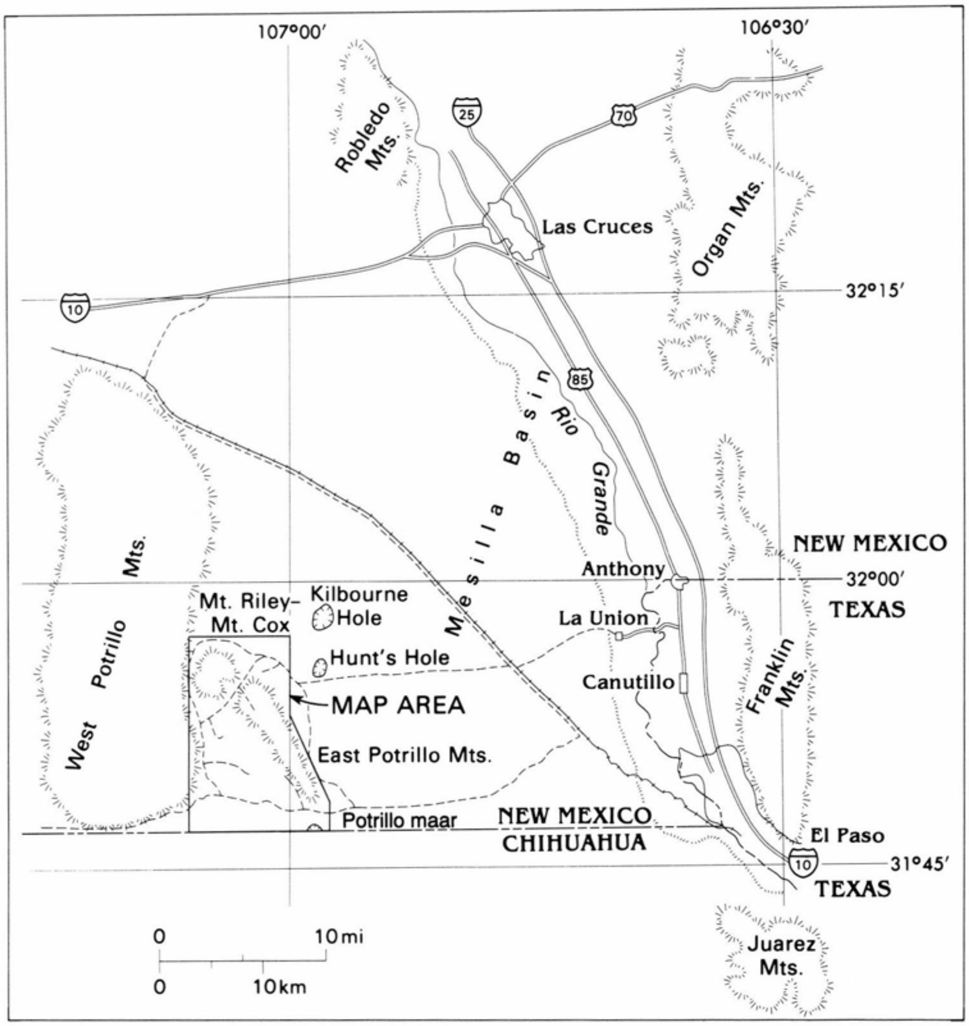

FIGURE 1—Map of East Potrillo Mountains, surrounding geologic and geographic features, and area included on geologic map (Sheet 1).

Abstract—Situated just north of the Mexico border in south-central Dona Ana County, New Mexico, the East Potrillo Mountains area reveals important details about Laramide, middle Tertiary, and late Tertiary deformation in the region, as well as data on Permian, Lower Cretaceous, and Cenozoic stratigraphy. The oldest rocks exposed in the area are limestone, dolomitic limestone, and minor silty or sandy beds, at least 300 m thick, that are believed to be correlative with the middle Permian Yeso and San Andres Formations of south-central New Mexico. Disconformably above these, Lower Cretaceous shelf limestone, siltstone, sandstone, and conglomerate are approximately 570 m thick in the northern part of the range, thinning southward. These rocks are Aptian-Albian in age and correlative with the Hell-to-Finish and U-Bar Formations of southwestern New Mexico. Orthoquartzite, possibly correlative with the Sarten or Mojado Formations, crops out locally. Cenozoic rocks are poorly exposed around the margins of the East Potrillo Mountains and on the flanks of the Mt. Riley-Mt. Cox volcanic domes. They include Upper Cretaceous and/or lower Tertiary conglomerate and sandstone (possible correlatives of the Love Ranch-McRae Formations), middle Tertiary intermediate-composition lavas and domes (correlative in part with the Rubio Peak-Palm Park Formations), and upper Oligocene or lower to middle Miocene fanglomerate. The latter rocks are thought to have been deposited in a broad "early-rift" basin extending 25 km or more to the west of the East Potrillo Mountains. Quaternary alluvial fans, basin-floor deposits, maar-rim tuffs, basalt flows, and cinder cones mantle older bedrock across much of the map area. Altogether approximately 1,350 m of Permian, Lower Cretaceous, and Cenozoic rocks are exposed.

Laramide, middle Tertiary, and late Tertiary structural features are exposed in the East Potrillo Mountains. Laramide thrust faults above overturned, thrust-faulted synclines trend N30°W ± 10°; drag folds indicate tectonic transport to the northeast. Erosion is not sufficiently deep to demonstrate whether thin-skinned overthrusting or rigid basement-block uplift has produced the observed shallow structure. Middle Tertiary deformation involved northeast or perhaps nearly north-south extension resulting in a system of closely spaced normal faults trending from N30°W to N60°W that cut Laramide structures. These faults presently dip from 0 to 50° (average approximately 30°) to the northeast or southwest. Original dips of major fault segments were as low as 20°-25° east and, this geometry as well as relationships in allochthonous and autochthonous rocks suggest that the structurally lowest parts of the system may have been initiated as a detachment fault zone. All faults in the system were rotated eventually approximately 25° southwestward, domino style, to their present attitude as middle Tertiary extension progressed. Late Tertiary uplift of the East Potrillo horst and subsidence of the adjoining grabens resulted from movement on modern, steep, range-boundary faults, trending from north-south to N20°W. Piedmont scarps along the range-boundary faults indicate continual movement into the late Quaternary. A symmetric positive gravity anomaly above the East Potrillo horst suggests that its uplift was not accompanied by significant rotation.

Introduction

Location, access, and physiography

The East Potrillo Mountains area is located in southwestern Doña Ana County, New Mexico, approximately 40 km west of El Paso and 55 km southwest of Las Cruces (Fig. 1). The Mexico-United States border forms the southern limit of the map, and the 107°06'30" meridian delineates the western edge of the map. Altogether approximately 235 km^2 of varied terrain—all starkly beautiful Chihuahuan Desert—were studied, and the results were incorporated into the map, cross sections, and text of this report.

The principal access roads are shown in Fig. 1; most are occasionally maintained. The roads that actually penetrate the mountains or hilly regions are unmaintained jeep trails. Most of the upland parts of the area are accessible only by foot or horseback.

The map area is geologically diverse, and this is reflected in a variety of landforms. The prominent volcanic domes of Mt. Riley and Mt. Cox dominate the view from most places within the area. These somewhat pyramid-shaped mountains ascend abruptly from the desert floor, towering 450 m or so above their surroundings. To the southeast, the narrow ridge known as the East Potrillo Mountains rises approximately 275-300 m above the bolson floor. In these mountains numerous transverse canyons with very steep or cliffy slopes and sparse desert vegetation provide superb exposures of sedimentary rocks and structural features. Basalt flows and cinder cones of the West Potrillo Mountains extend into the western margin of the study area. Although the relief here is low, the land is nevertheless rugged in its own way, owing to the youthfulness of the flows and to the uninhabited, desolate tracts of sand-strewn "malpais" (bad land). Between the mountains, ridges, and upland hills, broad dry piedmont slopes, striated by shallow arroyos, unify the landscape. Most of these broad flats are covered by a sheet of windblown sand or hummocky dunes anchored by mesquite or creosote. Countless fragments of white caliche strewn across the desert floor are a constant reminder of the thick carbonate-rich soil below the desert and of the geomorphic stability of these desert piedmont slopes and basin floors.

Previous work

Lee (1907) was the first to recognize Lower Cretaceous rocks in the East Potrillo Mountains. His collection of *Caprina occidentalis, Trigonia* sp. and *Octaeonella dolium* was identified by Stanton who assigned the collection to the Fredericksburg Group of the Comanche Series. Darton (1928, 1933) also observed the rocks of the East Potrillos and repeated the age assignments given by Lee (1907). Dunham (1935) recognized Permian and/or Pennsylvanian rocks in the northern part of the range and described a number of prospects and mineral occurrences. The first comprehensive study of the area was by Bowers (1960) who gave an account of the structure of the East Potrillo Mountains as well as the stratigraphy. He also recognized Permian rocks, which he assigned to the Hueco Formation, as well as Lower Cretaceous rocks, and he collected additional fossils to support the age assignments. He was the first to recognize the effects of compressional Laramide deformation in the range. Lokke (1964) described Lower Cretaceous *Orbitolina* from the East Potrillo Mountains.

Since 1960, Dr. J. Hoffer and his students at the University of Texas at El Paso (UTEP) have studied various aspects of igneous rocks in the map area, and they have been cited where appropriate in this text. More recently, Dr. R. Dyer, also of UTEP, and his students have mapped southernmost parts of the East Potrillo Mountains (Powell, 1983). Dr. H. Drewes, USGS, also is currently (1984) studying Laramide deformation in the range.

Present study

Our study of the East Potrillo area began in 1982 as part of the mapping for the Las Cruces and El Paso 1° x 2° sheets by Seager. Because the East Potrillo Mountains offered an excellent opportunity to study the style of Laramide deformation as well as the character of low-angle normal faulting, we decided to map the area in more detail than would be necessary for a 1° x 2° sheet. Detailed mapping began in the spring of 1983, continuing, when time was available, through the rest of 1983. Mapping was completed in May 1984. Stratigraphic studies by Mack were also completed in the summer of 1984.

Acknowledgments

The New Mexico Bureau of Mines and Mineral Resources financially supported this project. We would like especially to thank Dr. Frank E. Kottlowski for his encouragement and support of this project and Dr. Charles E. Chapin, Director of the New Mexico Bureau of Mines and Mineral Resources, for continuing to support our geological investigations in southern New Mexico. Richard Chamberlin, Ray Ingersoll, and Sam Thompson III improved the manuscript substantially with their careful, thoughtful reviews, and we thank them for their considerable effort. Discussions with Harald Drewes were helpful in understanding the geology of the East Potrillo Mountains. We thank W. A. Cobban for re-examining Bower's faunal list and Raymond C. Douglass for identifying *Orbitolina* for us. Discussions with David LeMone on Cretaceous stratigraphy and Calvin James on Cretaceous depositional environments were most helpful.

Stratigraphy

In the East Potrillo Mountains and adjacent areas, approximately 1,300 m of sedimentary and volcanic rock and alluvium, which range in age from middle Permian to Holocene, are exposed (Fig. 2). Permian and Lower Cretaceous sedimentary rocks, at least 300 m and 570 m thick, respectively, crop out extensively in the East Potrillo Mountains. Upper Cretaceous or lower Tertiary rocks correlative with the McRae and/or Love Ranch Formations of Las Cruces and Elephant Butte areas are poorly exposed around the perimeter of Mt. Riley and Mt. Cox. Middle Tertiary intermediate-composition volcanic rocks, largely flows and laharic breccia, are correlative in part with the Rubio Peak and Palm Park Formations, but again outcrops are limited. Extensive but poorly exposed outcrops of upper Oligocene(?)-middle Miocene fanglomerate assigned to the lower Santa Fe Group probably represent the fill of an "early-rift" basin. At least 300 m of fanglomerate are present. Quaternary deposits include alluvial-fan and basin-floor sediments, widespread basalt flows and cinder cones, and rim deposits of a maar volcano. Intrusive igneous rocks in the region consist of middle Tertiary dikes and sills of intermediate composition and the andesite-dadte volcanic domes of Mt. Riley-Mt. Cox.

Sedimentary rocks

Permian rocks

The oldest exposed stratigraphic unit (P) in the East Potrillo Mountains consists primarily of carbonate rocks and is believed to be middle Permian. Owing to faulting and other structural complications the exposed thickness of the unit is uncertain; it is clearly at least 300 m thick and may be 500 m thick.

Light-, medium-, and dark-gray, medium-bedded dolomitic limestones make up the lower two-thirds of the map unit. The dolomitic limestones locally exhibit silty laminations, small spar-filled vugs similar to birdseye structures, and thin horizontal to wavy laminations, which resemble stromatolites. Throughout the section limestone breccia, probably solution breccia, occurs sporadically in beds as thick as 1 m, as well as in thicker, more localized breccia masses. A few indurated brown siltstone beds approximately 1 m thick are present, and in the core of an anticline in the northwestern part of the range, yellow-brown, soft, fine-grained sandstone of typical Yeso or San Andres lithology crops out. No fossils were observed in the lower two-thirds of the map unit.

The upper third of the section consists of massive to medium-bedded, gray limestone, which contains chert nodules in the upper part. These rocks are marbleized in the middle and southern parts of the range. In thin section, the limestones are predominantly echinoderm wackestones that are partially recrystallized to neospar. An intraclast peloid, foraminifera, brachiopod, pelecypod packstone and an oolite grainstone are found near the bottom of the upper third. In the northern part of the range, echinoderm columnals, high-spired gastropods, pelecypods, and rare, small brachiopods and horn corals are locally visible in hand specimens.

Dunham (1935) believed these carbonate rocks corresponded to either the San Andres Formation or Hueco Formation of middle to Early Permian age, respectively. Bowers (1960), Hoffer and Hoffer (1981), and Powell (1983) assigned them to the Hueco Formation. We believe the carbonates are correlative with the Yeso and San Andres Formations in south-central New Mexico (Kelley and Silver, 1952; Kottlowski et al., 1956; Seager, 1981) and with the Colina and Epitaph Formations of southwestern New Mexico and southeastern Arizona (Zeller, 1965).

The Permian carbonate rocks in the East Potrillo Mountains differ from the Hueco Formation in several important ways. In the Robledo Mountains, in outcrops near Camel Mountain 30 km west of the East Potrillo Mountains, and in the Florida and Tres Hermanas Mountains, the Hueco Formation consists of a series of interbedded dark-gray limestone, shaly limestone, and shale. Fossils that are conspicuous at all of these localities include planispiral gastropods, phylloid algae, echinoid spines, productid brachiopods, and razor clams. In contrast, the carbonates in the East Potrillo Mountains contain much dolomitic limestone with solution breccias, occasional soft, yellow to tan sandstone beds, and little or no shale or argillaceous limestone. Furthermore, horn corals, not observed in the Hueco Formation, are present in the East Potrillo section, as are coquinas of equidimensional pelecypods with strongly convex shells

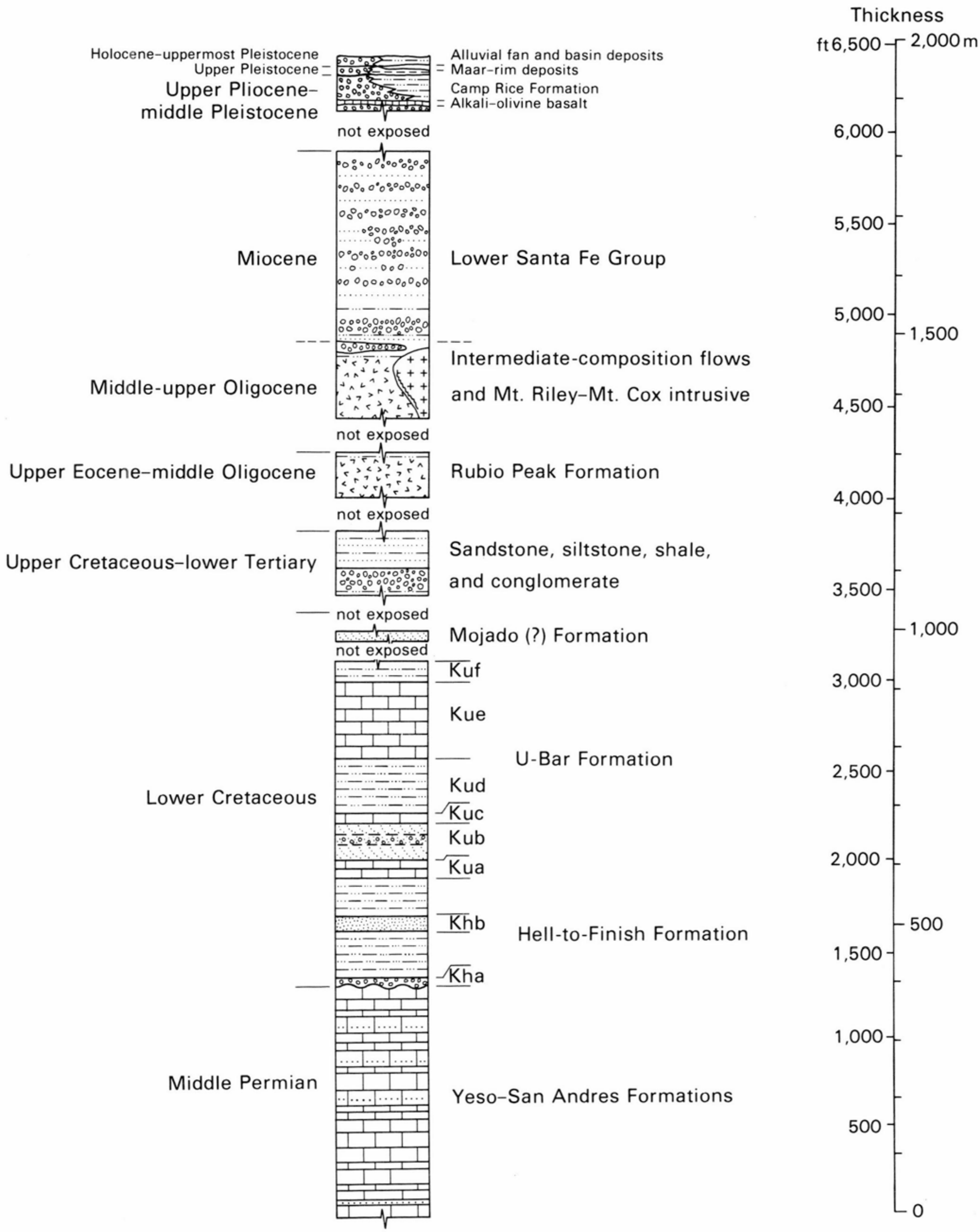

FIGURE 2—Generalized columnar stratigraphic section of mapped rock units in the East Potrillo Mountains and vicinity.

and curved beaks. Only rare, small nonproductid-type brachiopods and very few planispiral gastropods were observed in the East Potrillo section. Local coquinas of high-spired gastropods are unlike any known to us from the Hueco Formation. No algae of any type was noted. Based on these differences and on the close lithologic similarity to Yeso and San Andres carbonates in the Caballo and San Andres Mountains, we believe that the East Potrillo basal carbonate unit is middle Permian. The lower two-thirds of the unit is probably correlative with the Yeso Formation and the upper one-third with the San Andres Formation. Because the correlation is based only on lithology and on an overall faunal aspect that appears different from the Hueco Formation, we are currently collecting conodonts with the hope that they will provide an unequivocal age assignment.

Lower Cretaceous stratigraphy

Lower Cretaceous rocks in the East Potrillo Mountains were first described by Darton (1928), who correlated the East Potrillo section with the Fredericksburg Group of the Gulf Coast. The first rigorous stratigraphic work in the East Potrillo Mountains was done by Bowers (1960), who divided the Lower Cretaceous section into three formations, the basal Noria, Little Horse, and Restless. Fossil collections from the Little Horse and Restless formations suggested a correlation with the Trinity and/or Fredericksburg Groups of the Texas Gulf Coast (Bowers, 1960). Subsequent mapping in the East Potrillo Mountains followed or slightly modified the stratigraphic scheme of Bowers (Hoffer, 1976; Powell, 1983).

In the present study the criteria for selection of stratigraphic units are twofold: 1) to select stratigraphic units that conform to standard nomenclature in south-central and southwestern New Mexico, thereby simplifying regional correlation, and 2) to select mappable units that most clearly define the complex structure of the range. The stratigraphic subdivisions of Bowers (1960) fail to meet either criterion. The formation names of Bowers (1960) are used only in the East Potrillo Mountains and should be abandoned in favor of a more widely used terminology. Furthermore, the formations of Bowers (1960) do not provide the level of detail required for structural mapping of the range. Bowers (1960) also did not describe the uppermost two stratigraphic units recognized in this report.

The standard Lower Cretaceous section in southwestern New Mexico is in the Big Hatchet Mountains, where the section is divided into three conformable units, which are, in ascending order, the Hell-to-Finish, U-Bar, and Mojado Formations (Zeller, 1965). The lithologic characteristics of the Lower Cretaceous section in the Big Hatchet Mountains can be recognized also in the East Potrillo Mountains, justifying extension of the names Hell-to-Finish, U-Bar, and Mojado into the East Potrillo Mountains (Sheet 3). However, in order to provide the necessary map units the Hellto-Finish and U-Bar Formations in the East Potrillo Mountains are divided into informal members that cannot be directly correlated with members of the Hell-to-Finish and U-Bar Formations in the Big Hatchet Mountains. The Hellto-Finish Formation in the East Potrillo Mountains is divided into a conglomerate member (Kha) and an overlying mottled siltstone member (Khb). The U-Bar Formation in the East Potrillo Mountains is separated into six members, which are, in ascending order, the lower limestone (Kua), sandstone (Kub), rudistid limestone (Kuc), siltstone-limestone (Kud), massive limestone (Kue), and upper siltstone (Kuf) members. The Mojado(?) Formation (Km?) is not divided into members in this report. A more detailed discussion of regional correlations follows the descriptions of the stratigraphic units.

Hell-to-Finish Formation—The lower member of the Hell-to-Finish Formation (Kha) in the East Potrillo Mountains is a gray, cobble and boulder conglomerate, which weathers to rounded slopes and low ridges (Fig. 3). In the northern part of the range 39 m were measured, whereas in the southern section only 1 m of basal conglomerate was measured. The conglomerate is grain supported and has a well-sorted, medium- to coarse-grained sandstone matrix. Clasts consist of moderately well to well-rounded limestone and dolomite, similar to the underlying Permian carbonates, and subangular to rounded chert. Thin, medium- to coarse-grained sandstone interbeds are rare. The conglomerate member of the Hell-to-Finish Formation lies unconformably on Permian carbonate rocks but appears to be gradational and conformable with the overlying mottled siltstone member of the Hell-to-Finish Formation. Measurements of the lengths of the long axes of the 10 largest clasts in 3-m intervals in the northern section reveal a general upward-coarsening sequence to the 27-m level, followed by an upward-fining sequence to the top of the member (Fig. 4).

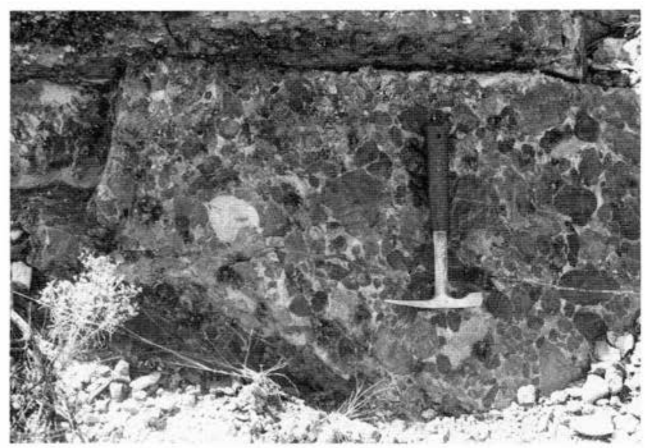

FIGURE 3—Lower Cretaceous conglomerate member of the Hell-to-Finish Formation (Kha) composed of well-rounded limestone and chert clasts. Hammer is 25 cm long.

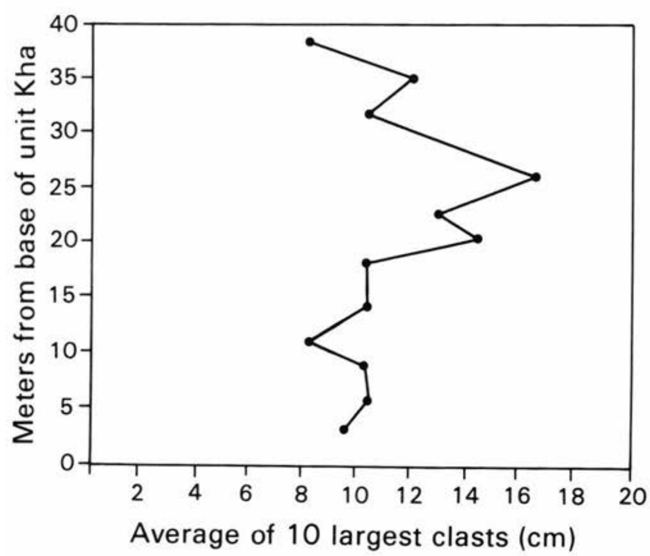

FIGURE 4—Clast size versus stratigraphic position in Lower Cretaceous conglomerate member of the Hell-to-Finish Formation (Kha) in the northern measured section. Size data represent the average of the 10 largest clasts in each 3-m interval.

The mottled siltstone member of the Hell-to-Finish Formation (Khb)is a medium- to thick-bedded, ledge-forming unit with a moderate amount of poorly exposed or covered slopes. The rock superficially resembles limestone and has been described as limestone (Bowers, 1960; Hoffer and Hoffer, 1983), but it is predominantly calcareous siltstone and sandstone. These rocks have a characteristic gray and tan-mottled appearance. The mottling is surficial, however, and the rock is gray on a fresh surface. In the northern part of the range, the mottled siltstone member is 168 m thick, whereas in the southern part of the range a thickness of 119 m was measured. In the northern section, the contact between the conglomerate and mottled siltstone member is covered, but in the southern section the contact appears to be conformable, with the conglomerate member grading up into fine sandstone of the mottled siltstone member. A conglomerate similar to those in the conglomerate member can be found approximately 10 m from the base of the mottled siltstone member. In both sections, the contact between the mottled siltstone member and the basal member of the U-Bar Formation is very sharp and conformable.

The most common lithologies in the mottled siltstone member are calcareous siltstone and very fine and fine-grained sandstone, with siltstone more abundant in the southern section than in the northern section. The siltstones and sandstones are massive and have horizontal laminations or irregular wavy and low-angle laminations, which resemble hummocky cross-stratification of Harms et al. (1975; Fig. 5). Horizontal and vertical burrows are also common, and there are a few examples of wave-oscillation ripples, pelecypods, and gastropods. The sandstones are quartzarenites, chertarenites, and subarkoses (Table 1).

Thin conglomerates are scattered throughout the mottled siltstone member. These conglomerates are generally less than or equal to 0.5 m thick, have an average maximum clast size of approximately 3 cm, and exhibit normal grading. Some conglomerates are composed of pebbles of detrital limestohe and chert, but most consist of intraformational rip-up clasts of siltstone and sandstone. The conglomerates have sharp basal contacts and gradational upper contacts.

In roughly the middle of both the northern and southern sections is a 6-m-thick, granular, coarse sandstone. This sandstone contains trough crossbeds and pelecypod and gastropod shell debris.

Dark-gray or green shale is exposed rarely but probably occupies many of the covered intervals. Locally the gray shale weathers to white, powdery shale. In the lower half of the northern section, there is a 5-m-thick bed of unfossiliferous, silty micrite.

U-Bar Formation—The lower limestone member of the U-Bar Formation (Kua) is a prominent, massive to thick-bedded, ledge-forming, gray limestone with conformable upper and lower contacts. In the northern part of the range, the lower limestone member is approximately 30 m thick, but it thins to only 13 m in the southern part of the range. In the northern section, the basal 5 m of the lower limestone member consist of foraminifera, pelecypod, peloid, and intraclast packstones. The foraminifera were identified by Raymond Douglass as *Orbitolina gracilis.* The upper 0.5 m of this interval has oncolites. The bulk of the unit, the middle 23 m, is a thick-bedded, sandy oolite grainstone. Sandy gastropod, pelecypod, intraclast packstone comprises the upper 2 m. In the southern section, the lower limestone member is composed of partially recrystallized pelecypod, gastropod wackestone and grainstones, some of which are silty. No oolite beds were recognized in the southern section.

The sandstone member of the U-Bar Formation (Kub) consists of ledge-forming sandstone, which has a diagnostic dark-brown color in weathered outcrops. The sandstone member has sharp and conformable upper and lower contacts. Locally the lower contact is a scoured surface 0.5 m deep overlain by a 3-m-thick set of low-angle crossbeds. Elsewhere the base of the sandstone member has a 10-cm-thick pebble layer. In the northern section the sandstone member is 66 m thick, but it thins southward to 25 m. The sandstone member has two lithologic groups: granular, pebbly sandstone and very fine, fine-, and medium-grained sandstone. The granular, pebbly sandstones are commonly crossbedded and contain pelecypod shell debris (Fig. 6). One crossbedded granular, pebbly sandstone also contains oolites. The finer sandstones are massive, horizontally laminated, or hummocky cross-stratified and occasionally exhibit burrows and pelecypod shells. The sandstones are arkoses and subarkoses (Table 1). In the northern measured section, the sandstone member has four beds of crossbedded granular, pebbly sandstone, ranging in thickness from 7 to 12 m. The southern section also has four granular, pebbly sandstones, but they range from 1.5 to 4 m in thickness. Less common lithologies include thin pebble conglomerate and a thin silty micrite in the northern section.

The rudistid limestone member of the U-Bar Formation (Kuc) is a relatively thin, although easily recognizable, ledge-forming, gray limestone. It has sharp, conformable contacts and varies in thickness from 13 m in the northern section to 6 m in the southern section. The northern section has a 1-m-thick basal sandy oolite grainstone that is absent in the southern section. The lower half of both the northern and

FIGURE 5—Hummocky cross-stratification in Lower Cretaceous mottled siltstone member of the Hell-to-Finish Formation (Khb). Hammer is 25 cm long.

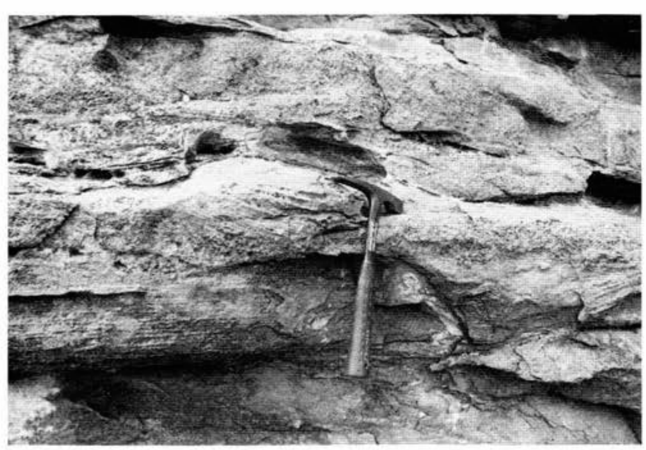

FIGURE 6—Medium-scale crossbeds in Lower Cretaceous granular coarse sandstone member of the U–Bar Formation (Kub). Hammer is 25 cm long.

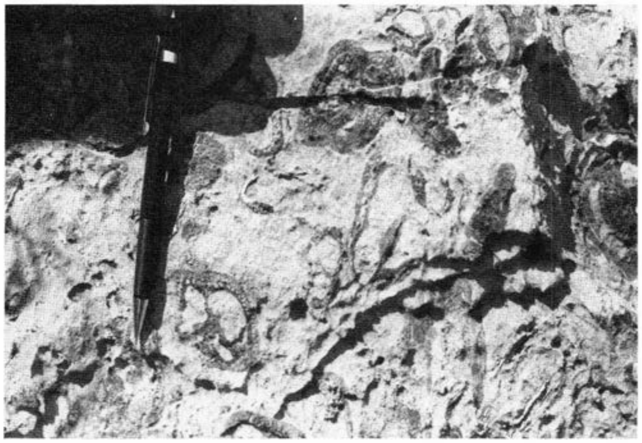

FIGURE 7—Rudistid pelecypods in rudistid limestone member of the U–Bar Formation (Kuc). Pen is 10 cm long.

southern sections is a massive silty rudistid boundstone, which, in thin section, can be seen to contain intraclasts, a few oolites, and the foraminifera *Orbitolina grossa* (R. C. Douglass, pers. comm. 1985; Fig. 7). Locally this limestone is recrystallized. The rudistid boundstone is overlain by medium-bedded turritellid gastropod and pelecypod wackestones. The fossils are easily recognizable in outcrop.

The siltstone-limestone member of the U-Bar Formation (Kud) consists of ledge-forming, gray and tan, calcareous siltstone and sandstone, with minor amounts of gray limestone and shale. A complete section, 90 m thick, was measured in the southern part of the range, where the upper and lower contacts are conformable. In the northern part of the range, the siltstone-limestone member has a conformable base, but the upper contact is not exposed; 81 m were measured in the northern section. The northern section has a higher sandstone/siltstone ratio (1.3 versus 0.09) and a higher limestone/sandstone and siltstone ratio (0.5 versus 0.04) than the southern section.

The dominant lithologies in both sections are calcareous siltstone and very fine and fine-grained sandstone. Both lithologies are massive, horizontally laminated, or hummocky cross-stratified. Fine fossil hash, megascopic gastropod and pelecypod shells, and burrowing are common. A few excellent examples of symmetrical ripples also can be found.

In the northern section, limestone beds vary in thickness from 0.2 to 5 m. Four different types of limestone were recognized in thin section. The most common is a gastropod, pelecypod packstone or wackestone, which may contain peloids and detrital silt. The second-most abundant limestone is a silty micrite with scattered gastropods and pelecypods. One bed of silty peloid, gastropod, pelecypod grainstone and one bed of silty peloid, oolite, gastropod, pelecypod packstone also were observed. The southern section has only a few gastropod, pelecypod packstones.

Shale was observed also in both sections, but it is rare. In the northern section the shale is a poorly exposed, white, powdery mudstone. In the southern section the shale is dark gray or green and very dense.

The massive limestone member of the U-Bar Formation (Kue) is exposed only in the southern part of the range, where it is 133 m thick and has conformable upper and lower contacts. The unit is characterized by massive to thick-bedded, ledge-forming, gray limestone. Limestones in the basal 20 m are silty and have a few gastropod and pelecypod fossils. Limestones in the remainder of the section have been recrystallized to coarse neospar and are crosscut by numerous calcite veins. No relict fossils were recognized in the upper limestone in either hand specimen or thin section. Two thin (<1 m), dark-gray to green, dense mudstones also exist in the massive limestone member.

The upper siltstone member of the U-Bar Formation (Kuf) is exposed only in the southern part of the range. The base is conformable with the massive limestone member, but the top is not exposed. Only 15 m of the upper siltstone member are exposed. The dominant lithology is gray and tan-mottled, calcareous siltstone, which is hummocky cross-stratified and bioturbated. A thin, recrystallized limestone was also found.

Mojado(?) Formation—Sandstones tentatively assigned to the Lower Cretaceous Mojado(?) Formation (Km?) are exposed in three low hills on the northwest side of the East Potrillo Mountains. The sandstone is a light-gray, massive, partially silicified, crossbedded quartzarenite (Table 1). Approximately 15 m of this formation are exposed. No fossils were found in this unit, and it cannot be traced laterally or vertically into other Lower Cretaceous strata. Consequently, the assignment of these rocks to the Mojado(?) Formation is tenuous.

Age and regional correlation—Paleontologic evidence for the age of the Cretaceous rocks in the East Potrillo Mountains is scanty, primarily because of poor fossil preservation and recrystallization of some limestone units. Despite the paucity of fossils, enough biostratigraphic control remains to allow assignment of the rocks, with a moderate degree of confidence, to the Lower Cretaceous Series and to the upper Aptian and/or lower Albian Stages (Craig, 1972; Hoffer and Hoffer, 1983; Powell, 1983). The age assignments are strengthened by lithologic correlation with fossiliferous Lower Cretaceous rocks in southwestern New Mexico.

The first attempt at age correlation in the East Potrillo Mountains was by Darton (1928), who cited a fossil collection of W. T. Lee that was examined by T. W. Stanton. This assemblage, which consisted of fossils of only four different species, was correlated with the Fredericksburg Group of the Texas Gulf Coast or middle Albian in terms of European stages. A larger fossil collection (13 species) made by Bowers (1960) was correlated with the Trinity Group of the Texas Gulf Coast, although a Fredericksburg age for the uppermost fossiliferous unit was considered possible. Reevaluation of the faunal list of Bowers (1960) by W. A. Cobban (pers. comm. 1985) indicates that many of the species are long ranging and of little use as index fossils. However, the gastropod *Nerinea hicoriensis* is found in Aptian rocks near Austin, Texas, and the coral *Pleuocora texana* is known in the middle Albian (W. A. Cobban, pers. comm. 1985).

The most important index fossil to date in the East Potrillo Mountains is the foraminifera *Orbitolina.* The taxonomy and biostratigraphy of *Orbitolina* in North America was delineated by Douglass (1960), who examined sections of Lower Cretaceous limestones in Texas, southeastern Arizona, and southwestern New Mexico exclusive of the East Potrillo Mountains. Douglass (1960) defined eight species of *Orbitolina* and demonstrated that they ranged from upper Aptian to lower Albian, upper Trinity in Gulf Coast terminology. Following Douglass' work, *Orbitolina* was found in the East Potrillo Mountains by several workers, including one of us (Mack) during the present study. Lokke (1964) identified *Orbitolina gracilis,* but the stratigraphic unit from which the fossils were taken is not clear from his description. Craig (1972) identified the species *Orbitolina texana* (Roemer). Thin sections of *Orbitolina-bearing* limestones collected during this study were sent to R. C. Douglass for identification. He recognized *Orbitolina gracilis* from the lower limestone member of the U-Bar Formation and *Orbitolina grossa* from the rudistid limestone member of the U-Bar Formation. The presence of *Orbitolina* suggests an early Albian age for at least the lower three members of the U-Bar Formation.

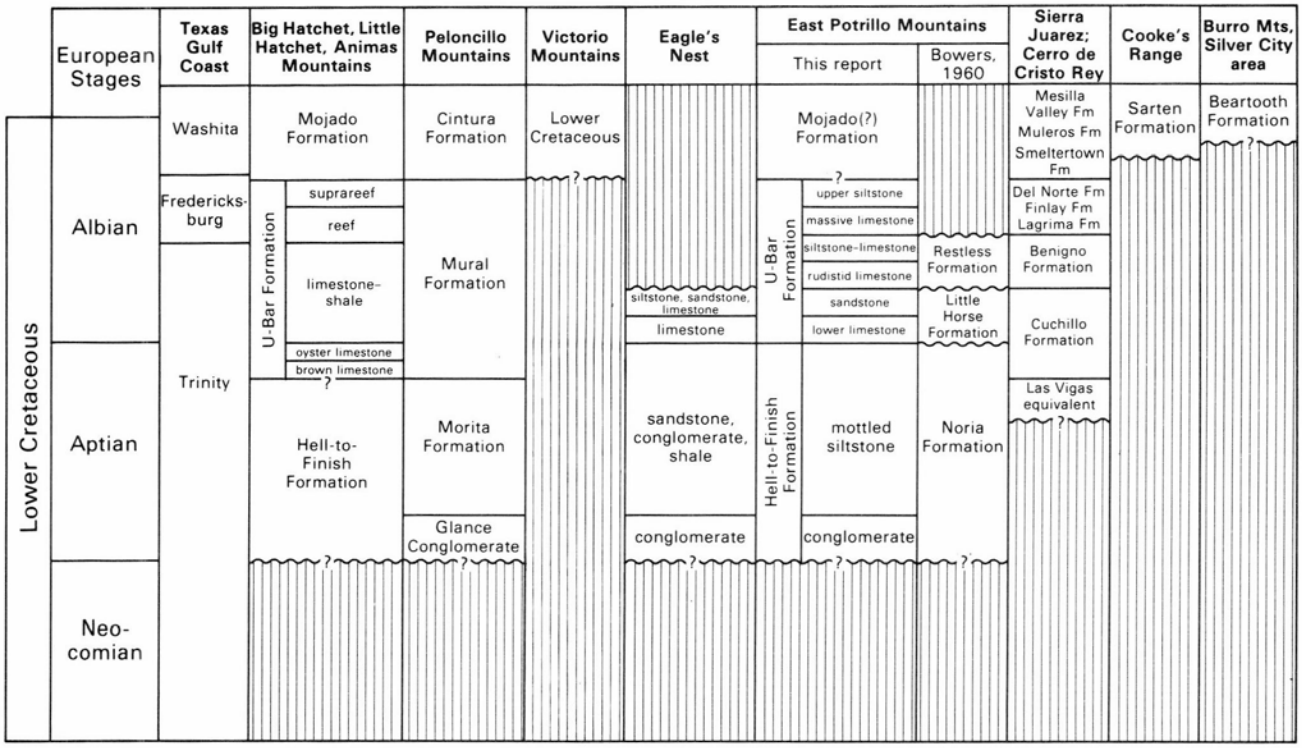

FIGURE 8—Correlation chart for Lower Cretaceous sedimentary rocks in southwestern and south-central New Mexico and Sierra Juarez and Cerro de Cristo Rey, Mexico. Adapted from Mack et al. (1986). For Peloncillo Mountains, terminology is that of Drewes and Thorman (1980a, b).

Regional correlation of the upper three members of the U—Bar Formation, as well as the Hell-to-Finish and Mojado Formations, is based on lithology and stratigraphic position.

In the Big Hatchet, Little Hatchet, and Animas Mountains of southwestern New Mexico, Lower Cretaceous rocks are divided into three conformable formations, which are, in ascending order, the Hell-to-Finish, U—Bar, and Mojado Formations (Zeller, 1965, 1970; Zeller and Alper, 1965).

The basal Hell-to-Finish Formation is approximately 480 m thick in the Big Hatchet Mountains, at least 700 m thick in the Little Hatchet Mountains, and 145 m thick in the Animas Mountains. The Hell-to-Finish Formation is composed of sandstone, shale, and conglomerate, with a minor amount of limestone. Where the base is exposed, the Hellto-Finish Formation has a diagnostic limestone- and chertcobble conglomerate, which can be as much as 30 m thick.

The U—Bar Formation is a mixed siliciclastic—carbonate unit 1,230 m thick, which was divided by Zeller (1965) in the Big Hatchet Mountains into five members. The lower brown limestone member consists of limestone, shale, and sandstone and is overlain by fossiliferous limestone, shale, and a few sandstones of the oyster limestone member. The third member, the limestone—shale member, is overlain by the cliff-forming reef limestone member. The uppermost member is the suprareef member, which is composed of medium-bedded limestone.

The Mojado Formation is 1,584 m thick in the Big Hatchet Mountains and consists of sandstone and shale, although a few beds of limestone occur in the lower and upper parts (Zeller, 1965).

Fossils in the oyster limestone member and the limestone—shale member of the U—Bar Formation in the Big Hatchet Mountains correlate with fossils in the Trinity Group (late Aptian and early Albian) of the Texas Gulf Coast (Zeller, 1965; Weise, 1982). The Aptian—Albian boundary is considered to be at the base of the limestone—shale member (Fig. 8; Zeller, 1965; Weise, 1982). Fossils in the reef and suprareef members correlate with those in the Fredericksburg Group of Texas, which is middle Albian in terms of European stages (Zeller, 1965; Weise, 1982). Fossils in the upper part of the Mojado Formation correlate with fossils of the Washita Group of Texas and indicate a late Albian and early Cenomanian age (Zeller, 1965). No index fossils are found in the Hellto-Finish Formation, but it is assumed to be Aptian or older because of its conformable contact with the U—Bar Formation (Fig. 8).

The Lower Cretaceous section in the East Potrillo Mountains can be correlated with Lower Cretaceous rocks in southwestern and south-central New Mexico (Mack et al., 1986). The basal two members in the East Potrillo Mountains correlate lithologically with the Hell-to-Finish Formation of southwestern New Mexico (Fig. 8). The conglomerate member closely resembles the basal limestone- and chert-cobble conglomerate of the Hell-to-Finish in the Big Hatchet and Animas Mountains. The mottled siltstone member correlates lithologically with the Hell-to-Finish Formation and perhaps with the lower two members of the U—Bar Formation in the Big Hatchet Mountains. The lower limestone, sandstone, and rudistid limestone members in the East Potrillo Mountains are similar in lithology to the brown limestone, oyster limestone, and limestone—shale members of the U—Bar Formation in the Big Hatchet Mountains, and the lower limestone and rudistid limestone members have a lower Albian fauna (Fig. 8). No index fossils were found in the siltstone—limestone, massive limestone, or upper silt-stone members in the East Potrillo Mountains, but these members are lithologically similar to the reef and suprareef members of the U—Bar Formation in the Big Hatchet Mountains (Fig. 8). The Mojado(?) Formation in the East Potrillo Mountains is lithologically similar to sandstones in the Mojado Formation in southwestern New Mexico, although the

TABLE 1—Detrital modes of Lower Cretaceous sandstones and one sandy oolitic limestone (EP–26). Data represent 300 points per thin section, exclusive of cement and matrix. Thin sections were stained with sodium cobaltinitrate to aid in feldspar identification. **MXQ** = monocrystalline quartz, **PXQ** = polycrystalline quartz, **KSP** = potassium feldspar, **PLG** = plagioclase feldspar, **CHT** = chert, **PRF** = pelitic rock fragment, **MRF** = metamorphic rock fragment, **VRF** = volcanic rock fragment, **QFF** = quartzofeldspathic rock fragment, and **ACC** = accessory minerals (biotite, amphibole, epidote, etc.).

	Lower Cretaceous unit	Sample	MXQ	PXQ	KSP	PLG	CHT	PRF	MRF	VRF	QFF	ACC
Hell-to-Finish Formation	Mottled siltstone member	EP–17	82.0	4.3	0.0	0.0	12.7	1.0	0.0	0.0	0.0	0.0
		EP–18	71.0	3.7	4.3	17.7	3.0	0.0	0.0	0.0	0.0	0.3
		EP–19	84.0	3.3	0.0	1.0	10.7	1.0	0.0	0.0	0.0	0.0
		EP–20	95.7	2.7	0.3	0.3	0.7	0.3	0.0	0.0	2.7	0.0
U–Bar Formation	Sandstone member	EP–33	64.3	4.7	13.7	10.7	3.3	0.7	0.0	0.0	2.7	0.0
		EP–34	82.6	1.0	9.7	4.0	2.3	0.3	0.0	0.0	0.0	0.0
		EP–35	66.0	2.0	18.0	12.3	0.7	0.0	0.0	0.0	1.0	0.0
		EP–36	74.0	5.0	7.3	6.3	6.7	0.7	0.0	0.0	0.0	0.0
	Rudistid limestone member	EP–26	93.0	2.7	1.3	2.0	0.7	0.3	0.0	0.0	0.0	0.0
Mojado Formation		EP–58	93.3	3.3	0.0	0.0	2.7	0.7	0.0	0.0	0.0	0.0

Mojado(?) Formation in the East Potrillo Mountains contains no index fossils and is poorly exposed.

In the Peloncillo Mountains of southwestern New Mexico, Gillerman (1958) mapped four Lower Cretaceous formations, the McGhee Peak, Carbonate Hill, Still Ridge, and Johnny Bull. Armstrong et al, (1978) used Gillerman's map units in their map of the central Peloncillo Mountains, except that they combined the Still Ridge and Johnny Bull Formations into one map unit. The McGhee Peak Formation correlates lithologically with the Hell-to-Finish Formation in the Big Hatchet, Little Hatchet, Animas, and East Potrillo Mountains (Fig. 8). The Carbonate Hill Formation is lithologically similar to the U-Bar Formation, although thinner, and contains an upper Aptian fauna (Gillerman, 1958). The Still Ridge and Johnny Bull Formations are lithologically equivalent to the Mojado Formation. Drewes and Thorman (1980a, b) extended Arizona stratigraphic terminology into the Peloncillo Mountains, applying the names Glance and Morita for the Hell-to-Finish, Mural for the U-Bar, and Cintura for the Mojado (Fig. 8).

In south-central New Mexico, Lower Cretaceous sections are incomplete and poorly exposed. At Eagle's Nest on the west side of the West Potrillo Mountains, approximately 250 m of Lower Cretaceous strata are overturned beneath a thrust fault whose hanging-wall block is cliff-forming Paleozoic limestone. The Lower Cretaceous section consists of four informal members, which in ascending order are 1) basal limestone- and chert-cobble conglomerate that is 15 m thick; 2) 70 m of poorly exposed arkosic sandstone, pebble conglomerate, and shale; 3) 27 m of rudistid boundstone, fossiliferous packstone, and oolitic grainstone; and 4) 137 m of siltstone, fine sandstone, and fossiliferous wackestone and packstone. The basal conglomerate is cut by the thrust fault, and the top of the stratigraphically highest member is not exposed. Hoffer and Hoffer (1983) collected the foraminifera *Dictyoconus walnutensis* from the middle limestone member and consider it to be a Fredericksburg index fossil. However, Wacker (1972) found the same fossil in the Juarez Mountains in the Benigno Formation, which has a Trinity fauna (Cordoba, 1969). Smith (1970) also found *Dictyoconus walnutensis* in the Trinity Benigno Formation in northern Coahuila, Mexico. Thus, the range of this fossil is not restricted to Fredericksburg but rather extends down into the Trinity. *Orbitolina* was also found by Hoffer and Hoffer (1983) in the middle limestone unit, suggesting a late Aptian or early Albian (Trinity) age (Douglass, 1960). A combination of lithologic trends and meager fossil record suggests that the basal conglomerate and overlying sandstone, conglomerate, and shale members probably correlate with the Hellto-Finish Formation in the East Potrillo Mountains. The upper two members at Eagle's Nest probably are equivalent to the middle limestone and sandstone members of the U-Bar Formation in the East Potrillo Mountains.

In the Victorio Mountains of south-central New Mexico, approximately 235 m of Lower Cretaceous rocks lie unconformably above the Silurian Fusselman Dolomite and are overlain unconformably by Tertiary conglomerates and sandstones. The Cretaceous section consists primarily of siltstones and fine sandstones, with lesser amounts of conglomerate and thin limestone. Conglomerate is most common in the lower half of the section. A 15-m-thick fossiliferous zone appears approximately 81 m above the base of the section. F. E. Kottlowski collected the pelecypod *Trigonia emoryi* from the Victorio Mountains section, indicating a late Albian age (Griswold, 1961). Thus, the Victorio Mountains section is tentatively correlated with the Mojado Formation (Fig. 8).

In the Tres Hermanas Mountains of Luna County, Kottlowski and Foster (1962) described a 467-m-thick section of sedimentary rocks that they correlated with Lower Cretaceous rocks of southwestern and south-central New Mexico. This section was subdivided into five parts: 1) basal clastic rocks consisting of 114 m of conglomerate, siltstone, and silty unfossiliferous limestone; 2) massive recrystallized limestone 120 m thick; 3) light-gray, crossbedded sandstone 12 m thick; 4) limestone- and chert-pebble and cobble conglomerate with interbedded sandstone 118 m thick; and 5) sparsely fossiliferous limestone and dolomite 103 m thick. Kottlowski and Foster (1962) speculated that a fault may exist between units 1 and 2 and that the top of the section is a fault contact. Although a Lower Cretaceous designation for the rocks in the Tres Hermanas Mountains has persisted (Hoffer and Hoffer, 1983), recent information raises serious doubts about the stratigraphic relationships. W. R. Seager and R. E. Clemons, working independently, have recognized a thrust fault between the upper limestone and dolomite (unit 5) and the middle conglomerate (unit 4). Furthermore, the upper limestone and dolomite resembles

Permian formations of southern New Mexico, such as the Yeso and San Andres Formations of south-central New Mexico or the Colina and Epitaph Formations of southwestern New Mexico. This observation is supported by the discovery of Leonardian (middle Permian) conodonts in the upper limestone and dolomite unit (Thompson, 1982). The middle conglomerate in the Tres Hermanas Mountains (unit 4) more closely resembles some outcrops of the lower Tertiary(?) Lobo Formation of the Florida Mountains than it does Lower Cretaceous rocks. The crossbedded sandstone (unit 3) recently was found to be interbedded with pinkish, recrystallized limestone and is lithologically similar to the Permian Yeso Formation. The unit in the Tres Hermanas section that is most similar lithologically to Lower Cretaceous rocks in southwestern New Mexico is unit 1, which resembles the Lower Cretaceous section in the Victorio Mountains and the Hell-to-Finish Formation in the Big Hatchet, Little Hatchet, Animas, and East Potrillo Mountains. The absence of index fossils and the poor exposure of unit 1 make this correlation tentative. Therefore, at this time, it is best to exclude the Tres Hermanas section from the correlation chart (Fig. 8) pending further work.

In the Cooke's Range and southern San Andres Mountains of south-central New Mexico, the Lower Cretaceous is represented by the Sarten Formation, which is composed of approximately 150 m of marine and nonmarine quartzarenite and dark shale. The Sarten Formation ranges in age from late Albian to early Cenomanian, indicating a correlation with the Mojado Formation (Fig. 8; Clemons, 1982; Cobban, 1987). The Beartooth Formation in the Burro Mountains and near Silver City is lithologically similar to the Sarten Formation but has no index fossils. The Beartooth Formation is no younger than the overlying Colorado Shale of Cenomanian age and is either latest Early Cretaceous or earliest Late Cretaceous (Chafetz, 1982).

An attempt is also made in Fig. 8 to correlate Lower Cretaceous rocks in southern New Mexico with Lower Cretaceous rocks in Sierra Juarez, Mexico, and Cerro de Cristo Rey, Mexico and New Mexico, based entirely on published descriptions and previous correlations. In Sierra Juarez the lowest exposed Cretaceous unit is called Las Vigas equivalent and is composed of approximately 82 m of shale and limestone (Swift, 1972, 1973). Las Vigas is overlain by the Cuchillo Formation, which is composed of a lower clastic unit, a middle carbonate unit, and an upper clastic unit (Swift, 1972, 1973). The Cuchillo Formation in Chihuahua, Mexico, is considered to be primarily Aptian in age, but the Aptian-Albian boundary may be within the formation (Cordoba, 1969; Swift, 1972, 1973). The Cuchillo Formation matches lithologically with the mottled siltstone, lower limestone, and sandstone members in the East Potrillo Mountains (Craig, 1972). The Cuchillo Formation is overlain by a massive rudistid limestone, the Benigno Formation, which is early Albian in age in northern Mexico (Cordoba, 1969; Swift, 1972). The Benigno Formation is similar to the rudistid limestone member in the East Potrillo Mountains, although the rudistid limestone member is thinner. Thin-bedded limestone and shale of the lower middle Albian Lagrima Formation overlies the Benigno Formation (Cordoba, 1969; Swift, 1972). The uppermost units in the Sierra Juarez are the Finlay Limestone, which is middle Albian in northern Mexico (Cordoba, 1969; Wacker, 1972), and the Benevides Formation, which is from middle to late Albian in age (Cordoba, 1969). The massive limestone member in the East Potrillo Mountains occupies the same stratigraphic position as the Finlay Formation.

Approximately 12 km north of Sierra Juarez at Cerro de Cristo Rey, Mexico and New Mexico, the upper part of the Finlay Formation is conformably overlain by approximately 170 m of Lower Cretaceous rocks and 213 m of Upper Cretaceous rocks (Lovejoy, 1976). Directly above the Finlay Formation are shale and limestone of the Del Norte Formation, which correlates with the Fredericksburg Group of Texas and may contain the boundary between the middle and upper Albian (Lovejoy, 1976). The Del Norte Formation is overlain by the Smeltertown, Muleros, and Mesilla Valley Formations, which have Washita or upper Albian fossils that correlate with fossils in the Mojado and Sarten Formations of southwestern and south-central New Mexico (Lovejoy, 1976). The upper Albian formations at Cerro de Cristo Rey consist of shale, limestone, and sandstone and have a lower proportion of sandstone and a higher proportion of limestone than the Mojado and Sarten Formations (Lovejoy, 1976). Above the Mesilla Valley Formation is the Anapra Sandstone of earliest Cenomanian age, which correlates with the uppermost Mojado and Sarten Formations (Fig. 8). Above the Anapra Formation are the Del Rio and Buda Formations, which also may correlate with the uppermost Mojado Formation, although fossil evidence is sparse.

Depositional environments—The conglomerate member of the Hell-to-Finish Formation is the most difficult unit to interpret in terms of depositional environment because it contains the fewest features diagnostic of depositional processes. The conglomerates are grain supported, have well-rounded clasts, and have a well-sorted medium- to coarse-grained sand matrix (Fig. 3). These textural features indicate that the conglomerates were water laid. Although poor bedding in the conglomerate member may be viewed as evidence of gravity-flow processes, such as debris flow, there is no evidence of matrix support or poorly sorted, clay-rich matrix, which are commonly found in debris-flow deposits. There is no evidence of marine processes either. The thickness of the conglomerate member is highly variable, a characteristic feature of the basal Hell-to-Finish conglomerate in southwestern New Mexico and of the Glance Conglomerate in southeastern Arizona (Zeller, 1965; Bilodeau, 1982). The thickness variations indicate that the conglomerate member was deposited on a surface of locally high relief, and the similarity between clasts in the conglomerate member and the underlying Permian carbonates implies a local provenance.

Two different depositional models are presented for the conglomerate member. One possibility is deposition in a streamflood-dominated alluvial fan. This model is supported by the coarse grain size, texture, local provenance, and variable thickness. Streamflood processes are favored over debris-flow processes on low-angle fans with large radii (Steel, 1976). The dominance of streamflood fades also may reflect midfan deposition. Another model for the deposition of the conglomerate member is a proximal braided stream, termed the Scott type by Miall (1977). In a Scott-type braided-stream system, grain-supported conglomerates fill shallow channels and are superimposed vertically with little or no interbedded sand. The braided-stream model is supported by the absence of diagnostic alluvial-fan facies, such as debris-flow and sheetflood fades, and by the relatively minor thickness of the unit (<40 m). In the braided-stream model, the conglomerate member was deposited as a thin veneer of fluvial gravel.

The conglomerate member displays vertical changes in average maximum clast size (Fig. 4). Within the context of either depositional model, these grain-size cycles probably are tectonic in origin (Steel and Wilson, 1975; Steel et al., 1977). The upward-coarsening sequence may have been produced by initial uplift, whereas the upward-fining sequence represents the reduction in relief of the source terrane by erosion following the cessation of tectonism. Consequently, by the end of deposition of the conglomerate member much of the local relief probably was reduced substantially. An alternative control on the grain-size cycles climate, although no paleoclimate indicators are present it the conglomerate member to support or reject this control,

The conglomerate member is overlain by shallow-marine facies. The transition from braided stream or alluvial fan tc shallow marine without intervening coastal-plain or shoreline deposits may appear anomalous. However, most Cretaceous sedimentary rocks of the western United States wen deposited during regressive cycles, and there is little or nc sediment record of the transgressive periods (Ryer, 1977), As a result, lower offshore shales directly overlie shoreline or nonmarine fades in many areas.

The mottled siltstone member of the Hell-to-Finish Formation through the upper siltstone member of the U—Bar Formation have evidence of deposition in a shallow-marine setting. The mottled siltstone, sandstone, siltstone—limestone, and upper siltstone members contain a high proportion of calcareous siltstone and very fine and fine-grained sandstone. The siltstones and sandstones are massive, horizontally laminated, or hummocky cross-stratified; are moderately to heavily bioturbated; and have scattered pelecypod and gastropod shells or discrete shell layers less than 20 cm thick. All of the features are characteristic of modern sediment on the Atlantic and Gulf continental shelves of the United States (Berryhill, 1976; Bouma et al., 1982). Particularly diagnostic of shallow-marine processes is hummocky cross-stratification, which consists of sets of low-angle laminations with erosional lower bounding surfaces (Harms el al., 1975; Fig. 5). Hummocky cross-stratification is interpreted to be the result of combined unidirectional and osdilatory flows produced by storm waves (Hanes et al., 1975; Harms et al., 1982; Hunter and Clifton, 1982; Swift et al., 1983). Storm processes also may be responsible for the thin intraformational conglomerates that occur in the mottled siltstone member. Horizontally laminated siltstone and sandstone, as well as shale, were probably deposited from suspension during fair-weather periods and were not re-suspended by storms. Massive siltstones and sandstones also may represent deposition from suspension, or they may have been deposited originally by storms and subsequently reworked by burrowing organisms. The evidence from the fine-grained facies suggests deposition between normal wave base and storm wave base, depths that can vary between 15 m and 200 m (Swift et al., 1972; Walker, 1979).

Thin carbonates interbedded with the siltstones and sandstones in the mottled siltstone, siltstone—limestone, and upper siltstone members are primarily silty micrites and fossiliferous wackestones and packstones, which also are consistent with a shallow-marine setting. A few intraclast and oolite packstones in the siltstone—limestone member suggest periods of higher energy.

Crossbedded, gravelly sandstone and conglomerate 3-12 m thick are found interbedded with siltstone and fine sandstone in the mottled siltstone and sandstone members (Fig. 6). This coarse fades commonly coarsens upward, and, in some cases, a thin zone of intraformational rip-up clasts is found at the base. Trough crossbeds exhibiting polymodal paleocurrents are dominant, but horizontal laminations and low-angle laminations also are found. Pelecypod and turritellid gastropod shells are scattered throughout the beds or occur in discrete layers less than 20 cm thick. Burrows are rare, although a few thin horizons within the crossbedded fades are heavily bioturbated. One crossbedded sandstone bed in the sandstone member is oolitic. The mottled siltstone member has only one bed of crossbedded, granular sandstone, which is 6 m thick. In the northern measured section, the sandstone member has four beds of crossbedded, gravelly sandstone and conglomerate, ranging in thickness from 7 to 12 m and representing 60% of the thickness of the member. The sandstone member in the southern section also has four beds of the coarse facies, but they range in thickness from 1.5 to 4 m and represent only 42% of the member thickness.

The coarse, crossbedded fades in the mottled siltstone and sandstone members is interpreted to be offshore marine bars because they are interbedded with fine-grained shallow-marine fades and because they show no evidence of transitions to shoreline or nonmarine fades. The coarse, crossbedded fades is similar to descriptions of offshore-bar fades in the Jurassic of Wyoming (Brenner and Davis, 1974) and the Upper Cretaceous of Wyoming, Colorado, and northern New Mexico (Porter, 1976; Brenner, 1978; La Fon, 1981). The best modern analog is storm-generated marine bars on the Atlantic shelf of the southeastern United States (Duane et al., 1972; Swift et al., 1973; Stubblefield et al., 1975; Swift and Field, 1981). In this model linear sand ridges as much as 12 m thick and tens of kilometers long are formed by the transportation of coarse sediment as dune or sand-wave bedforms by storm-induced currents. Storm currents moving up the ridge flanks and the wave surge on ridge crests result in upward-coarsening sequences. During fair weather or less intense storms, finer sediment is deposited, or the bars may be bioturbated. The size and shape of the bars in the mottled siltstone and sandstone members were not determined.

The siliciclastic-rich members thin southward and display a southward decrease in the percentage of conglomerate and in the ratio of sandstone to siltstone. These data suggest relative southward sediment dispersal. This interpretation is consistent with clastic-dispersal data collected from the Hell-to-Finish and lower U—Bar Formations in southwestern and south-central New Mexico (Mack et al., 1986).

Limestones of the lower limestone, rudistid limestone, and massive limestone members of the U—Bar Formation also were deposited in a normal marine setting. The lower limestone member in the northern part of the range changes upsection from a fossiliferous packstone to sandy oolite grainstones. In the southern part of the range, the lower limestone member is a sandy fossiliferous wackestone. The oolite beds probably represent shallow shoals, such as those that occur on the Bahama Bank or in Florida Bay (Ball, 1967). Obviously these shoals were restricted areally because they are absent in the southern section. The oolite shoals may have been localized by a slope break or by irregular bottom topography. Modern oolite shoals exist in water as shallow as 5 m (Bathurst, 1971), and the change from the mottled siltstone member to the lower limestone member may be associated with a regression. The decrease in detrital sediment accompanying the change from the mottled siltstone to lower limestone member may reflect a shift in the locus of detrital depocenters or a tectonically or climatically controlled decrease in influx of detrital sediment.

The rudistid limestone member consists of a lower rudistid biostrome and upper turritellid gastropod wackestones. A paleoecological analysis by Pickens (1984) revealed monopleurid rudistid colonies at the base of the biostrome. With time caprinid rudistids replaced the monopleurids, probably in response to a decrease in influx of fine clastics. Subsequently, the caprinids declined and were replaced by solitary monopleurids. Rudistid biostromes are interpreted to have been deposited under normal-marine conditions anywhere from the inner shelf to the shelf—slope break (Wilson, 1975; Pickens, 1984). Thus, a significant change in water depth is not necessary during the transition from the sandstone member to the rudistid limestone member. As is the case with the lower limestone member, a decrease in detrital influx due to a shift in the detrital depocenter, tectonism, or climatic change is indicated by the appearance of the rudistid limestone member.

The thickest limestone in the East Potrillo Mountains is the massive limestone member. Unfortunately, all but the basal few beds are too recrystallized for microfacies study. The basal beds are silty gastropod, pelecypod wackestones, suggesting quiet-water marine deposition.

The Mojado(?) Formation in the East Potrillo Mountains is too poorly exposed to allow interpretation of depositional environment.

Provenance—Data used in provenance interpretations include lithologic identification of gravel-size clasts in the conglomerate member of the Hell-to-Finish Formation and thin-section point counts of nine sandstones of the mottled siltstone member of the Hell-to-Finish Formation, sandstone member of the U-Bar Formation, and Mojado(?) Formation and one sandy oolite limestone from the rudistid limestone member of the U-Bar (Table 1). Gravel clasts in the conglomerate member are restricted to limestone, dolomite, and chert. The carbonate clasts resemble Permian carbonates exposed in the East Potrillo Mountains, suggesting that the clasts in the conglomerate member are locally derived. Sedimentary source rocks for sandstones in the mottled siltstone, sandstone, and rudistid limestone members and Mojado(?) Formation are indicated by the high percentage of monocrystalline quartz, as well as the presence of chert and pelitic rock fragments. Carbonate grains are also common in the sandstones, but because of the possibility that some are intrabasinal rather than detrital in origin, they are excluded from detrital modes.

Sandstones of the mottled siltstone, sandstone, and rudistid limestone members also contain detrital feldspar, both K-feldspar and plagioclase (Table 1). In general, the percentage of feldspar increases upsection from the mottled siltstone member to the sandstone member and then decreases in the rudistid limestone member. The high percentage of total feldspar (≤30%) and the absence of volcanic rock fragments indicate granitic or gneissic source rocks, an interpretation supported by the presence of a few quartzofeldspathic rock fragments (Table 1). The most likely pre-Lower Cretaceous granitic or gneissic source rocks are Precambrian basement rocks. The Mojado(?) Formation in the East Potrillo Mountains consists primarily of monocrystalline quartz with a minor amount of chert, polycrystalline quartz, and pelitic rock fragments. The Mojado(?) Formation lacks feldspar and was derived exclusively from sedimentary source rocks.

Provenance data from the East Potrillo Mountains are consistent with sandstone provenance data collected from other Lower Cretaceous rocks in southwestern and south-central New Mexico (Mack, 1987). Regionally, the Hell-to-Finish and U-Bar Formations display a change upsection from quartz- and chert-rich sandstones to arkosic sandstones. This compositional trend probably reflects erosion of Paleozoic sedimentary rocks and unroofing of the crystalline basement-cored Burro uplift in west-central New Mexico (Elston, 1958; Mack, 1987). By late Albian time, during deposition of the Mojado, Sarten, and perhaps part of the Beartooth Formations, the sediment contribution from the crystalline basement diminished and eventually ceased, and the source rocks were exclusively sedimentary for the Sarten and Beartooth Formations and mixed sedimentary and volcanic for the Mojado Formation (Mack, 1987). Paleocurrents from the Mojado Formation in the Big Hatchet and Animas Mountains suggest a source to the west, whereas paleocurrents from the Sarten Formation in the Cooke's Range suggest a source to the northwest (Mack et al., 1986).

Upper Cretaceous and/or lower Tertiary rocks

Sedimentary rocks that are believed to be of latest Cretaceous or early Tertiary age (TKs) crop out in a few gullies adjacent to the Mt. Riley-Mt. Cox volcanic domes. The rocks dip moderately northeastward under and away from the summits of the eroded domes. No fossils were noted, so the age assignment is based solely on lithologic comparison with similar rocks elsewhere in the region.

Outcrops in three localities were mapped, one on the north side of Mt. Riley, one between Mt. Riley and Mt. Cox, and one on the southern flank of Mt. Cox. Different sections are exposed in all three areas, and the cumulative thickness is perhaps 200 m.

Exposures on the southwest side of Mt. Cox reveal approximately 30-40 m of pink mudstone, which are overlain by an equal thickness of limestone-pebble and cobble conglomerate, tan to olive sandstone and siltstone, and red shale. The conglomerate, sandstone, siltstone, and shale are interbedded to some extent but generally become finer grained upsection. Clast content in the conglomerate includes Lower Cretaceous and Permian lithologies similar to those cropping out in the East Potrillo Mountains. Overall, the conglomerate resembles the Love Ranch Formation (TKlr) of Las Cruces area, interpreted by numerous workers as syntectonic to posttectonic deposits of the Laramide orogeny (Lucas and Ingersoll, 1981).

Exposures between Mt. Riley and Mt. Cox and north of Mt. Riley are mostly tan, brown, reddish-brown, olive to green sandstone and shale with some beds of grayish-purple, brownish-purple and grayish-maroon siltstone and shale. A few limestone boulders weather from thin conglomerate units. Some of the sandstone and siltstone beds are volcaniclastic, derived from intermediate-composition volcanic or intrusive rocks. Because of the volcaniclastic component and because they appear to be stratigraphically higher than the rocks southwest of Mt. Cox, these rocks may be somewhat younger than the Love Ranch conglomerate; they may represent an upward transition into a middle Tertiary volcanic (Rubio Peak Formation, Trp) sequence. This is far from certain, however, and the whole section may prove to be correlative with part or all of the McRae Formation of the Elephant Butte area (Bushnell, 1953), which is Late Cretaceous and early Tertiary in age.

Santa Fe Group

The Santa Fe Group is the rock-stratigraphic unit that comprises the bulk of intermontane-basin deposits along the Rio Grande rift (Hawley et al., 1969; Gile et al., 1981). The oldest parts of the formation are fanglomerate and volcanic rocks of late Oligocene-early Miocene age that were emplaced during earliest stages of extensional block faulting in the region. The top of the Santa Fe is the youngest basin-fill surface predating entrenchment of the Rio Grande valley. This surface is generally middle Pleistocene (-0.4 Ma).

Most of the older parts of the Santa Fe Group that crop out in the southern Rio Grande rift are Miocene in age and were assigned by Kottlowski (1953, 1958, 1960) to the lower Santa Fe Group. In the Hatch-Rincon area, Hawley et al. (1969), Seager et al. (1971), and Seager and Hawley (1973) divided thick (1,500 m) lower Santa Fe beds into two formations: the older Hayner Ranch Formation and younger Rincon Valley Formation. Together, the two formations compose the bulk of the Santa Fe Group in the region and represent the major periods of bolson deposition associated with evolution of the Rio Grande rift. In general, the two formations are weakly to moderately tilted and locally have been offset a thousand or more meters along range-boundary faults.

Kottlowski (1953, 1958, 1960) designated the uppermost part of the Santa Fe Group as upper Santa Fe Group and considered it to be mainly Pleistocene in age, although now it is known that the oldest parts of the unit are late Pliocene

(Hawley, 1978). Hawley et al. (1969), Seager et al. (1971), and Seager and Hawley (1973) correlated the upper Santa Fe Group with the Camp Rice Formation of Strain (1966, 1969a, 1969b) and extended Camp Rice terminology into southern New Mexico. The Camp Rice Formation includes deposits of the ancestral Rio Grande as well as coeval piedmont-slope and basin-floor deposits. Unlike older parts of the Santa Fe Group, the Camp Rice Formation is thin (100-200 m) and nearly undeformed. This has resulted in a conspicuous unconformity between Camp Rice and older formations in all outcrop areas of the Santa Fe in southern New Mexico. Near central parts of modern basins, however, drilling has suggested that an unconformity between Camp Rice and older basin-fill deposits is lacking.

Both the Camp Rice Formation and fanglomerates correlative with the lower part of the Santa Fe Group crop out in the East Potrillo Mountains area.

Lower part of the Santa Fe Group—Tilted fanglomerate (Tsf) crops out over a broad area adjacent to the Mt. Riley fault zone, which is located a few kilometers west of the East Potrillo Mountains (geologic map, Sheet 1). Although the rocks are indurated, they are poorly exposed, generally being covered by a thin mantle of soil and clasts weathered from underlying rock. The exceptions are two hills of partly silicified, well-exposed fanglomerate in the northwest part of the outcrop belt and good outcrops in the deeper gullies that transect more southerly stretches of the belt. Exposed thickness is approximately 300 m. We believe these fanglomerates to be "early-rift" basin deposits of middle Tertiary age based on three observations: 1) similarity of the fanglomerate's volcanic dasts to the Rubio Peak and younger middle Tertiary volcanic sequences, 2) lithologic similarity with upper Oligocene to middle Miocene fanglomerate of the Hayner Ranch and Rincon Valley Formations north of Las Cruces, and 3) westward tilting of the fanglomerate by as much as 25°, which precludes its assignment to the Camp Rice Formation.

A varied source terrane and midfan depositional site are indicated by the composition of and structures in the fanglomerate. Conglomerate and conglomeratic sandstone are the dominant rock types, although sandstone and mud-stone are interbedded with coarser-grained rocks. Cementation is tight with calcite or silica, and the rock is generally well indurated. Bedding ranges from planar to lenticular and channelform as would be expected on an alluvial fan (Bull, 1972). Trough crossbedding is common. Colors range from dark brown to pale purple brown. Clasts generally are grain-supported, rounded to subangular, cobble to boulder size, and up to 28 cm in diameter. They include: Lower Cretaceous and Permian limestone and siltstone similar to outcropping rocks of the East Potrillo Mountains, lithologies akin to Rubio Peak or Palm Park andesite, basaltic andesite, felsic rocks including flow-banded rhyolite, and minor red Precambrian granite.

The granite clasts are of special interest because they indicate uplift of basement-cored blocks in middle Tertiary or earlier time. The source of the clasts might have been recycled Love Ranch fanglomerate, which, as xenoliths in the maar deposits of Kilbourne Hole, contain Precambrian clasts. Alternatively, the source might have been Precambrian bedrock. The nearest outcrop of Precambrian granite, at Eagle's Nest, is 32 km northwest of the East Potrillo Mountains, far beyond the modern drainage net. At that locality, Cretaceous rocks overlie the granite on what we presume to be a southeastern extension of the Burro uplift (Elston, 1958). Apparently, this uplift, or a basement-cored Laramide uplift, or Love Ranch fanglomerate associated with a basement-cored Laramide uplift was exposed by middle Tertiary faulting in the East Potrillo area, and it furnished small amounts of granitic detritus to the Santa Fe fanglomerate.

Camp Rice Formation—The upper part of the Santa Fe Group consists of alluvial-fan and basin-floor deposits of late Pliocene to middle Pleistocene age, which are assigned to the Camp Rice Formation (Strain, 1966; Hawley et al., 1969, 1976; Seager et al., 1971, 1984; Gile et al., 1981). In contrast with lower Santa Fe Group fanglomerates, whose source rocks are largely buried or otherwise obscure, alluvial-fan deposits in the Camp Rice clearly are derived from modern uplands and mountains.

In the Rio Grande region the uppermost beds of the Camp Rice Formation mark the final aggradation of bolson floors before entrenchment of the Rio Grande. The constructional top of these deposits forms a widespread basin-floor surface known as La Mesa surface (Ruhe, 1964, 1967). Fluvial deposits of the ancestral Rio Grande underlie La Mesa surface east of the East Potrillo Mountains and compose the bulk of the Camp Rice Formation. In and around the East Potrillo area, however, these fluvial strata pinch out westward into other kinds of basin-floor deposits or into alluvial fans.

Both basin-floor and alluvial-fan deposits compose the Camp Rice Formation in the map area. Basin-floor sediments (Qcb) include deposits of alluvial flats, playas, and dune fields. Generally they are poorly to moderately indurated calcareous mudstone, siltstone, and sandstone, pink, tan, pale yellow or gray in color. The nearly flat constructional surface of these deposits—La Mesa surface—is preserved over substantial parts of the map area. Soil-carbonate horizons as thick as 1.5 m have developed in upper parts of the basin-floor sediments below La Mesa surface. A basalt flow that overlies La Mesa surface on the northern rim of Potrillo maar is 1.2 Ma, which indicates that the surface west of the East Potrillo fault as well as the underlying Camp Rice strata is still older. To the east of the East Potrillo fault, La Mesa surface may be as young as 0.5 Ma (Seager et al., 1984). Evidently, La Mesa surface varies in age from place to place, especially across range-boundary fault zones. One reason for this may be that uplift along such faults isolated upthrown parts of the surface from further aggradation, while in downthrown areas Camp Rice sedimentation continued uninterrupted (J. W. Hawley, pers. comm. 1982).

Camp Rice alluvial-fan deposits include a lower sequence of well-cemented gravel and an upper sequence of less indurated gravel (Qcp). In many places the two sequences cannot be readily distinguished and were mapped as one (QTc). The older sequence consists of calcite-cemented conglomerate derived from local mountains and uplands. Cementation is thorough, probably the result of precipitation from ground water when the gravels were temporarily in the zone of saturation (J. W. Hawley, pers. comm. 1982). Exposed thickness of these gravels is only a few meters, but the unit may be many tens or hundreds of meters thick in the subsurface. Best exposures are in the bottom of the deepest gullies, on piedmont slopes above the East Potrillo fault, and on the flanks of Mt. Riley-Mt. Cox.

Unlike the tightly cemented lower gravels, the younger fan deposits are poorly indurated except for the upper meter or so, which is lithified by pedogenic calcrete. Source terranes are the same for upper and lower gravel units, as are general clast and bedding characteristics, which suggest transport by debris-flow, sheetflood, and arroyo-channel mechanisms. The upper deposit is generally less than 7 m thick, and its constructional upper surface—the Jornada I surface (Gile and Hawley, 1968)—grades downslope into La Mesa surface on basin floors. Dating of the basalt flows or ashes in southern New Mexico demonstrates that oldest parts of Camp Rice fans are late Pliocene (3.1 Ma) in age, whereas youngest parts are middle Pleistocene (0.4-0.3 Ma; Gile et al., 1981; Seager et al., 1984).

Upper Pleistocene deposits

Upper Pleistocene alluvial-fan and basin-floor deposits (Qpo) are inset against or overlie Camp Rice deposits. They are most extensively developed along major modem drainageways, and they have as their source the modern mountains or uplands in the map area. Lithologically, both alluvial-fan and basin-floor deposits are similar to Camp Rice strata, but there is a lesser degree of soil development in upper parts of the upper Pleistocene deposits. Soil-carbonate horizons generally are 1 m thick or less, and except for the oldest upper Pleistocene soils, pedogenic calcrete is absent. Thus, most of the upper Pleistocene deposits are rather loosely cemented or uncemented, even in the soil horizons. Near mountain fronts upper Pleistocene fan gravel may be as much as 10 m thick, thinning downslope to veneers 3 m or less.

Uppermost Pleistocene to Holocene deposits

Uppermost Pleistocene to Holocene deposits include the sandy or gravelly alluvium of modern arroyo floors as well as piedmont-slope gravels at the mouths of such drainageways. Fine-grained basin-floor deposits, playa deposits, and eolian quartz sand dunes or sand sheets also mantle broad parts of the map area. All of these deposits are completely nonindurated, and soils are only weakly developed. Soil carbonate occurs as thin threads, small nodules, or veins; in finer-grained soils on basin floors, slight accumulations of organic material and clay may dominate over carbonate. Thicknesses are less than 5 m.

Igneous and volcaniclastic rocks

Rubio Peak Formation

Volcanic and volcaniclastic rocks (Trp) that we believe to be correlative with the Rubio Peak Formation, Palm Park Formation, and Orejon Andesite (Elston, 1957; Kelley and Silver, 1952;, Dunham, 1935) crop out along the northwestern flank of Mt. Riley-Mt. Cox domes. Exposures are confined to a few small gully bottoms. Purple to purplish- or bluish-gray porphyritic flows and laharic breccia of intermediate composition are most typical of the unit. Phenocrysts are predominantly plagioclase with lesser numbers of hornblende and biotite. Purplish volcaniclastic sandstone and siltstone also are exposed, and light-colored tuffaceous sandstone with intercalated andesite flows compose stratigraphically highest exposures. Brown sandstone- and limestone-pebble conglomerate crops out locally and may represent a transition into an underlying Upper Cretaceous-lower Tertiary section. The total exposed thickness is a few hundred meters. Based on radiometric dates of correlative rock units in southern New Mexico, the Rubio Peak Formation in this area is probably late Eocene to early Oligocene in age (Clemons, 1982).

Fine-grained intermediate-composition volcanic rocks

A series of intermediate-composition volcanic and intrusive rocks (Tvi) that may be younger than the Rubio Peak Formation is exposed in the group of hills along the road 3-5 km south of Mt. Cox. Unlike the Rubio Peak Formation, these igneous rocks are generally nonporphyritic or sparsely porphyritic, although several units contain tiny needlelike phenocrysts of hornblende. Texturally and mineralogically many units resemble the andesitic to rhyodacitic rocks of the Mt. Riley-Mt. Cox volcanic domes. Colors range from bluish gray to dark purplish brown, but shades of maroon and purple are most common. Some rocks are vesicular; others are very aphanitic and dense. Siliceous laharic breccia is locally common. Flow banding is moderately developed in hill 4322, and this may be part of an intrusive mass. We presume, however, because of the variety of lithologies, that the unit consists mostly of flows.

Along the eastern side of hill 4474 brown to purplish-brown, volcaniclastic sandstone, mudstone, and siltstone, presumably part of the volcanic section, are interbedded with siliceous breccia and fanglomerate. The breccia and fanglomerate beds are lithologically similar to the lower Santa Fe fanglomerate described earlier and are in fault contact with a well-exposed section of lower Santa Fe rocks. We interpret these interbedded breccias and fanglomerates within the volcanic section to indicate that the volcanic sequence intertongues upward with the Santa Fe "early-rift" rocks. Assuming that the Santa Fe rocks are lower Miocene—an assumption based on lithologic correlation with dated lower Santa Fe fanglomerates elsewhere—the volcanic section is somewhat older than the Santa Fe but younger than the Rubio Peak Formation. Tentatively then, we consider these volcanic rocks to be middle to late Oligocene in age, perhaps 35-26 Ma.

Mt. Riley-Mt. Cox volcanic domes

Millican (1971) studied the intrusive-extrusive rocks of Mt. Riley-Mt. Cox (Tar). The following petrographic description is taken from his study and from the account by Hoffer (1976). The domes consist of fine-grained, pale greenish-gray microporphyritic andesite to rhyodacite. Flow banding is developed locally along wall-rock contacts but is either faint or nonexistent a few meters from the contact. Phenocrysts, which average approximately 7% of the rock, are plagioclase, hornblende, and biotite and average approximately 4 mm in length. Tabular phenocrysts are generally subparallel. The groundmass is trachytic to pilotaxitic and consists of subparallel plagioclase laths and interstitial alkali feldspar and quartz. Some of the quartz is xenocrystic, and a variable quartz content generally accounts for the compositional range.

This unit is probably the erosional remnant of a viscous lava dome, most of which stands above the land surface on which it erupted. Contacts with country rock are not well exposed owing to talus and alluvial fans that overlap high onto the intrusive. Two exposures, however, reveal inward-dipping flow banding in the intrusive at the wall-rock contact. The wall rock itself locally dips beneath the dome and elsewhere away from the dome. The interpretation of the volcanic mass as an eroded, steep-sided, probably funnel-shaped dome is based largely on its subvolcanic texture and on the inward dips of observed contacts.

The age of the Mt. Riley-Mt. Cox intrusive is not known. The intrusive may be correlative with the intermediate-composition flows and intrusions (Tvi) that crop out along the road south of Mt. Cox. Similar compositions and textures support this possibility. These flows seem to be middle to late Oligocene in age. If the correlation is correct, then the Mt. Riley-Mt. Cox domes also are middle to late Oligocene in age, perhaps 35-26 Ma.

Other intrusive rocks

Andesite-latite dikes and minor sills (Ta) similar in composition and texture to the Mt. Riley-Mt. Cox domes crop out sparsely in the East Potrillo Mountains. The dikes trend east-northeast and are 1-2 m wide and as much as 0.5 km long. They transect Laramide structures but probably (not certainly) are cut by middle Tertiary low-angle faults. We presume they are correlative with the Mt. Riley-Mt. Cox domes and therefore are middle to late Oligocene in age.

A single rhyolite dike (Tr), trending northeasterly, cuts Lower Cretaceous strata in the southern part of the East Potrillo Mountains. The rock lacks flow banding, being very fine grained to aplitic in texture. Its age is unknown.

West Potrillo Basalt

The West Potrillo Mountains are an uninhabited stretch of uplands lying just west of the East Potrillo map area. Rather than consisting of deformed and uplifted rock, the West Potrillo "Mountains" are a collection of more than 150 cinder cones, maar volcanoes, and voluminous basalt outpourings (Qb), extensively covered by a thin mantle of sand but still rugged enough and desolate enough to compare favorably with a lunar landscape. The basalts overlie a deep northerly trending gravity low as well as a gravity high (Keller and Cordell, 1983). The gravity high represents the northwestern extension of the East Potrillo Mountains; the low is a major early-rift basin modified by late-rift faulting. Thus the basalts have buried major rift basins and uplifts. The basalt field, which covers approximately 500 km^2, extends onto the western edge of the map area (Sheet 1).

Renault (1970), Page (1973), Hoffer (1976), Bersch (1977), Ortiz (1979), and Sheffield (1981) have studied various aspects of the West Potrillo Basalt. The basalts are alkali-olivine basalts, whose silica content averages approximately 44.5%, total alkalis approximately 5%, and TiO_2 approximately 2.3% (Hoffer, 1976). Typical flows are hypocrystalline with microphenocrysts of olivine and some plagioclase and pyroxene. The groundmass is mostly pyroxene and plagioclase with small amounts of opaques, glass, olivine, and feldspathoids. In addition, many of the flows carry xenocrysts of anorthodase, plagioclases, pyroxene, and amphibole, as well as xenoliths of olivine-pyroxene-spinel aggregates (Hoffer and Hoffer, 1973; Hoffer, 1976; Ortiz, 1979).

The age of the West Potrillo Basalt is not well constrained. The stage of soil development, especially pedogenic carbonate, on most flows or cinder cones resembles the stage of soil development on uppermost Camp Rice strata. Based on this comparison and on the fact that upper Camp Rice beds are middle Pleistocene (about 0.5-0.3 Ma), it seems likely that the Potrillo basalts are largely, if not entirely, pre-middle Pleistocene. One K-Ar whole-rock age of 1.2 ± 0.06 Ma was reported for a basalt flow and spatter cone located on the northern rim of Potrillo maar (Seager et al., 1984). This may be a representative age for many of the flows in the field, although we would not be surprised if ages between about 3.0 Ma and 0.5 Ma were obtained eventually. On the basis of morphology, petrography, and other characteristics, Hoffer (1976) indicated that older and younger groups of flows and cones existed in the field, but he was unable to give any absolute age assignments.

Potrillo maar rim deposits

Potrillo maar, described by Reeves and De Hon (1965), straddles the Mexico-New Mexico border a few kilometers south of the East Potrillo Mountains. Phreatomagmatic in origin, it is very similar in morphology, structure, and composition to the better-known Kilbourne Hole and Hunts Hole (Reiche, 1940; De Hon, 1965a, b; Shoemaker, 1957). However, unlike the other maars, Potrillo maar exhibits a group of small cones and short flows on its floor (in Mexico), the final stage in the phreatomagmatic activity that built the structure. Seager et al. (1984) obtained an age of 0.18 ± 0.03 Ma for these flows, which is probably only slightly younger than the maar itself. The premaar basalt flow on the northern rim of the maar is 1.2±0.06 Ma (Seager et al., 1984).

Rim ejecta (Qt) is well exposed in the United States on the northern rim of the maar (geologic map, Sheet 1). It consists of thinly bedded sand- to silt-size fragments of disaggregated Camp Rice Formation (on which the tuffs are lying), abundant sideromelane, accretionary lapilli, and many accidental xenoliths as much as 1.0 m in diameter. Lherzolite, granulite, granite, gneiss, and volcanic and sedimentary xenoliths probably represent many levels of the crust and upper mantle. Most of the fragments are moderately well rounded, partly as a result of abrasion during their transport to the surface. Volcanic and limestone lasts are derived, in part at least, from disrupted upper and lower Santa Fe Group fanglomerate, so these clasts may owe their rounding to sedimentary as well as to volcanic processes. Occasional sag structures in the finer-grained rim deposits were produced by impact of one xenolith or another. Low-angle crossbedding in the rim deposits suggests that repeated base-surge eruptions below a dense water- and ash-laden eruption cloud was as common at Potrillo maar as it was at Kilbourne Hole (Shoemaker, 1957; Hoffer, 1976).

Structural geology

Structural features (folds, faults, and homoclinal, tilted strata) in the map area formed during three deformational events. Although not absolutely dated, the three events clearly are Late Cretaceous and Cenozoic in age based on the general structural style and on the fact that Lower Cretaceous, Tertiary, and Quaternary rocks are deformed.

The three sets of structures can be relatively dated by crosscutting relationships in the East Potrillo Mountains where all of the structures are best displayed. Oldest of the structures are folds and associated thrust faults that are the products of Laramide compression in Late Cretaceous-early Tertiary time. These structures are in turn dissected by a system of low-angle normal faults that are associated with moderately rotated strata; we believe these structures resulted from early extension (late Oligocene-early Miocene) in the Rio Grande rift. Finally, the most recent set of structures are high-angle normal faults that cut middle Tertiary and Quaternary fanglomerate and border the modern fault-block uplifts and basins. Movement on these faults has continued into the middle or late Quaternary.

Laramide faults and folds

Laramide compressional structures are exposed in the central and northern parts of the East Potrillo Mountains. Cross sections *A-A'* and *B-B'* (Sheet 2) best illustrate the general structural style, which is also present in sections *C-C'*, *D-D'*, *H-H'*, and *I-I'*. In both northern and central areas thrust faults dipping 30°-60°W carried allochthonous hanging-wall rocks northeastward; the direction of tectonic transport is abundantly clear from the asymmetry of drag folds. Below the thrust faults, Lower Cretaceous footwall rocks are folded into tight, eastward-overturned synclines, locally with thrust-faulted cores (Fig. 9). The thrust faults and folds trend N10°W-N30°W, approximately perpendicular to the movement direction of the thrust.

Laramide deformation between the northern and central parts of the range does not appear to be continuous, although this is not certain owing to disruption of the Laramide structures by younger faulting. Some Laramide folds and faults in the central area die out northward within 23 km. Indeed, the deformation in this central area is less severe compared to that in the northern extremity of the range. In fact, the general structure in the central region may be best described as a simple, overturned anticline-syncline pair with a conspicuous west-dipping thrust fault in the core of the anticline (sections *H-H'* and I-I'*)*. This core thrust has moved the right-side-up limb of the anticline eastward onto the adjacent near-vertical to overturned limb of the syncline. Because the thrusts and folds die out within 2-3 km north of their best developed exposure and because Lower Cretaceous rock units on both sides of the thrust fault are identical in thickness and fades, it seems clear that displacement on the fold-thrust zone is relatively small, a kilometer or less.

The Laramide folds and fault just described are in the hanging wall of a younger low-angle normal fault zone, which will be described in the next section. Below the normal fault, Laramide folds and thrusts are exposed again, but they cannot be matched with certainty to the structures above the low-angle normal fault (section I-I'*)*. The most important feature below the normal fault is a westward-dipping thrust, which translated Permian carbonates eastward over the Lower Cretaceous mottled siltstone member of the Hell-to-Finish Formation (Khb). Does this represent a deeper structural level of the thrust-faulted anticline above the normal fault, or is it a separate, deeper thrust, perhaps with major slip? We do not see enough evidence to favor any one interpretation.

Although similar in overall structural style to that of the central area, Laramide deformation in the northern part of the range exhibits four important, or at least distinctive, variations. First, the deformation appears to be more severe, the resulting structures larger. This is especially true of the overturned syncline below the major west-dipping thrust fault, as an examination of sections *A-A'* through *C-C'* will show. Second, intense crumpling and thrust faulting of the synclinal core is present locally (section *A-A').* Southeastward along strike of the syncline axis, thrust faulting in or near the axis increases in apparent displacement as indicated by juxtaposition of strata. Eventually, in the western overturned limb of the syncline, inverted beds, which dip 10°-30° westward, are in fault contact with near horizontal beds of the upright limb (sections *D-D'* and, espedally, *E-E';* Fig. 10). Third, nearly flat to moderately west dipping thrust faults locally break the (eastern) right-sideup limb of the syncline, resulting in repetition of Cretaceous rock units (sections *A-A'* and *B-B').* Drag folding adjacent to these thrusts has resulted in some spectacular outcrops of deformed rocks on mountain sides (Fig. 11). Shortening by movement on individual thrusts and associated drag folds in this right-side-up limb is approximately 100-200 m, and the total displacement across the system of from two to three thrusts (eastern half of *B-B')* is probably less than 0.6 km.

The fourth and most important variation that distinguishes the northern belt of deformation is the character of thrust faulting above and to the west of the overturned syncline. What appears to be essentially a single (locally two) thrust fault dipping 35°-60°W has carried Permian carbonates onto the overturned limb of the adjacent syncline. In most places along the thrust, the thrust sheet of Permian rocks lies on overturned Permian carbonates that form the overturned limb of the syncline (sections *A-A'* through *CC').* Locally, the thrust transgresses downward (stratigraphically upward) through the overturned limb into underlying Lower Cretaceous siltstone. In these areas, Permian rocks are thrust over inverted Cretaceous rocks (sections D-D'and *E-E';* Fig. 10). Permian rocks in the thrust sheet locally exhibit tight to open folds with no obvious direction of asymmetry. In general, the Permian strata dip westward 20°-40°, complicated locally by zones of intense shattering, bedding-plane slippage, and silicification. No clear top- or bottom-bedding indicators were identified in carbonates of the thrust sheet; we presume that these rocks are in normal stratigraphic order, although they clearly lie on and are semiparallel to inverted Permian and Cretaceous rocks below the thrust surface. This assumption of normal order of strata in the thrust sheet is based largely on analogy with

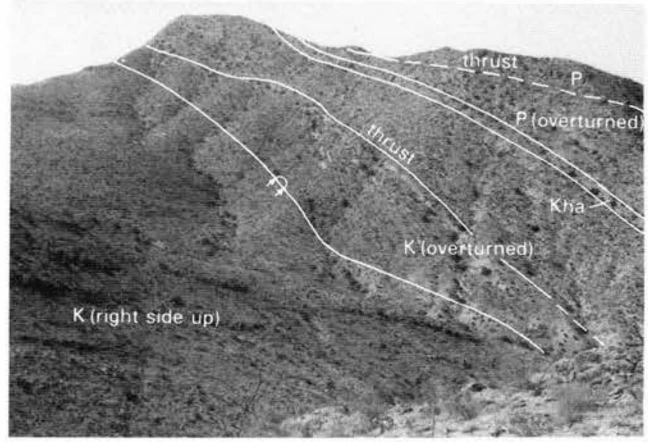

FIGURE 9—View looking southwestward at overturned syncline at northern end of East Potrillo Mountains. Thrust sheet of Permian rocks (in normal order?) at top of ridge has moved northeastward (toward viewer) over the overturned limb of the syncline. **Kha** = basal Lower Cretaceous conglomerate, **K** = Cretaceous rocks, and **P** = Permian rocks.

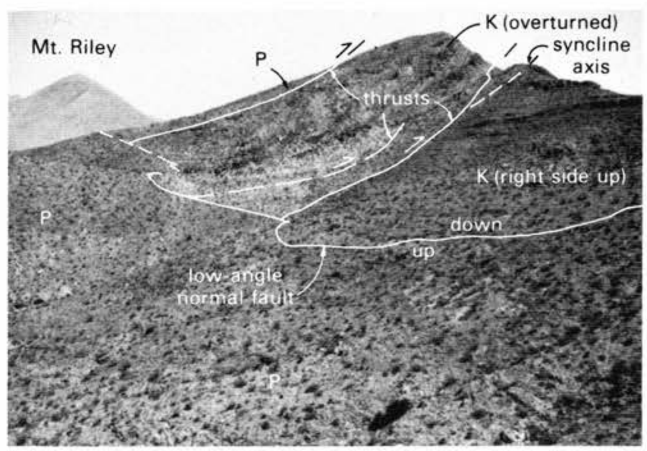

FIGURE 10—View looking northeastward at thrust-faulted, overturned syncline at northern edge of East Potrillo Mountains. Thrust sheet of Permian rocks (in normal order?) on skyline has been thrust northeastward over the overturned limb of the syncline. Cross sections *D–D'* and *E–E'* show relationships in this photo. **K** = Cretaceous rocks and **P** = Permian rocks.

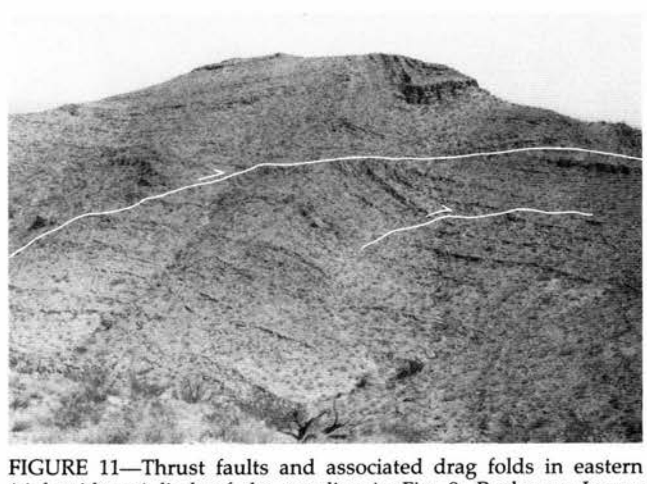

FIGURE 11—Thrust faults and associated drag folds in eastern (right-side-up) limb of the syncline in Fig. 9. Rocks are Lower Cretaceous units **Kub** through **Kud**. Cross section *B–B'* shows structure in this photo. View looks north–northwest.

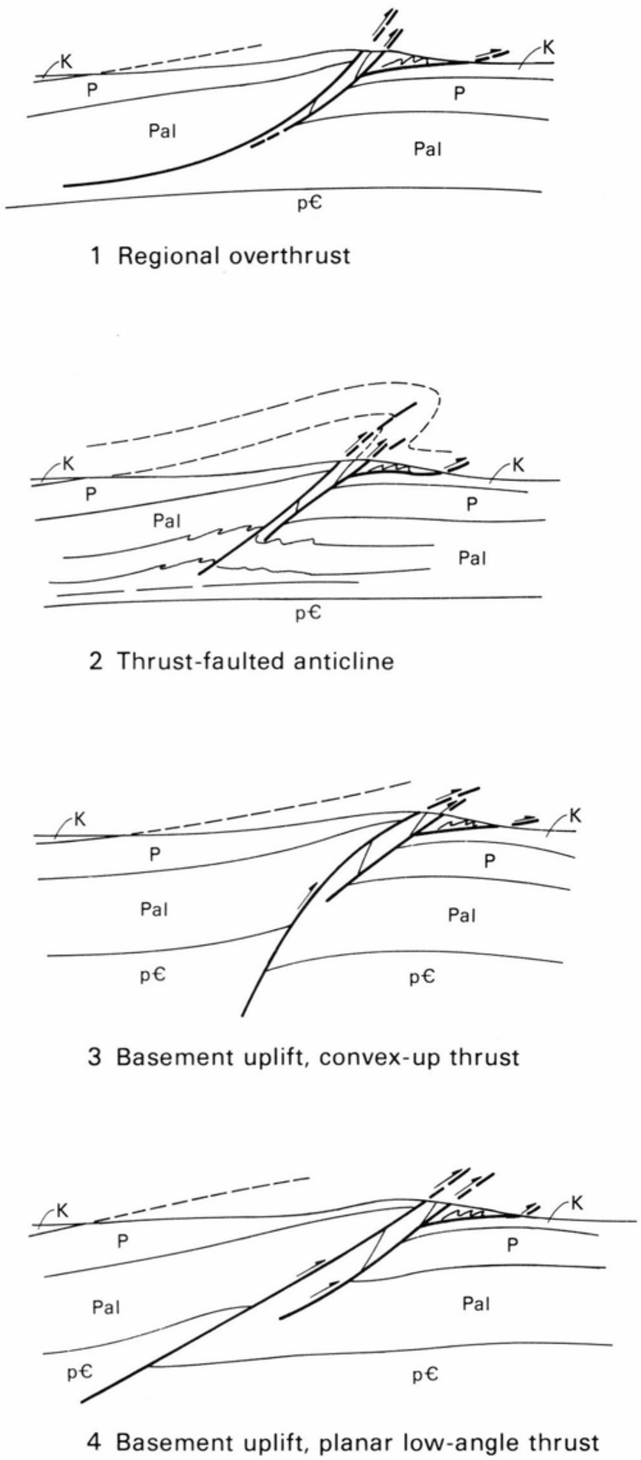

FIGURE 12—Four models of thrusting that could account for the style of Laramide deformation observed in the East Potrillo Mountains are: **1,** thin-skinned regional overthrusting similar to that interpreted for the Juarez Mountains, Mexico, and other ranges in southwestern New Mexico by Corbitt and Woodward (1973), Woodward and DuChene (1981), and Drewes (1978); 2, thrust-faulted anticline similar to the small-scale structure mapped in the central part of the East Potrillo Mountains (sections *G–G'* and *H–H')*; 3, rigid basement-block uplift bounded by upward-flattening "upthrust"; crustal shortening in this model is less significant than vertical uplift; and **4,** rigid basement uplift bounded by low- to moderate-angle, planar thrust; crustal shortening in this model may be significant. Models three and four are similar to Laramide structure northeast of the East Potrillo Range in the San Andres and Caballo uplifts and at San Diego Mountain. **K =** Cretaceous rocks, **P =** Permian rocks, **Pal =** Paleozoic rocks, and **pC =** Precambrian rocks.

the right-side-up Cretaceous section above the thrust in the central part of the range.

Middle to late Tertiary rotation of Laramide structure

As discussed in a later section, the strata in the East Potrillo Mountains apparently were rotated approximately 25° southwestward in middle Tertiary time. If we remove this amount of deformation from Laramide structures, their geometry may be more correctly revealed. Two effects of this operation would be: 1) to flatten Laramide thrust-fault dips to 10°-30°W dips, and 2) to accentuate the asymmetry of overturned synclines in front of the thrusts. A word of caution may be appropriate here, however. Although approximately 25° appears to be a reasonable estimate of middle Tertiary rotation for the southern two-thirds of the range, the amount of tilting in the northern part of the range, where Laramide deformation is extensive, is poorly constrained. Tilting may be less in that area compared to farther south, and certainly cross sections *A—A'* through *E—E'* show no obvious effects of a 25° rotation.

Laramide models

Fig. 12 shows four possible models for Laramide deformation in the East Potrillo Mountains. Corbitt and Woodward (1973), Drewes (1978), and Woodward and DuChene (1981, 1982) have favored overthrust models for southwestern New Mexico, while Brown and Clemons (1983),

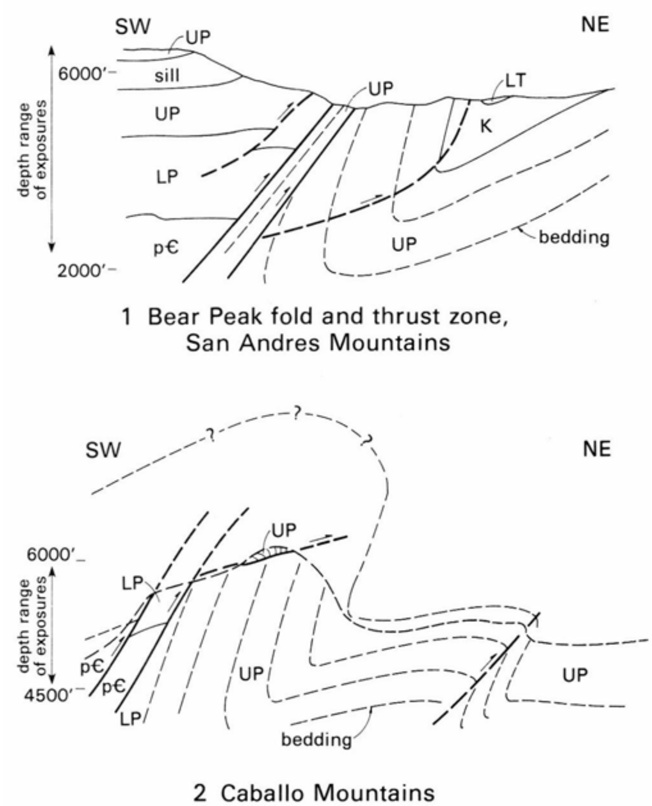

FIGURE 13—Cross sections of **1,** the Bear Peak fold and thrust zone in the southern San Andres Mountains (after Seager, 1981) and **2,** part of the southern Caballo Mountains (Seager, unpubl. maps; Kelley and Silver, 1952). The shallow structure of these deformed zones is similar in geometry with fold and thrust structure in the East Potrillo Mountains. In the San Andres and Caballo areas, however, exposures are deep enough to show that thrust faults transect basement. **LT =** lower Tertiary fanglomerate, **K =** Cretaceous rocks, **UP =** upper Paleozoic rocks, **LP =** lower Paleozoic rocks, and **pC =** Precambrian rocks.

Seager (1983), and Seager and Mack (1986) preferred basement-cored uplift and convergent wrenching models. Clearly, exposures are not sufficient in the East Potrillo Mountains to distinguish among the four models, but some comparisons with other Laramide structures in south-central New Mexico and adjacent areas may be instructive.

Superficially, it is easy to see how an overthrust, "thin-skinned" model of Laramide deformation could apply to the East Potrillo Mountains. A similar style of folding and thrusting has been described from the Juarez Mountains 50 km to the southeast (Lovejoy, 1980). Although the level of erosion is not deep enough to prove it, deformation in the Juarez Mountains has been considered a good example of regional "thin-skinned" overthrusting (Drewes, 1978, 1982; Corbett and Woodward, 1973).

On the other hand, the geometry of folds and thrusts in the East Potrillo Mountains is also identical to that of the higher structural levels of Laramide deformation in the southern San Andres and Caballo Mountains (Fig. 13). In these areas and at San Diego Mountain north of Las Cruces, the level of erosion is sufficiently deep to demonstrate that the thrust-fault systems transect basement rocks at angles of 30°-60°. Although the East Potrillo structures may be readily interpreted as thin-skinned overthrusts, this interpretation cannot be confirmed with the data at hand, and analogous structures not far to the north support the possibility that the East Potrillo thrust system may transect basement and mark the edge of a basement-cored uplift. Precambrian clasts in Love Ranch fanglomerate (from Kilbourne Hole) and in middle Tertiary fanglomerate also hint at the possibility that basement was uplifted and exposed in the East Potrillo area in Laramide time (see section on Lower part of the Santa Fe Group, p. 16).

Middle Tertiary low-angle normal faults

Laramide folds and thrust faults are cut by a system of younger low-angle normal faults. Although the precise age of these faults is in doubt, they are clearly younger than Laramide structures and appear to be older than the through-going East Potrillo fault, the eastern boundary fault of the modern horst. We believe that the low-angle faults are late Oligocene-early Miocene in age. The faults are structurally similar to better-dated upper Oligocene-middle Miocene low-angle normal-fault systems elsewhere in the Rio Grande rift (e.g. Organ-San Andres-Franklin Mountains, Seager, 1981; Lemitar Mountains, Chamberlin, 1983). Furthermore, the faults and adjacent rocks are locally silicified and have controlled sulfide mineralization; this also suggests a temporal association with middle Tertiary magmatism.

Because of their low angle of dip, many of the faults, especially west-dipping or flat ones, may be mistaken for thrust faults. However, the low-angle faults differ from Laramide thrust faults in two important ways. First, whereas Laramide thrust faulting caused older rocks to move above younger ones, resulting in a vertical repetition of rock units, the younger low-angle faulting invariably placed younger strata on older, resulting in omission of part of the section. In outcrop, map view, and cross sections, the dip of faults and downward relative movement of the hanging wall is always typical of normal faults (e.g. sections G-G'and *H-H')*. Second, although small drag folds or broad zones of gentle warping are locally present adjacent to low-angle normal faults, there are no related, large-scale, tight folds as there are associated in every case with Laramide thrusts. Except where Laramide folding is apparent, footwall and hanging-wall blocks of the low-angle normal faults are homoclinally tilted to the southwest from 15° to 50°, averaging 20°-25°.

The array of closely spaced low-angle faults that disrupt strata in the central two-thirds of the range apparently constitutes a single fault system. The present geometry of the system is best described as shovel shaped with a trough that plunges shallowly north-northeast. Inasmuch as the fault system appears to have been rotated west-southwestward 20°-25° with adjacent strata, removal of this tilt should restore the fault system to its original shape. We will address this problem after describing the system's present geometry.

Present fault geometry

The geologic map, cross sections *C-C'* through *K-K'* (especially and Fig. 14 (in pocket) illustrate the geometry of the low-angle fault system. Major low-angle faults trend predominantly N30°W, parallel to both the Laramide structural grain and to the trend of the modern East Potrillo Mountains boundary fault and uplift. Low-angle faults dip northeast in the northern part of the range and southwest in the southern part at angles of 0°-50°. Because they are nearly parallel to bedding, the southwest-dipping faults are difficult to detect. In transverse canyons, however, they can be seen to gently truncate bedding (sections J-J'and *K-K')*. Faults in the major zones branch and rejoin on both small and large scales (e.g. section *F-F');* locally they also change strike abruptly, apparently reflecting both small- and large-scale grooves or mullion structures on the fault surface.

Near the center of the range (Fig. 14, point A) both northeast- and southwest-dipping fault segments bend abruptly to more easterly strikes, with dips changing to a more northerly direction at shallow to moderate angles. Along the eastern face of the range, the bending continues in certain strands of the fault zone until these faults strike east-northeast and dip northwest approximately 30° (Fig. 14). Seven faults to the north of and parallel to these east-northeast-striking fault strands repeat strata along the eastern base of the range; they are also low angle and are down-thrown to the northwest. These faults die out upsection and at least locally exhibit small amounts of reverse drag in the hanging wall.

As shown in Fig. 14, the overall present geometry of the fault system, including the northeast-striking set of imbricate fractures, may be interpreted as being shovel shaped, the shovel opening and deepening to the north-northeast. Faults in the northern and central parts of the range suggest this geometry most strongly, but structurally lower, southwest-dipping faults in the southern part of the range seem to be part of the same shovel-shaped system. Stratigraphic separation across the whole system is approximately 500 m. The slip direction is not well constrained, but all except one of five slickensided surfaces suggest slip in a N20°EN35°E direction, essentially down the trough axis of the shovel-shaped system. These slickensides were taken from all parts of the main fault zone as well as from one of the northeast-trending short faults. The anomalous slickenside trends N55°E, essentially a dip-slip direction for N30°Wstriking fault segments.

Original geometry and nature of fault system

Interpretations of low-angle fault-zone geometry affect one's view about the character of extensional tectonics in a region. Are the low-angle faults simply the flatter parts of listric faults with stratal rotation a consequence of fault geometry? Or were the fault zones originally high angle and planar or nearly so and subsequently rotated with adjacent strata to low-angle positions, domino style (Morton and Black, 1975; Chamberlin, 1983)? Could the faults have been initiated as comparatively planar low-angle normal faults, i.e. "detachment faults" as defined by Wernicke et al. (1985)?

One way to distinguish between the listric and domino-style models of extensional faulting is by comparing tilts in hanging walls and footwalls of low-angle faults. Listric faulting requires greater rotation of hanging-wall strata relative to footwall blocks (Wernicke and Burchfield, 1982). Although locally this appears to be the case in the East Potrillo Mountains, in general, footwall and hanging-wall blocks exhibit subequal tilts. (This situation doesn't hold where obvious Laramide structure is present in hanging walls or footwalls.) Therefore, it is unlikely that the low-angle normal faults are merely the flat parts of listric faults and that the rotated strata are related to movement on listric faults. "Domino-style" rotation seems to be more consistent with these subequal tilts, and so we need to try to restore original fault geometry by removing the effects of middle to late Tertiary rotation.

Largely because of the extensive Laramide deformation in the range, particularly in the northern and central areas, it is difficult to determine how much stratal rotation is due to middle and late Tertiary extension and how much is inherited from the Laramide. Actually, the Laramide structures (especially in the northern part of the range) appear to require little or no rotation to account for their present geometry. Nevertheless, a pervasive, generally west-southwest, approximately 20^0-25° tilt of strata, particularly in the central and southern parts of the range, suggests that the range was tilted by this amount in middle to late Tertiary time. The most convincing evidence for this is the west-southwest tilt of approximately 20°-25° of middle to upper Tertiary fanglomerate (Tsf) cropping out along the western edge of the East Potrillo horst (geologic map; section Y-Y'). Clearly, west-southwest stratal tilting was concurrent with or followed deposition of these fanglomerates. We believe that this tilting was concurrent with rotation of the normal faults to their present low dip in middle Tertiary time. Little rotation appears to have accompanied latest Tertiary uplift of the East Potrillo horst (see discussion of Late Tertiary high-angle normal faults, p. 23).

Removal of 25° of west-southwest tilting by rotation about a N30°W-trending axis (the average strike of the low-angle fault system and the southwest-tilted rocks) presumably would restore the low-angle fault system to its original geometry. This operation has been performed for all areas of strike and dip control in the fault system, and the result is shown in Fig. 15 (in pocket). Two features are noteworthy: 1) low-angle dips remain along some fault segments and 2) curvilinear trend of the system.

Although restored fault dips range from 60° to 20°, it is significant that major segments of the fault system were apparently initiated at angles of 40° or less. This is illustrated particularly well by the (present) west-dipping, near-bedding-plane fault in the southern part of the range (sections *J-J'* and *K-K'*, Sheet 2), which is the structurally lowest fault exposed in the normal fault system. Removal of 25° of west-southwest tilt leaves this fault dipping approximately 20°E, and the rotation also reduces stratal dips to near horizontal. (As shown in cross sections *1-1'* and *K-K'*, hanging-wall dips may locally exceed footwall tilt by a few degrees.) Because of the low restored hanging-wall dips and little evidence in general for greater rotation of hanging-wall rocks relative to footwall rocks, this fault and others in the system cannot be regarded as flat parts of listric faults. Rather, these faults, as well as major segments of the structurally higher normal faults farther north, appear to have originated as low-angle faults.

Generally, the restored fault zone strikes N30°W but bends to N50°W-N60°W strike near point A in Fig. 15. In this reconstruction the present shovel-shaped geometry is far less obvious than it was in Fig. 14. The restored curviplanar fault geometry is probably best interpreted as a large-scale groove or mullion, dipping approximately 40°-45° northward, created by the change in strike of the major fault zone from N30°W to N60°W at point A. In this respect it may be similar to large creases in the East Potrillo fault surface where that fault makes abrupt changes in strike (R. M. Chamberlin, pers. comm. 1985). Rotation of slickenside measurements suggests slip in a northerly direction, and this in turn suggests important left slip on the N30°E-trending fault segments (see section on Direction and amount of extension, p. 23).

As for the zone of northeast-trending, northwest-dipping faults along the eastern flank of the range, restoration of them indicates original strikes of N60°W-N75°W and dips of 45°-50°NE; they are essentially parallel to the main fault zone southeast of point A (Fig. 15). These faults may merge into the main fault zone at depth; the southernmost member of the system appears to do so at the surface. Reverse drag adjacent to these faults is locally well developed, indicating that these faults may flatten as they merge into the main fault zone and are therefore listric in geometry.

In summary, it seems reasonably clear that the normal-fault system in the East Potrillo Mountains originally dipped eastward at shallow to moderate angles, that the structurally lowest fault zone in the system dipped most shallowly (20°-25°), and that the structurally highest zone consisted of a series of short imbricate(?) normal faults, dying out upward and possibly flattening and merging downward with major strands of the system. Other than regional tilt, little or no deformation is present in any exposed footwall rocks below the structurally lowest fault zone. Except locally, little or no differential tilt of hanging wall and footwall accompanied movement on the major faults, suggesting that these fault surfaces are planar or nearly so. However, we recognize that changes in strike created megagrooves or mullions, which appear today as shallowly dipping, shovel-shaped fault surfaces. On a smaller scale, the fault zone contains smaller wavelength mullions as well as complex systems of braiding and anastomosing fault strands of variable strike and dip.

Wernicke et al. (1985) defines normal faults that initiate at low angles as detachment faults. Based on this usage, the low-angle fault system in the East Potrillo Mountains may be described also as a detachment, at least in its early history. Detachments in the Colorado River basin and southern Nevada are further characterized as lower boundary shear zones, above which rocks are severely extended by normal faults and rotation relative to the rocks beneath (Davis et al., 1980; Anderson, 1971). In the East Potrillo Mountains there is little evidence of extension below the structurally lowest fault exposed, but the strata above have been extended by 75-100%. Perhaps the structurally lowest, flattest zones in the East Potrillo system functioned as a detachment whereas the short, imbricate(?), listric(?) faults along the northeastern flank of the range served to extend or dilate allochthonous hanging-wall rocks in the manner described by Wernicke et al. (1985) in the Mormon Mountains, Nevada. On the other hand, detachment faults also are considered to be first-order structures of extensional terranes, exhibiting slip of several miles (Wernicke et al., 1985; Davis et al., 1980; Shackelford, 1980; Reynolds and Spencer, 1985). In the East Potrillo Mountains, total stratigraphic separation across the low-angle fault system is only approximately 486 m, although the unknown total slip may be greater. If the East Potrillo fault system has functioned as a detachment, it does not appear to be comparable in scale to other Basin and Range structures unless it is merely part of a much larger but mostly unexposed system.

Regardless of whether the fault system in the East Potrillo Mountains functioned as a detachment (or part of a detachment) zone or represents a less glamorous system of closely spaced but otherwise ordinary normal faults, it seems clear that during the course of extension the faults as well as adjacent strata were rotated southwesterly approximately 25° about a N30°W-trending horizontal axis. Thus, although the evidence is inconclusive and somewhat contradictory, we conclude that this "domino-style" rotation of initially higher-angle faults accounts for much of the observed low-angle dip of fault surfaces. If basal strands of the system also served originally as low-angle detachment surfaces then both detachment and subsequent "domino-style" rotation collaborated to accommodate regional extension during early phases of rifting in southern New Mexico.

Direction and amount of extension

Regional studies indicate that middle Tertiary ("early-rift") extension in the Rio Grande rift and Basin and Range province was northeast-southwest to east-northeast-west-southwest (Rehrig and Heidrick, 1976; Zoback et al., 1981; Golombek et al., 1983; Price and Henry, 1984; Seager et al., 1984; Aldrich et al., 1986). This direction is consistent with the overall N30°W trend of the low-angle faults in the East Potrillo Mountains but is inconsistent with slip direction (north-northeast to north, after rotation) suggested by four of five sets of slickensides previously described. These slickensides suggest significant left slip on N30°E-trending low-angle fault segments, but no other evidence confirming such strike slip was noted, and the one anomalous slickenside set from a N30°W-trending segment of the main fault zone indicated dip slip. At this point we are unable to choose between fault trend or sparse slickenside data as the most reliable indicator of extension direction in the East Potrillo area.

Middle Tertiary extension in the East Potrillo area was on the order of 75-100%, at least in strata above the structurally lowest fault. This is based on palinspastic reconstruction of cross sections G-G' through I-I' (Sheet 2) and on equations for calculating extension developed by Thompson (1960) and Angelier and Colleta (1983). Similar estimates of middle Tertiary extension have been published by Chamberlin (1983) and Chamberlin and Osburn (1984) for parts of the central Rio Grande rift.

Possible transverse shear zone and oppositely tilted blocks

The poorly exposed, northeast-dipping Upper Cretaceous-lower Tertiary strata adjacent to Mt. Riley and Mt. Cox may be structurally significant. The stratal tilt here appears to be opposite of that in the East Potrillo Mountains. The two oppositely tilted blocks may be separated by a very speculative northeast-trending shear zone shown on the geologic map (Sheet 1) and in Fig. 16. Bending of strata from northwest to northeast strike at the northern edge of the East Potrillo Mountains could be viewed as drag against such a shear zone. Such oppositely tilted domains, separated by northeast-trending shear zones, have been documented in the Socorro area (Chapin et al., 1978; Chamberlin and Osburn, 1984) and in the Basin and Range and elsewhere by Stewart (1980) and Bally (1981). The transverse faults may be thought of as continental transforms (Bally, 1981; Fig. 16). Relationships in the Mt. Riley-Mt. Cox block are not well enough exposed to document further this intriguing but speculative possibility.

Late Tertiary high-angle normal faults

The most recent faults in the East Potrillo area are the boundary faults or fault zones of the modern East Potrillo uplift and adjacent basins. Trending northwest, the East Potrillo fault and Mt. Riley fault zone form the northeastern and southwestern boundaries, respectively, of the uplift (geologic map, Sheet 1). The uplift, which appears to be a

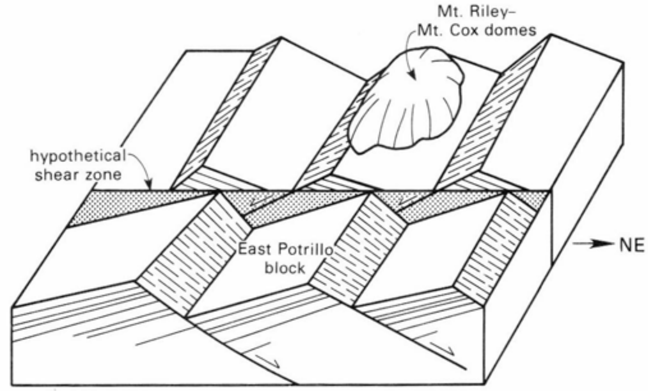

FIGURE 16—Diagram illustrating speculative northeast-trending shear zone separating domains of oppositely tilted blocks in East Potrillo area. After Bally (1981).

rather simple horst, is flanked on the northeast by the broad Mesilla Basin and on the southwest by a poorly known middle to late Tertiary basin. Based on regional studies, Seager et al. (1984) and Chapin and Seager (1975) suggest formation of these blocks as well as other modern ranges and basins in southern New Mexico by movement on high-angle normal faults (such as the East Potrillo fault) beginning about 9 Ma and continuing to the present.

Of the two boundary fault zones, the East Potrillo fault is far more prominent and well known. Its location is flagged by a conspicuous east-facing piedmont scarp, as much as 15 m high, cutting lower to middle Pleistocene Camp Rice fanglomerate and gravel. In the fall of 1984, the fault surface was exposed in a roadside ditch adjacent to the county road that crosses the scarp 2.5 km north of the Mexico-U.S. border. At that location the fault is nearly vertical and cuts caliche at the top of the Camp Rice Formation on the upthrown side and maar-rim tuffs associated with Potrillo maar on the downthrown side. The tuffs have been dragged into a 10°-15°E dip by movement on the fault. These relations indicate movement on the fault within the last 180,000 yrs the age of the maar deposits. Cumulative displacement of hundreds to thousands of meters is suggested by a steep gravity gradient across the fault (Keller and Cordell, 1983) and by drill-hole data (C. Swanberg and **J.** Snyder, pers. comm. 1984). The drilling further indicates that bedrock descends into the Mesilla Basin stepwise on a series of faults rather than by movement on the single East Potrillo fault (J. Snyder, pers. comm. 1984).

By contrast, the Mt. Riley fault zone, located at the western margin of the East Potrillo horst, is far more poorly defined. It seemingly is located 5-6 km west of the East Potrillo Mountains at or near the western edge of bedrock hills of volcanic rock and fanglomerate (geologic map, Sheet 1). These hills presumably are part of the horst. An exposure of the fault or one of the faults in the zone may be the northwest-trending, down-to-the-west fault located 5 km south of Mt. Cox. Other evidence for this fault zone includes:

1) the rather abrupt termination of bedrock hills along the inferred zone
2) the straight eastern edge of "early-rift" fanglomerate (lower Santa Fe Group) outcrops
3) the apparent offset of La Mesa surface along this zone by 10-40 m
4) eastward back tilting of down-faulted parts of La Mesa surface into the fault zone
5) gravity data (Keller and Cordell, 1983) that indicates deepening of a middle to late Tertiary clastic basin west of the fault zone. Exposures of thick "early-

rift" fanglomerates within the fault zone suggest that this basin may be largely of middle Tertiary age. If so, this early basin has been modified by a shallower late Tertiary graben and extensively covered by Quaternary basalt flows.

Unfortunately, there is little unequivocal evidence that would allow us to separate tilting that may have accompanied uplift of the late Tertiary East Potrillo horst from middle Tertiary domino-style rotation of fault blocks. Nevertheless, we presented reasons earlier for concluding that most, if not all, of the observed range tilt was associated with middle Tertiary domino-style rotation. Gravity maps (Keller and Cordell, 1983) also support this interpretation by showing that the late Tertiary East Potrillo horst is relatively symmetrical. Apparently, rotation of the East Potrillo horst has not been sufficient to produce an asymmetric positive gravity anomaly. Although this and other arguments may be equivocal by themselves, we believe that taken together the evidence supports 1) middle Tertiary domino-style rotation of fault blocks and 2) uplift of the late Tertiary East Potrillo horst as an essentially unrotated block.

Economic geology

Dunham (1935) and Bowers (1960) presented summaries of the mineralization in the East Potrillo Mountains. Prospects are widely scattered across the range but in most cases are localized by Laramide or Tertiary low-angle faults. In each digging the mineralization seems to be feeble. Silicification along low-angle fault zones (both thrust and normal faults) or as bed replacement, as well as locally intense brecciation, accompanies much of the mineralization.

Only a few mineral species have been recognized. Spotty malachite with hematite and limonite are common mineral occurrences in prospect dumps. Barite, quartz, pyrite, and calcite are common accessory minerals. Dunham (1935) also reported galena with supergene cerussite, pyromorphite, and malachite at a group of prospects at the northern edge of the range. Dunham (1935) mentioned further that pockets of rich gold ore were said to have been mined from a quartzite bed on the east side of the range approximately 5 km south of the road passing through the gap between the East Potrillo Mountains and Mt. Riley-Mt. Cox.

In the summer of 1983 drilling was completed in several of the silicified outlying hills of Permian limestone just west of the northwestern corner of the East Potrillos. Presumably the aim was to test the precious-metal potential of these outcrops. No follow-up work was done in 1983 or 1984.

In 1962 Pure Oil Company drilled the Pure No. 1 Fed "H" test in the central part of the East Potrillo Mountains at the eastern base of the range. The well was spudded in middle Permian limestone approximately 0.6 km west of the East Potrillo boundary fault. According to Kottlowski et al. (1969):

> . . . Drilling penetrated what appears to be a normal sequence to a depth of 4,200 feet. Gouge-like material was encountered at this depth and drilling continued in similar material to the total depth of 7,346 feet. Below 6,960 feet some diorite is present in the cuttings. It is suggested that the east-boundary fault zone of the East Potrillo Mountains was penetrated somewhere around 4,200 feet and was followed to the total depth.

An important consequence of the interpretation of Kottlowski et al. (1969) is that in order for the well to penetrate the East Potrillo fault zone, the fault must dip west beneath the range and would therefore be a reverse fault. We are convinced, however, that range-boundary faults in southern New Mexico as well as throughout the Rio Grande rift and Basin and Range province are normal faults. Thus, we believe some other fault zone or structure was penetrated by the Pure well. Thompson (1982) indicates a reverse fault was cut by the well at a depth of 4,405 ft, but he does not identify it as a range-boundary fault.

From an examination of cuttings, Thompson (1982) interprets the following down-hole lithologies:

0-625 ft Lower Cretaceous rocks (based on Bowers' [1960] geologic map); the well was actually spudded in middle Permian limestone, so this 625 ft probably is Permian.

625-3,850 ft Permian to Mississippian limestone, dolostone, chert, and mudstone.

3,850-4,060 ft Percha Shale, dark, radioactive mudstone. 4,060-4,405 ft Fusselman Dolomite, light, coarsely crystalline dolostone.

4,405-6,240 ft Repeat of Permian to Mississippian section.

6,240-6,815 ft Marbleized Paleozoic rocks.

6,815-TD Tertiary diorite.

Drill-stem tests at 4,380-4,400 ft revealed a slight show of gas in drill mud, and another at 4,354-4,412 recovered 3,480 ft of mud-cut water.

The middle Permian limestone and dolomitic limestone in the East Potrillo area (Yeso-San Andres and Colina-Epitaph equivalents) may be good hydrocarbon source beds. They appear to be thick, probably more than 300 m, and are generally dark and fetid, locally carrying abundant fossil debris. However, they are decidedly nonporous except for locally extensive fracture porosity. Although Dunham (1935) recognized these rocks as probably Pennsylvanian or Permian, they have generally been regarded as Lower Permian (Hueco) since Bowers' (1960) study. For reasons described earlier, we believe the strata are middle Permian. The presence of a thick middle Permian section with source-rock potential coupled with Laramide structures, whether of overthrust or block-uplift and basin style, present a heretofore unrecognized petroleum target in this region.

Finally, in the past, marbleized Permian limestone has been quarried in the southern part of the range.

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Selected conversion factors*

TO CONVERT	MULTIPLY BY	TO OBTAIN
Length		
inches, in	2.540	centimeters, cm
feet, ft	3.048×10^{-1}	meters, m
yards, yds	9.144×10^{-1}	m
statute miles, mi	1.609	kilometers, km
fathoms	1.829	m
angstroms, Å	1.0×10^{-8}	cm
Å	1.0×10^{-4}	micrometers, μm
Area		
in^2	6.452	cm^2
ft^2	9.29×10^{-2}	m^2
yds^2	8.361×10^{-1}	m^2
mi^2	2.590	km^2
acres	4.047×10^{3}	m^2
acres	4.047×10^{-1}	hectares, ha
Volume (wet and dry)		
in^3	1.639×10^{1}	cm^3
ft^3	2.832×10^{-2}	m^3
yds^3	7.646×10^{-1}	m^3
fluid ounces	2.957×10^{-2}	liters, l *or* L
quarts	9.463×10^{-1}	l
U.S. gallons, gal	3.785	l
U.S. gal	3.785×10^{-3}	m^3
acre-ft	1.234×10^{3}	m^3
barrels (oil), bbl	1.589×10^{-1}	m^3
Weight, mass		
ounces avoirdupois, avdp	2.8349×10^{1}	grams, gr
troy ounces, oz	3.1103×10^{1}	gr
pounds, lb	4.536×10^{-1}	kilograms, kg
long tons	1.016	metric tons, mt
short tons	9.078×10^{-1}	mt
oz mt^{-1}	3.43×10^{1}	parts per million, ppm
Velocity		
ft sec^{-1} (= ft/sec)	3.048×10^{-1}	m sec^{-1} (= m/sec)
mi hr^{-1}	1.6093	km hr^{-1}
mi hr^{-1}	4.470×10^{-1}	m sec^{-1}

TO CONVERT	MULTIPLY BY	TO OBTAIN
Pressure, stress		
lb in^{-2} (= lb/in^2), psi	7.03×10^{-2}	kg cm^{-2} (= kg/cm^2)
lb in^{-2}	6.804×10^{-2}	atmospheres, atm
lb in^{-2}	6.895×10^{3}	newtons (N)/m^2, N m^{-2}
atm	1.0333	kg cm^{-2}
atm	7.6×10^{2}	mm of Hg (at 0° C)
inches of Hg (at 0° C)	3.453×10^{-2}	kg cm^{-2}
bars, b	1.020	kg cm^{-2}
b	1.0×10^{6}	dynes cm^{-2}
b	9.869×10^{-1}	atm
b	1.0×10^{-1}	megapascals, MPa
Density		
lb in^{-3} (= lb/in^3)	2.768×10^{1}	gr cm^{-3} (= gr/cm^3)
Viscosity		
poises	1.0	gr cm^{-1} sec^{-1} *or* dynes cm^{-2}
Discharge		
U.S. gal min^{-1}, gpm	6.308×10^{-2}	l sec^{-1}
gpm	6.308×10^{-5}	m^3 sec^{-1}
ft^3 sec^{-1}	2.832×10^{-2}	m^3 sec^{-1}
Hydraulic conductivity		
U.S. gal day^{-1} ft^{-2}	4.720×10^{-7}	m sec^{-1}
Permeability		
darcies	9.870×10^{-13}	m^2
Transmissivity		
U.S. gal day^{-1} ft^{-1}	1.438×10^{-7}	m^2 sec^{-1}
U.S. gal min^{-1} ft^{-1}	2.072×10^{-1}	l sec^{-1} m^{-1}
Magnetic field intensity		
gausses	1.0×10^{5}	gammas
Energy, heat		
British thermal units, BTU	2.52×10^{-1}	calories, cal
BTU	1.0758×10^{2}	kilogram-meters, kgm
BTU lb^{-1}	5.56×10^{-1}	cal kg^{-1}
Temperature		
°C + 273	1.0	°K (Kelvin)
°C + 17.78	1.8	°F (Fahrenheit)
°F − 32	5/9	°C (Celsius)

**Divide by* the factor number to reverse conversions.
Exponents: for example 4.047×10^{3} (see acres) = 4,047; 9.29×10^{-2} (see ft^2) = 0.0929.

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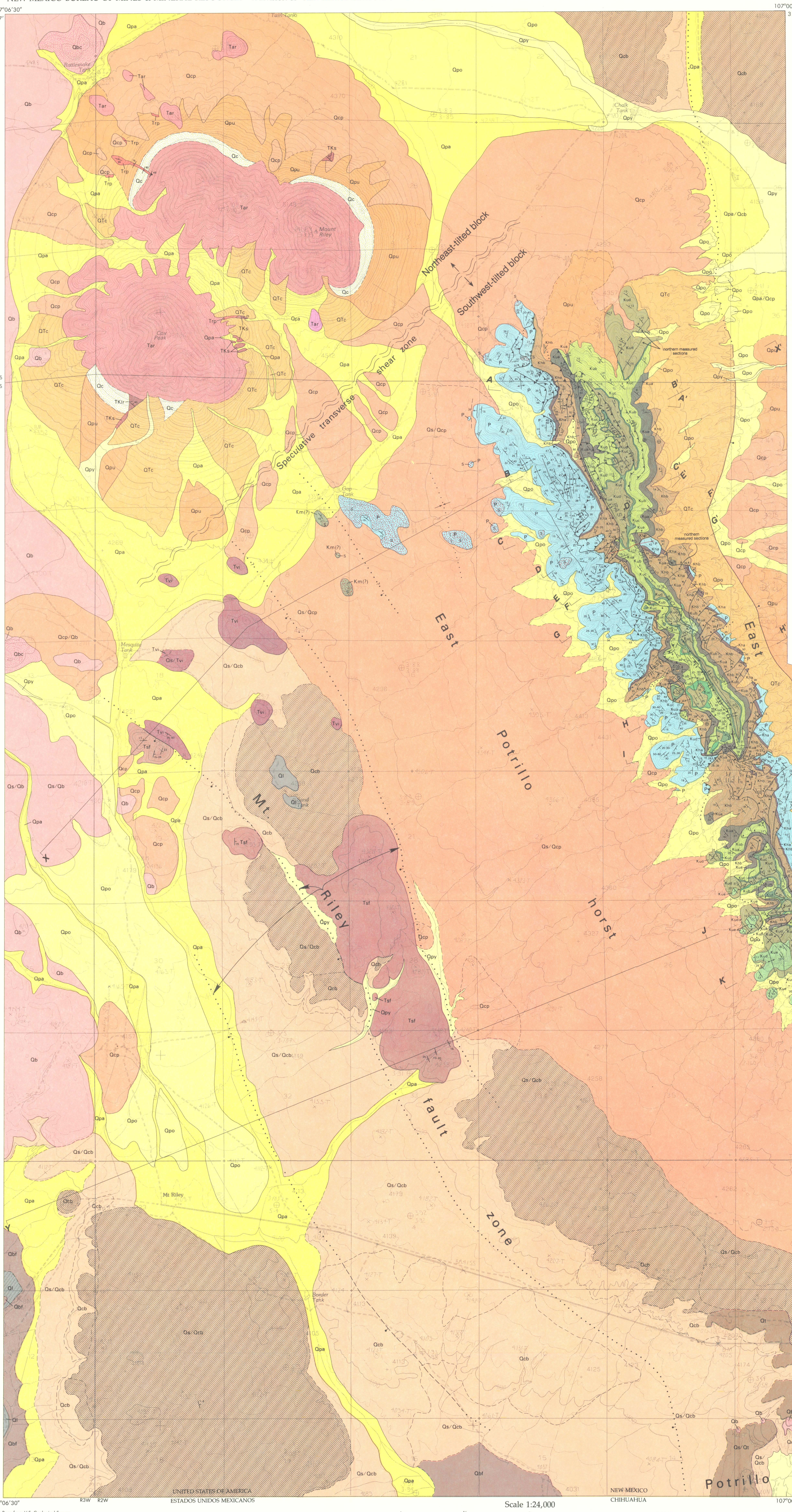
Quantity: 1,000

107°06′30″ 31°57′ — 107°00′ 31°57′



31°47′02″ 107°06′30″ R3W R2W — 107°00′ R2W R1W — 106°58′08″ 31°47′02″

Base from U.S. Geological Survey
Geology by W. R. Seager, 1982–1984

Scale 1:24,000

1 ½ 0 1 MILE

1 .5 0 1 KILOMETER

Geologic color separations and layout by C. A. Salisbury, R. J. Titus, and M. W. Wooldridge

Edited by J. C. Love

CORRELATION OF UNITS

	Piedmont slope	Basin floor	Volcanic rocks
Holocene and upper Pleistocene	Qpa, Qpy, Qs, Qc, Qpu	Qbf, Ql, Qs	
Pleistocene	Qpo; unconformity; Qcp	unconformity	Qt; Qb/Qbc
Lower Pleistocene upper Pliocene	QTc	Qcb; unconformity	
Miocene		Tsf	
Oligocene		Tvi, Tar, Ta	Tr
Oligocene and upper Eocene		Trp	
Lower Tertiary and Upper Cretaceous		TKs, TKlr; unconformity	
Lower Cretaceous		Km(?), Kuf, Kue, Kud, Kuc, Kub, Kua, Khb, Kha; unconformity	
Middle Permian		P	

DESCRIPTION OF UNITS

Qpy **Younger piedmont-slope deposits**—Unconsolidated sand, gravel, and loamy sediments of drainageways, which cross and are inset below or bury older piedmont surfaces, and of fans constructed on distal slopes at lower end of such drainageways; thickness to 5 m (16.5 ft)

Qs **Eolian sand**—Dunes and hummocks of tan to pale-orange quartz sand, which mantle various rock units, especially Qcp, Qcb, and Qb; thickness to 3 m (10 ft)

Qbf / Ql **Basin-floor sediments (Qbf)**—Mostly loam, silt, or clay, locally with thin pebble zones; thickness to 5 m (16.5 ft). **Playa-lake deposits (Ql)**—Mostly clay and silt in small nonalkaline lakes or depressions; thickness to 5 m (16.5 ft)

Qpo **Older piedmont-slope deposits**—Fan and terrace deposits and erosion-surface veneers on piedmont slopes graded to closed-basin floors; mostly weakly consolidated gravel and sandy gravel with thin surficial and buried horizons of soil-carbonate and clay accumulation; thickness to 10 m (33 ft)

Qpa **Undifferentiated Qpo and Qpy**

Qc **Colluvium**—Slope gravel of Holocene to early Pleistocene age on the middle slopes of Mt. Riley–Mt. Cox; locally cemented with pedogenic carbonate

Qt **Air-fall tuff and breccia of Potrillo maar**—Crossbedded juvenile basaltic lapilli and accretionary lapilli and disaggregated sand and silt from the Camp Rice Formation; angular to rounded upper mantle xenoliths and lower, middle, and upper crustal xenoliths include lherzolite, granulites, gneisses, conglomerate, limestone, and volcanic rocks; thickness to 7 m (23.5 ft)

Qb / Qbc **Alkali-olivine basalt**—Basalt flows (Qb) and cinder cones (Qbc); thickness at least 20 m (65 ft) locally

Qcp **Santa Fe Group, Camp Rice Formation, piedmont-slope facies**—Weakly to moderately cemented boulder to cobble fan deposits and erosion-surface veneers in proximal fan positions grading to gavelly silt and loam on distal piedmont slopes. Surficial layers, as much as 3 m (10 ft) thick, contain prominent horizons of soil-carbonate accumulation that form resistant pedogenic calcrete zones 1–2 m (3–6 ft) thick; thickness generally less than 7 m (24 ft)

QTc **Santa Fe Group, Camp Rice Formation, fanglomerate facies**—Unit comprises Qcp as well as underlying, massive calcite-cemented fanglomerate; oldest beds in this unit may be late Pliocene in age. As much as 15 m (50 ft) exposed; probably considerably thicker locally

Qpu **Undifferentiated Qpo and Qcp**

Qcb **Santa Fe Group, Camp Rice Formation, basin-floor facies**—Loam, silt, clay, and sand in basin-floor position. Sediments represent deposits of alluvial flats, ephemeral nonalkaline lakes, eolian activity, and possible fluvial deposition in braided channels of the ancestral Rio Grande. Constructional top of these deposits is La Mesa surface, shown on the map with diagonal lines. Surficial layers, as much as 3 m (10 ft) thick below La Mesa surface, have multiple horizons of soil-carbonate accumulation, the uppermost of which forms a prominent resistant pedogenic calcrete zone as much as 1.5 m (5 ft) thick. Holocene eolian sand (Qs) discontinuously mantles the unit; exposed thickness approximately 10 m (35 ft)

Tsf **Santa Fe Group, older fanglomerate**—Brown to purplish-brown conglomerate and conglomeratic sandstone derived from volcanic rocks, Cretaceous and Permian rocks of the East Potrillo Mountains, and Precambrian granite; locally silicified; moderately deformed by tilting and faulting; thickness at least a few hundred meters (800 ft)

Tvi **Intermediate-composition volcanic and volcaniclastic rocks**—Gray, purplish to maroon-gray latite, dacite, and andesite flows, intrusives, and breccia; aphyric to slightly porphyritic; very fine grained; unit probably also includes conglomerate, mudstone, and sandstone (derived from intermediate-composition volcanic rocks) exposed on the northeast side of hill 4474, 5 km (3 mi) south–southwest of Mt. Cox; thickness at least 100 m (330 ft)

Trp **Rubio Peak Formation**—Purplish-gray, greenish-gray, and dark-gray porphyritic andesite to latite flows and interbedded breccia and tuffaceous epiclastic rocks; siliceous tuffaceous siltstone as well as limestone-pebble conglomerate near base of unit; thickness at least a few hundred meters (656 ft)

Tar **Andesite porphyry of Mt. Riley–Mt. Cox**—Pale greenish-gray andesite–rhyodacite; sparse phenocrysts of hornblende, biotite, and plagioclase; nearly structureless; flow banding developed locally along contacts with wall rock. Unit is probably a volcanic dome or cluster of domes

Ta **Andesite dikes of East Potrillo Mountains**—Greenish-gray, sparsely porphyritic dikes approximately 1 m (3 ft) wide; only a few small outcrops noted as well as a single 0.5-km-long (0.3-mi-long) dike

Tr **Rhyolite dike of East Potrillo Mountains**—Cream, nonflow-banded rhyolite

TKs **Upper Cretaceous and/or lower Tertiary sedimentary rocks**—Gray, purplish-gray, maroon, tan, brown, and olive-brown sandstone, shale, siltstone, and pale-red mudstone derived in part from intermediate-composition volcanic rocks; minor conglomeratic sandstone with limestone (Paleozoic–Cretaceous) clasts; outcrops widely scattered and poorly exposed around the margins of the Mt. Riley–Mt. Cox intrusion. Unit is probably at least a few hundred meters (700 ft) thick

TKlr **Love Ranch Formation**—Upper unit of gray, reddish-gray limestone-cobble and boulder fanglomerate with interbedded red shale and siltstone and tan to olive-brown sandstone; clasts are Lower Cretaceous and Permian rocks similar to those of the East Potrillo Mountains; grades downward into lower unit of red mudstone. Each unit is 100 m (330 ft) or more thick; exposed at southwest margin of Mt. Cox

Km(?) **Mojado Formation**—Gray, partly silicified (s), crossbedded orthoquartzite; approximately 15 m (50 ft) exposed

Kuf **U–Bar Formation, upper siltstone member**—Gray and tan-mottled, massive to hummocky-laminated, calcareous siltstone; conformable basal contact; upper contact covered; exposed only in southern part of East Potrillo Mountains; thickness at least 15 m (50 ft)

Kue **U–Bar Formation, massive limestone member**—Light-gray, massive, recrystallized limestone with light streaks of calcite; few relict bivalve and gastropod shells near base; conformable upper and lower contacts; exposed only in southern part of East Potrillo Mountains; thickness 133 m (435 ft)

Kud **U–Bar Formation, siltstone–limestone member**—Tan and gray-mottled, massive to horizontally or hummocky-laminated, bioturbated siltstone and very fine grained sandstone; bivalve and gastropod shells; bivalve and gastropod packstones increase in abundance upward; conformable upper and lower contacts; thickness 81–90 m (268–298 ft)

Kuc **U–Bar Formation, rudistid limestone member**—Gray, ledge-forming limestone; lower half is massive rudistid boundstone; upper half is medium-bedded gastropod and bivalve wackestone; conformable contacts; thickness 6–13 m (20–45 ft)

Kub **U–Bar Formation, sandstone member**—Interbedded (on a scale of 1–5 m [3–17 ft]), brown, ledge-forming, crossbedded, granular to very coarse grained sandstone and granule-pebble conglomerate with bivalve shells and gray, massive to horizontally or hummocky-laminated, bioturbated very fine and fine-grained sandstone with bivalve shells; conformable contacts; thickness from 25 to 66 m (85–220 ft), thinning southward

Kua **U–Bar Formation, lower limestone member**—Gray, thick-bedded, oolitic grainstone with scattered large bivalve and gastropod shells in northern part of range; silty bivalve and gastropod grainstone in southern part of range; conformable contacts; thickness from 13 to 33 m (45–110 ft), thinning southward

Khb **Hell-to-Finish Formation, mottled siltstone member**—Tan, gray-mottled calcareous siltstone and very fine grained sandstone with horizontal and hummocky laminations and bioturbation; predominantly ledge forming with covered intervals; six 0.5-m-thick (1.5-ft-thick) beds of siltstone-, limestone-, and chert-pebble conglomerate and one 6-m-thick (19.5-ft-thick) bed of crossbedded, granular, coarse-grained sandstone with bivalve and gastropod shells; conformable contacts; thickness from 119 to 168 m (400–550 ft), thinning southward

Kha **Hell-to-Finish Formation, conglomerate member**—Gray, massive, grain-supported cobble and boulder conglomerate with medium-grained sand matrix; clasts are well-rounded chert and limestone derived from underlying middle Permian section; conformable upper contact, unconformable lower contact; thickness from 1 to 40 m (3–133 ft)

P **Yeso–San Andres Formations**—Gray, cream, olive-gray, and rusty-gray micrite, echinoid grainstone, and dolomitic limestone in medium to massive beds; local bivalves, rugose corals, and chert; silicified or marbleized locally (s); thickness at least several hundred meters (1,000 ft)

s **Silicified rocks (P, Km?)**

Contact—Dashed where inferred or approximately located; arrow shows direction and amount of dip

Strike and dip of bed

Strike and dip of overturned bed

Horizontal bed

Vertical bed

Strike and dip of foliation

Normal fault—Ball on downthrown side; dashed where inferred, dotted where buried; arrow shows direction and amount of dip; double-headed arrow shows trend and plunge of slickensides

Thrust fault—Teeth on hanging wall; dashed where inferred, dotted where buried; arrow shows direction and amount of dip

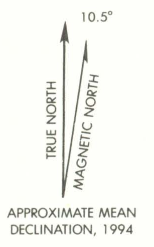
Anticlinal axis

Synclinal axis—Dashed where inferred, dotted where buried

Overturned anticlinal axis

Overturned synclinal axis

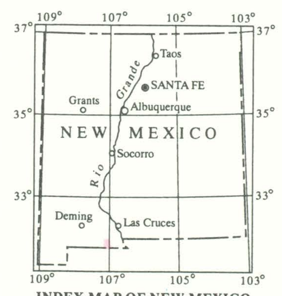
Measured-section locality—See also Sheet 3



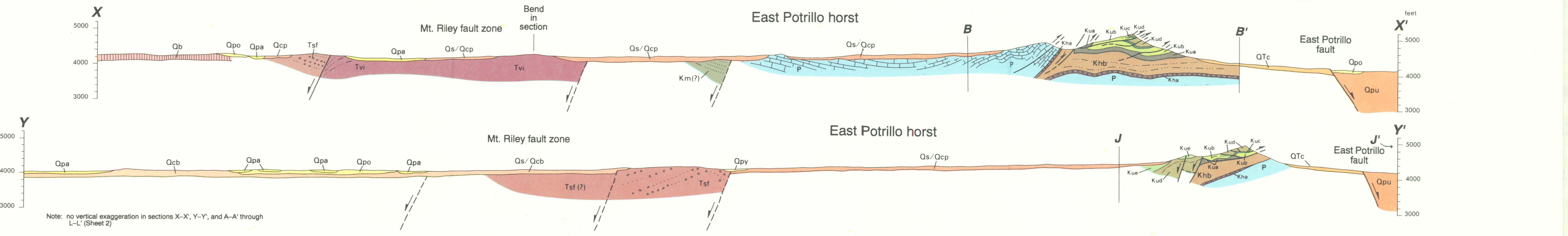
Geologic map of East Potrillo Mountains and vicinity, Doña Ana County, New Mexico

by

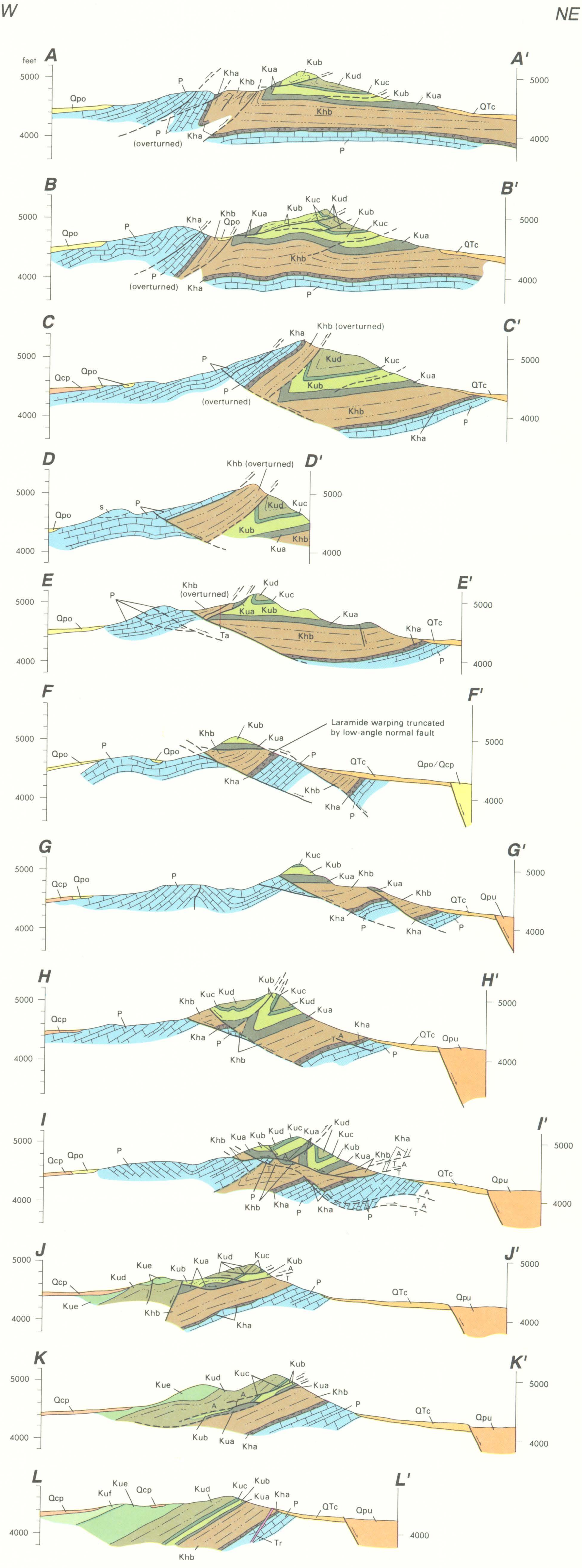
William R. Seager, 1994



INDEX MAP OF NEW MEXICO



Note: no vertical exaggeration in sections X–X′, Y–Y′, and A–A′ through L–L′ (Sheet 2)



Symbols are the same as those used on geologic map
except A and T on faults, which indicate movement away from and toward the viewer.

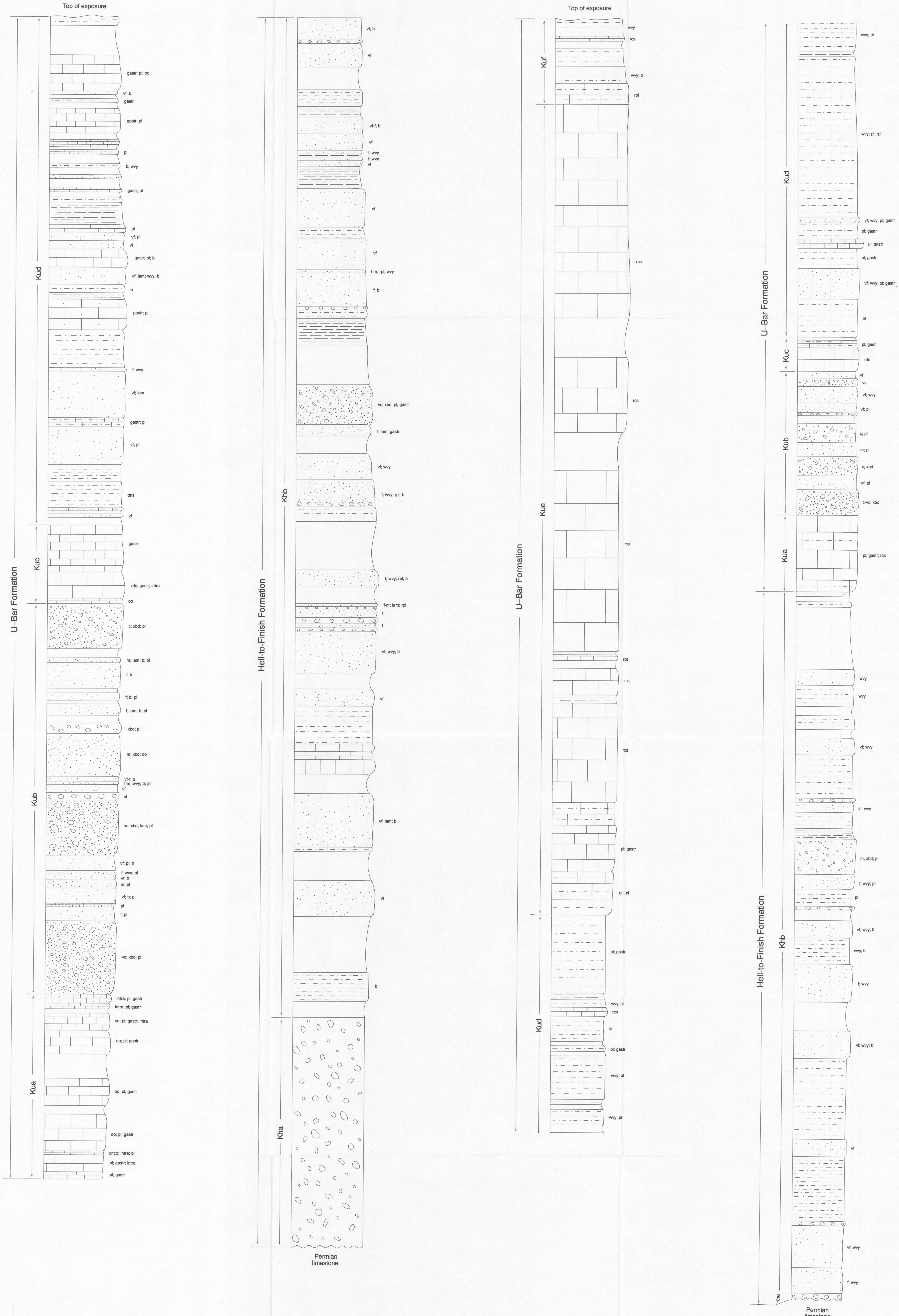
Geologic cross sections of East Potrillo Mountains and vicinity, Doña Ana County, New Mexico

by

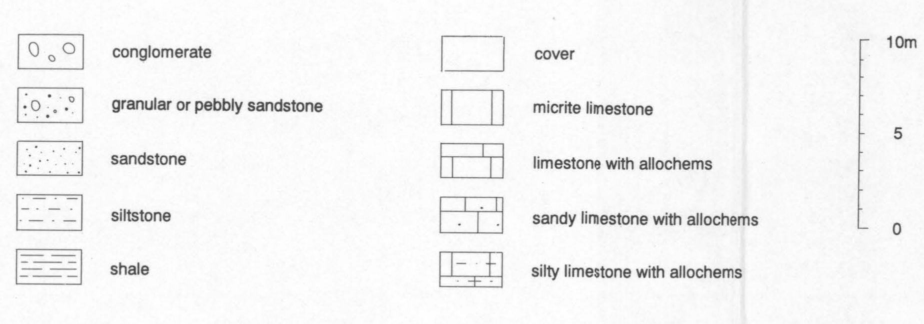
William R. Seager, 1994

Northern part of East Potrillo Mountains

Southern part of East Potrillo Mountains



SYMBOLS

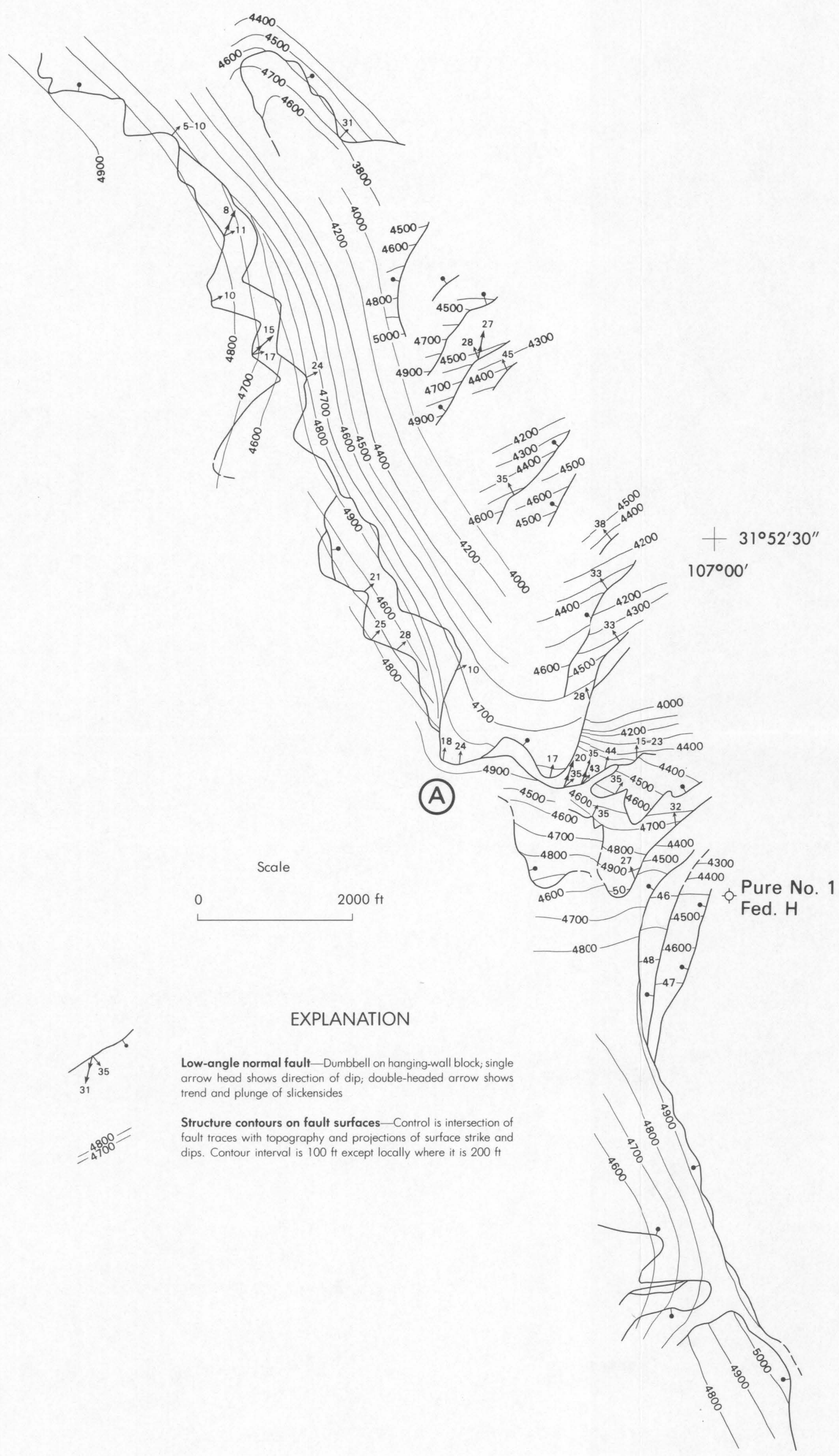


ABBREVIATIONS

vc	very coarse grained sandstone	wvy	irregular, wavy, or hummocky laminations	pl	pelecypods
c	coarse-grained sandstone	xbd	crossbeds	gastr	gastropods
m	medium-grained sandstone	rpl	ripple marks or ripple cross-laminations	rds	rudistids
f	fine-grained sandstone	intra	intraclasts	rcs	recrystallized (neosparite)
vf	very fine grained sandstone	oo	oolites	dns	dense
b	burrows, bioturbation	onco	oncolites		
lam	horizontal laminations				

Lower Cretaceous rocks, northern and southern parts of East Potrillo Mountains, Doña Ana County, New Mexico

by
Greg H. Mack and William R. Seager, 1994

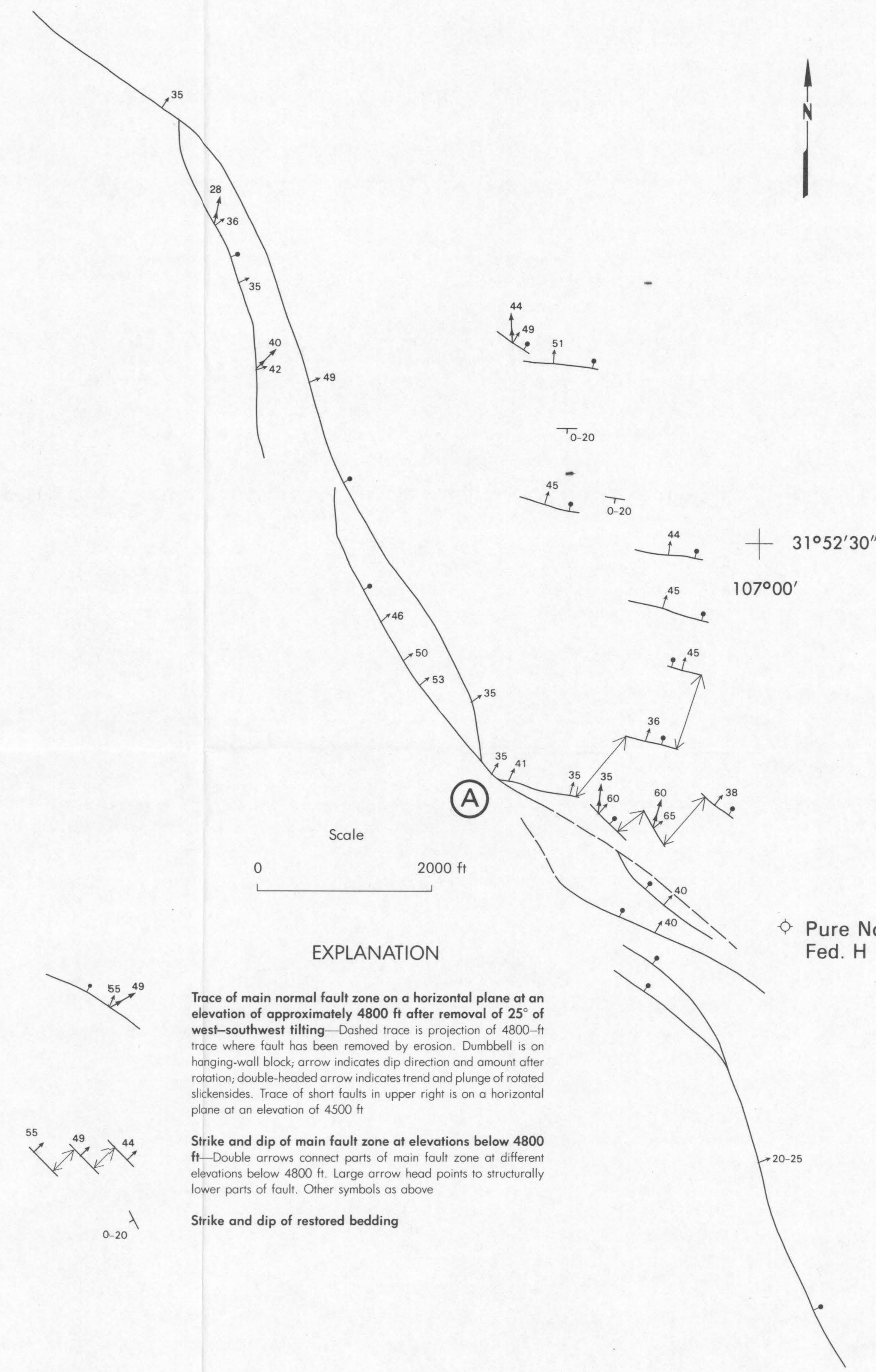


EXPLANATION

Low-angle normal fault—Dumbbell on hanging-wall block; single arrow head shows direction of dip; double-headed arrow shows trend and plunge of slickensides

Structure contours on fault surfaces—Control is intersection of fault traces with topography and projections of surface strike and dips. Contour interval is 100 ft except locally where it is 200 ft

Map of low-angle normal faults
in central part of East Potrillo Mountains
shows structure contours on fault surfaces



EXPLANATION

Trace of main normal fault zone on a horizontal plane at an elevation of approximately 4800 ft after removal of 25° of west–southwest tilting—Dashed trace is projection of 4800–ft trace where fault has been removed by erosion. Dumbbell is on hanging-wall block; arrow indicates dip direction and amount after rotation; double-headed arrow indicates trend and plunge of rotated slickensides. Trace of short faults in upper right is on a horizontal plane at an elevation of 4500 ft

Strike and dip of main fault zone at elevations below 4800 ft—Double arrows connect parts of main fault zone at different elevations below 4800 ft. Large arrow head points to structurally lower parts of fault. Other symbols as above

Strike and dip of restored bedding

Normal-fault patterns
in central East Potrillo Mountains
after removal of 25° of west–southwest tilt