Description and Development of the Cordilleran Orogenic Belt in the Southwestern United States and Northern Mexico

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By HARALD DREWES

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A tectonic synthesis of systematic changes in style and age of deformation; direction, amount, and rate of tectonic transport, and interaction with magmatism and sedimentation

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By HARALD DREWES

ABSTRACT

The Cordilleran orogenic belt forms a continuous zone of compressive deformation of Late Jurassic to Eocene age resulting from collision of the converging North American plate with northeast Pacific plates. The segment of the belt between El Paso, Texas, and Las Vegas, Nevada, differs from adjacent segments in that subordinate tectonic zones are truncated along an unusual kind of frontal zone. This difference is believed to result from a combination of oblique collision of the deformation zones with the semicratonic Colorado Plateau block, the effects of pre-orogenic basement faults along the southwest margin of that block, southward tilting and deep erosion along this frontal segment, and overprinting of extensional tectonism.

The style of deformation across this segment of the orogenic belt varies systematically. In the fold-and-thrust zone (northern Chihuahua, Mexico), folds are large, open to moderately tight, and only slightly inclined, but disharmonic, over (that is, truncated against) genetically related thrust faults and small backthrusts in Permian and Mesozoic rocks. In an eastern intermediate zone (southwestern New Mexico and southeasternmost Arizona), folds are smaller, tighter, and more markedly inclined northeastward toward related shingled thrust faults and larger backthrusts. Likely, detachment occurred along the top of the crystalline basement of Precambrian age. In a western intermediate zone (much of southeastern and central southern Arizona), many folds are also small and are strongly inclined northeastward toward related thrust faults, but some erosional remnants of larger folds involve major thrust plates. Locally, crystalline basement rocks occur in these major plates. In the hinterland zone, many folds are recumbent, and thrust faults are numerous and commonly are parallel to bedding in ductily deformed thin plates of Phanerozoic rock and to foliation in massive plates of basement rock.

Areas of mild deformation of a foreland zone lie northeast of, and beneath the fold-and-thrust zone. They are viewed as being external to the main part of the orogenic belt because their faults probably are not continuous with those of the orogenic belt proper in the level of relatively brittle deformation. This zone extends across the Colorado Plateau and east to the Rocky Mountain front. The style and age of deformation in this zone and that of the fold-and-thrust zone are most typically the Laramide phase of the orogeny.

The record of orogenic sedimentation suggests that a series of northwest-trending foreland basins developed northeast of the advancing and younging orogenic front. Much of the black shale, siltstone, and subgraywacke of these basins was probably derived from the volcanic cap or arc terrane of the active part of the belt. This voluminous detritus was first transported eastward subaerially to the basin axes and then southeastward along depocenter axes. More local sources of detritus contributed coarser clastic components, such as quartz sand, gravel, and even some landslide deposits. The coarsest of these deposits may have been shed from local uplifts adjacent to paleo-foreland zone structures, which were subsequently overprinted by styles of deformation characteristic of the fold-and-thrust zone (and?) or intermediate zones. Older basin deposits were cannibalized to help fill newer ones.

The record of igneous activity is closely tied to that of tectonism. Both plutonic and volcanic rocks are younger to the northeast, although there is considerable range in the age of plutonism, especially where the orogenic time was more protracted, to the southwest. This activity dwindled to the northeast, was most abundant (or is best preserved) and most varied in the intermediate zones, and its record is considerably eroded or covered to the southwest or west. Andesitic volcanic rocks are the most voluminous and widespread; rhyolitic ash-flow and airfall tuffs usually cap them but are not found east of Arizona. Plutonic rocks intrude all rocks as young as the orogenic andesitic suite. Metamorphic aureoles are narrow to the east; they widen and seem to merge as a semiregional metamorphic terrane rarely of more than greenschist facies in the western half of the western intermediate zone. Metamorphism in the hinterland zone is of greenschist or lower amphibolite grade, except locally to the far west where some terranes reached the upper amphibolite grade. Major mineral deposits of Cordilleran orogeny age were optimally emplaced and preserved in the western intermediate zone.

The age of deformation along the Texas-California transect was older and longer lived to the west, and was younger and briefer to the east. This time-transgressive characteristic of orogenesis reflects the quasi-ductile nature of rocks, through which the compressional stress propagated northeastward at about 1 cm per year between the Late Jurassic and early Eocene. Where the orogeny was longest lived, 3 phases of deformation occur, the oldest possibly correlative with the Sevier phase of the Cordilleran orogeny. Two tectonic phases are

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recognized in the intermediate zones, and only one is known in the foreland zone and the subfold-and-thrust belt terrane.

Tectonic transport was eastward near Las Vegas during the main or intermediate of the 3 phases, and tectonic transport gradually shifted northeast near El Paso. Diverse transport directions are recorded for some deformation phases, for some local stress fields near reactivated segments of northwest-trending, strike-slip, basement faults, and for backthrusts. There is some evidence either that *sc* (from French words for cleavage (*s*cission) and schist (*s*ciste)) fabric is an inconsistently reliable indicator of transport direction or that irregular ductile flowage patterns may develop. The transport amount was largest in the hinterland, was possibly about 200 km in the western intermediate zone, and dwindled to a few tens of kilometers near El Paso. If a transport amount of 200 km is accepted, the average rate of tectonic transport near Tucson was about 2.5 cm per year.

During peak deformation time, the thickness of cover was 8–10 km in the intermediate zones and much of the hinterland zone. The cover thinned to 3–4 km near El Paso, and thickened to 12–15 km nearest to Las Vegas.

The hypothesis that the Cordilleran orogenic belt is continuous between El Paso and Las Vegas is amply supported by these new data on age, direction of movement, and environment of deformation, obtained from a much broader base than was originally available. My earlier hypothesis must be modified, however, to show that features of tectonic style were not uniform but changed systematically across the belt. No broad alternative synthesis has been proposed; the one available hypothesis of deformation of local extent is found wanting.

Identification of the continuity of the orogenic belt between El Paso and Las Vegas, together with explanations for the irregularities along the northeastern margin of this segment of the belt, permit construction of a comprehensive synthesis for the region. This synthesis emphasizes the interrelations among tectonism, sedimentation, and magmatism. These interrelations are first shown through a series of idealized profiles and then are transposed into a series of palinspastically controlled paleogeographic maps. During an early stage of development the orogenic front extended nearly straight south from Las Vegas through southeastern California. A volcanic terrane to the west probably overlay (by many kilometers) the train of Late Jurassic and Early Cretaceous batholiths, located some 70 km or more west of the present site of the Colorado River, and possibly marking an orogenic core. A foreland basin in western Arizona accumulated the thick and possibly once extensive McCoy Mountain Formation and other units.

The northern part of this region first responded to a compressive stress derived from the northwest, as determined from orientation of oldest folds and *sc* fabric, probably the Sevier phase of the orogeny. The entire belt then was affected by a stress field propagating eastward (and in the more southern part of this region, east-northeastward), in response to which all systems (sedimentation, deformation, magmatism, and so on) likewise shifted eastward. Where the western salient of the semicratonic Colorado Plateau block was encountered, the stress trajectory was deflected to the northeast on the south side of this buttress. By late Early Cretaceous time, the western basin deposits and their overlying andesitic volcanic rocks were being cannibalized and redeposited as the Bisbee Formation in the fluvial and marine basin of central southern Arizona.

With continued east-northeastward propagation of the compressive stress, the Bisbee depocenter was also partly destroyed. The earliest signs of the demise of this basin probably showed up as the local uplifts near reverse faults of a paleo-foreland zone. An upland was also growing southeastward along the present Mogollon rim, thereby keeping the successor foreland basins tilted south or southeast and providing other sources for coarse-grained detritus interbedded in the siltstone and shale.

By Late Cretaceous time, the Bisbee rocks and their overlying volcanic deposits were, in turn, being cannibalized, to be redeposited farther east in such smaller and more continental basins as that of the Ringbone Formation and early Tertiary conglomerates, such as the Lobo Formation. During the Paleocene the older members of the 2-mica garnet-bearing granite were emplaced across Arizona at a depth of 8–10 km, where the first arrivals received a mild tectonic stamp from the last of the orogenic movement.

The orogenic process died out near El Paso by late Paleocene to early Eocene time. A fold-and-thrust zone developed against, or was deflected by, a final obstacle in the form of a northwest-trending fault block northeast of the Texas lineament. Residual stress was dissipated northward and eastward during early Eocene time, to produce scattered local folds and faults of the foreland zone and terminated the youngest orogenic basins of deposition, such as the Love Ranch basin (or basins). Continued uplift along the paleo-Mogollon uplift at least as far southeast as the Burro uplift, and possibly to the Diablo Platform, led to deep erosion of the northeastern edge of the orogenic belt, and ultimately to the stripping back of part of the frontal line across southwestern New Mexico.

INTRODUCTION

The Cordilleran orogenic belt is one of the two largest orogenic belts in the world (fig. 1). It extends several

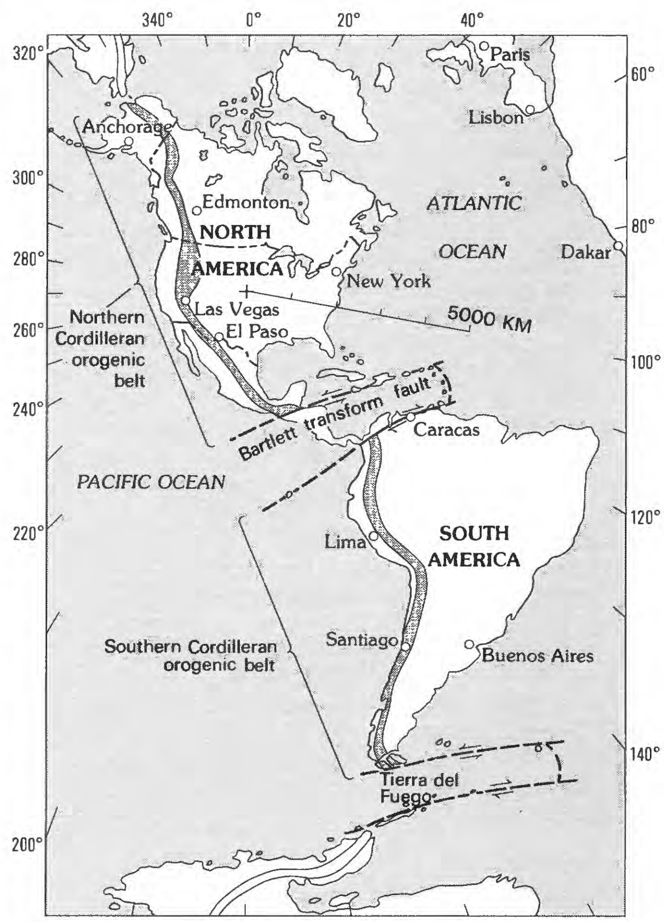


FIGURE 1.—Map showing full extent of the Cordilleran orogen, shown stippled except for its likely extensions slightly beyond North and South America. (Based on an azimuthal equidistant projection centered on 40° N., 100° W., National Map Atlas, p. 329.)

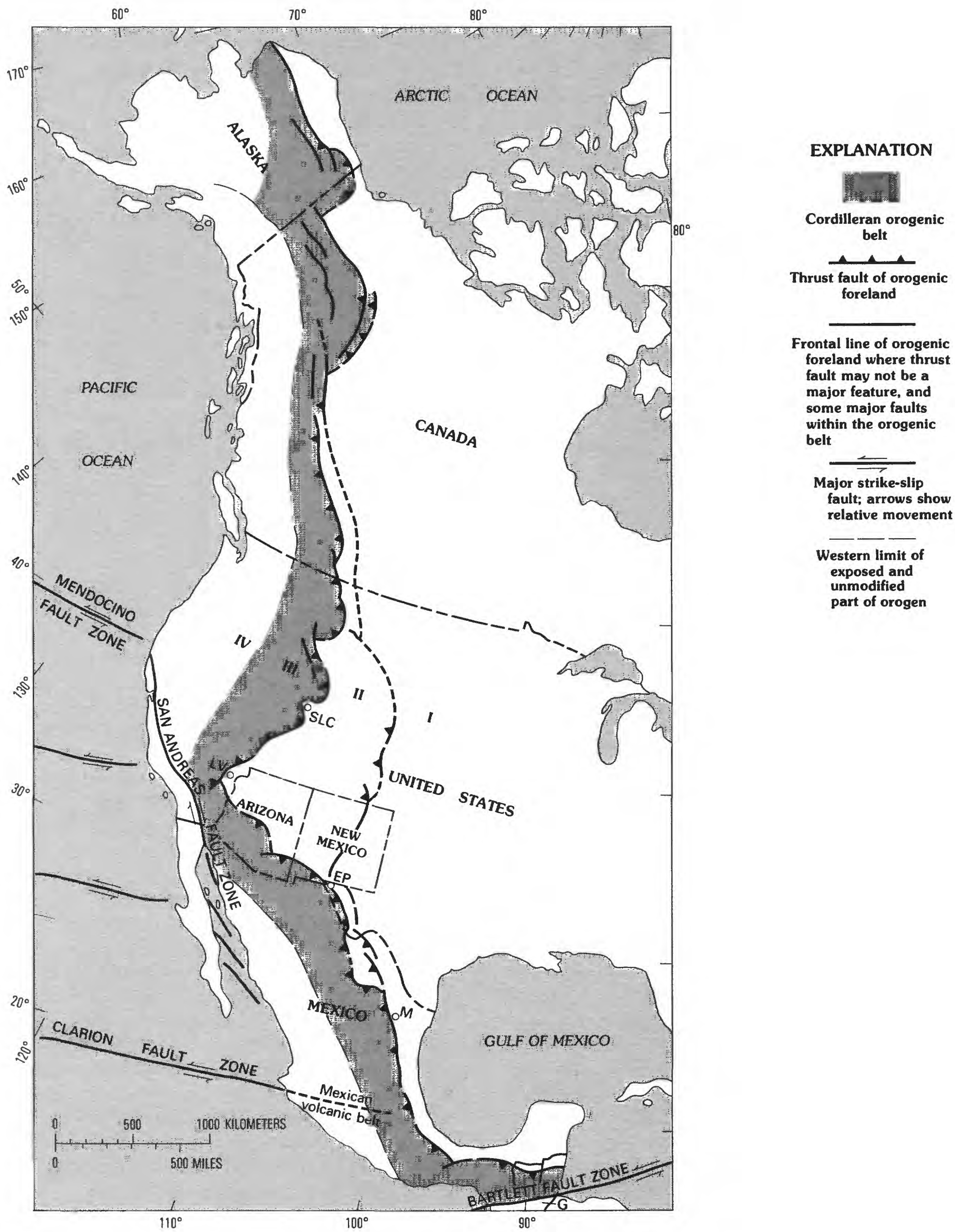


FIGURE 2.— Map of North America showing distribution of the Cordilleran orogenic belt. Modified after King (1969) and King and Edmonston (1972). Region I, craton (includes some pre-Cordilleran eastern terranes and post-Cordilleran cover); II, Colorado Plateau and Rocky Mountain (semi-cratonic region); III, Cordilleran orogenic belt; IV, terrane heavily covered by post-orogenic deposits or modified by post-orogenic tectonic events. Sites: SLC, Salt Lake City; LV, Las Vegas; EP, El Paso; M, Monterrey; and G, Guatemala.

tens of thousands of kilometers from at least northern Alaska to the Palmer Peninsula of Antarctica. Rocks along this belt are compressively deformed, massively intruded by magmas of acid to intermediate composition, and extensively covered by volcanic rocks. Vertical movements along and near this belt have left a distinctive record of erosion and deposition. The stresses and magmas that produced this vast orogenic belt were caused by the convergence of two groups of global plates, the one group driven eastward from the mid-Pacific spreading center and the other group driven westward from the mid-Atlantic spreading center.

The geologic history of such a large orogenic belt as the Cordilleran, varies considerably from region to region. Major divisions of the orogenic belt are marked by systems of transform faults between which the orogenic belt has been offset far to the east, once in southern Central America, and also south of Tierra del Fuego. That division of the orogenic belt northwest of the Bartlett transform fault of Guatemala is the North American or northern part of the Cordilleran belt, although for convenience it will herein be called simply the Cordilleran orogenic belt (fig. 2).

The present study is confined to that segment of the Cordilleran orogenic belt in the Southwestern United States between El Paso, Texas, and Las Vegas, Nevada (fig. 3). This segment is noted for a tectonic pattern atypical of that of much of the rest of the belt (King and Edmonston, 1972). A distinct foreland basin, for example, appears to be missing. A fold-and-thrust zone is also seemingly absent. Thrust faults are discontinuous and seem to have had a diverse geologic history. The thickness of cover rocks has been thought to be less than is common to other parts of the belt and perhaps the rocks were too weak to transmit stresses adequately.

Some of these atypical features and concerns are real; others are only apparent. The challenge is to explain the real differences in the light of the total development of this segment of the orogenic belt. The apparent differences need to be examined through deciphering the geologic record, primarily by detailed field mapping but with laboratory support, particularly for geologic dating. In this synthesis I provide geologic descriptions and consequent inferences that demonstrate a compatibility of time and style of Cordilleran deformation between the El Paso–Las Vegas segment and other segments of the orogenic belt. In other words, no gap exists in the continuity of the North American Cordilleran orogenic belt; rather, the characteristics of the belt change because the overriding rock mass obliquely crossed a part of cratonic margin that had a preexisting strongly developed structural grain, and because a subsequent tectonic event masked many of the Cordilleran features.

Field studies conducted over 24 years form the bulk of this tectonic synthesis. In these studies that were strongly oriented toward geologic mapping, I examined at least parts of 25 mountain ranges. Resolutions to structural and stratigraphic problems are based on field-generated solutions, rather than on broadly regional conceptual models, and as such constitute the support for my conclusions. However, a sharp distinction between these field-generated and regional conceptual models is not possible, nor is it, indeed, desirable. The very process of making a comparison between the development of the enigmatic segment of the orogenic belt under study and the development of another segment of the belt that already has been studied is tantamount to applying a model of what the belt may be like. Thus, my synthesis, while largely field oriented, also benefited from interaction with model-based approaches. Perhaps as a consequence of this operating style the synthesis grew slowly. For example, it took 10 years to marshall enough support for the initial proposal that the low-angle faults of southeastern Arizona were mainly regional thrust faults, rather than entirely local gravity faults. Finally, with support of a Gilbert Fellowship from the U.S. Geological Survey, the study was extended from the core area of parts of Pima, Santa Cruz, and Cochise Counties, Arizona, and Hidalgo, Grant, and Luna Counties, New Mexico, east to the Rio Grande and west to the Colorado River. Colleagues, students, and numerous published reports have contributed to my investigation and are acknowledged in the appropriate sections of the text.

Technical reviews were made by Mitchell Reynolds during 1987 and by C. H. Thorman during 1988, with additional help from Thomas Kohnen in 1989. Their patience and understanding has lead to many improvements and is gratefully acknowledged. The work in Chihuahua was done in conjunction with help from the field camp operation of the University of Texas at El Paso, then under Professor Russ Dyer.

In this synthesis I first review the main features of the geologic setting that bear on the development of the Cordilleran orogenic belt. Then, as the principal part of the synthesis, I describe the structural styles of five tectonic zones, four of them comprising the main part of the orogenic belt, and one that is in front of the belt. These structural styles change systematically among the tectonic zones. Data, such as the age of deformation, the thickness of cover beneath which deformation occurred, the amount and direction of tectonic transport, the relation among tectonic features, sedimentation, and magmatism, and the characteristics of the frontal line along the southwest margin of the Colorado Plateau where the tectonic zones are variously disrupted and eroded, are essential in the presentation. From the structural styles and tectonic data, I conclude by developing a tectonic evolution of the region through a series of structures sections generalized to fit any segment of the belt, and through a series of palinspastically controlled maps of the El Paso–Las Vegas region.

GEOLOGIC SETTING

Several major pre-orogenic structural features had a strong influence on the geologic development of the El Paso–Las Vegas segment of the Cordilleran orogenic belt, and other post-orogenic features modified the belt in this region (fig. 3). The orogenic belt generally follows the western or southwestern side of the North American craton, which is cut by a set of northwest-trending basement faults. The Texas lineament probably is one of these basement faults, reactivated, perhaps diversely, at various localities (Kelly, 1955; Albritton and Smith, 1975). The Mojave-Sonoran megashear is a somewhat similar but more nebulous feature proposed to cross northern Mexico (Silver and Anderson, 1974). The Cordilleran orogenic belt was developed on this deformed craton margin, in part along the margin itself, but extending well beyond the craton, mostly along the Pacific side of North America. Subsequently, the basin-and-range features and the Rio Grande graben were developed largely on the terrane already deformed, to disrupt the continuity of the older structures of the El Paso–Las Vegas segment of the orogenic belt.

Probably the chief factor to generate the character of this segment of the orogenic belt is its position along the southwest margin of the North American craton. The continental craton adjacent to the belt is characterized by a crustal thickness of about 60 km. In contrast, the mobile belt to the southwest has a crustal thickness of only about 30 km. Between these two terranes is the Colorado Plateau–Rocky Mountain block (fig. 3), having a crustal thickness of 50 km, most nearly like that of the craton, but having a somewhat more mobile tectonic history than the craton. The southwestern part of this large block, which, henceforth, will simply be referred to as the Colorado Plateau block, acted as a semirigid salient around which the more mobile orogen was wrapped. The Colorado Plateau block also had a history of development somewhat independent of that of the craton. Whereas the major frontal line of the mobile belt followed the southwestern and western sides of the Colorado Plateau–Rocky Mountains block, a minor branch followed the eastern side.

The basement rocks of the El Paso–Las Vegas segment of the belt consist of Proterozoic crystalline rocks, mainly schist, gneiss, and granite plutons dated at 1,650 to 1,450 m.y. old. These rocks are cut by a system of subparallel to anastomosing northwest-trending high-angle faults. A northwest alignment of many of the granite plutons even suggests an earlier origin of this structural grain. Reactivation of segments of many of these faults occurred at various times and in response to various stresses, as illustrated by Drewes (1981a, fig. 2).

Several faults in southeastern Arizona, for example, were initiated during the Proterozoic as left-slip structures with large displacement (Swan, 1975; Drewes, 1981a, 1982b). Some segments of these faults were reactivated as normal faults or as left-slip faults of moderate displacement during Jurassic to mid-Tertiary time. While the early movement was chiefly subhorizontal, some segments had substantial vertical components of movement, too. Some segments of these faults occur in the present ranges; most occur in low terrain, and a few segments separate the high from the low terrains, where rocks of diverse competence were juxtaposed. For example, the Gold Hill–Dividend fault system near Bisbee, Ariz., was activated (reactivated?) as a reverse or normal fault during the Late Jurassic or Early Cretaceous (Ransome, 1904; Hayes and Landis, 1964), and the Sawmill Canyon fault zone 50 km south-southeast of Tucson was reactivated many times between Jurassic and Oligocene times (Drewes, 1972a).

The Texas lineament of Baker (1935) is probably one member of this northwest-trending right-slip fault system (fig. 3). While originally proposed to exist in west Texas, it was subsequently suggested to extend not only between El Paso and Las Vegas, but to tie in with another lineament known as the Walker Lane (see review by Kelly, 1955). This extension of the Texas lineament, however, was never substantiated as a single through-going fault, such as the San Andreas fault system. Currently, the Texas lineament is viewed as part of a broad zone of faults having a varied history of movement (Kelly, 1955).

The northwest-trending fault system may have been a controlling factor in the development of a transition zone between the Colorado Plateau and the Basin and Range province of Arizona. During mid-Tertiary time this zone was physiographically continuous with the plateau; but block faulting genetically related to basin-and-range faulting, possibly reactivating segments of the ancient northwest-trending fault system, separated this transition zone from the plateau. This zone lies along the transition from the 50-km-thick crust to the 30-km-thick crust.

The Mojave-Sonoran megashear is a proposed major strike-slip fault that lies largely southwest of the region here under consideration (Silver and Anderson, 1974; Silver, 1983). The megashear is proposed to be a Jurassic structure that has a vast amount of left slip—similar to a

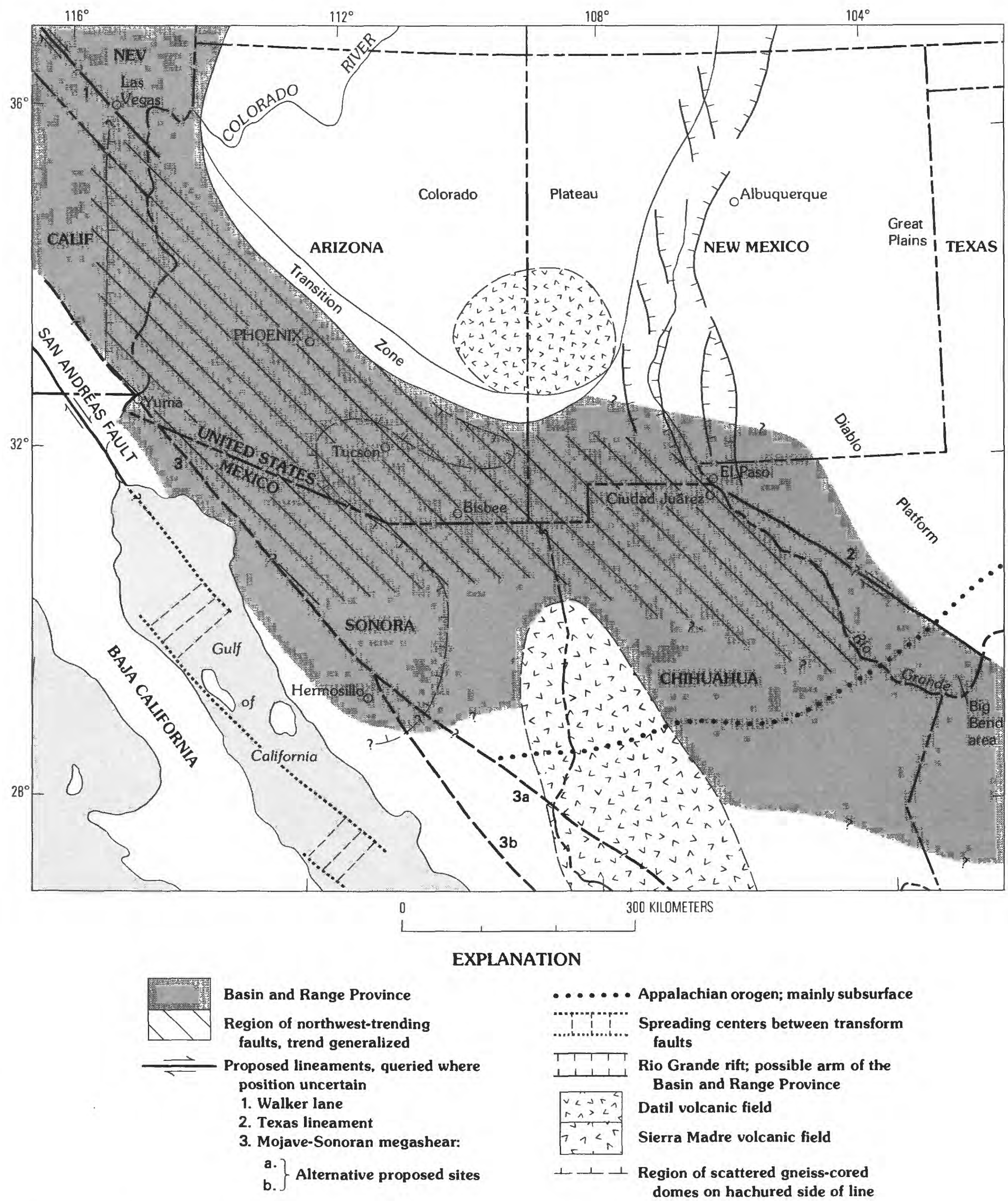


FIGURE 3.—Map of selected geologic terranes and features in the region around the Texas-California transect of the Cordilleran orogenic belt.

transform fault related to a model of plate movements in the Gulf of Mexico far to the southeast. Thus far, the megashear concept is better supported by indirect evidence predicated on how one assumes older rocks may have been distributed before the Jurassic than by direct evidence of any mapped fault. Indeed, the proposed site of this fully concealed fault has been shifted southwestward (fig. 3, from site 3a to site 3b) as field control on the northeastern locations, such as that obtained by Poole and Madrid (1986), presented difficulties in positioning. Should this southwestward shift in the proposed positioning of the megashear continue, it may be possible to merge the megashear with a structure in the Gulf of California, such as the younger San Andreas fault system.

The El Paso–Las Vegas region lies in the southeastern part of the Basin and Range province, a terrane noted for its characteristic array of fault-bounded linear ranges and valleys. These formed in response to tensional stress beginning in late middle Miocene time (Eardley, 1951, p. 481), and perhaps even during Oligocene or Eocene time. This earlier movement is deduced from the fact that in many parts of the Basin and Range province, block faulting and effusion of voluminous deposits of rhyolitic material occurred together, apparently both as response to crustal tension. Inasmuch as this volcanism was commonly initiated during late Oligocene time, and locally even earlier, the earliest block faulting may also have been Oligocene or earlier. Evidence of such early range-front faulting occurs in the Santa Rita Mountains (Drewes, 1972b, p. 7 and pl. 1), where late Oligocene rhyolite was emplaced concurrently with the movement on the San Cayetano fault, along the eastern foot of the San Cayetano Mountains. More typically, the volcanic rocks cover the oldest range-bounding faults, only to be offset themselves by numerous Miocene faults (20 m.y. and younger) that are readily detected and dated.

As young as this basin-and-range event of tectonism, volcanism, and sedimentation is, relative to the Cordilleran orogeny, its main effects were to disrupt and conceal the older structures. Such effects are incidental, if also aggravating, to the focus of this synthesis. Current studies in the Basin and Range province, such as those of Reynolds (1980, 1982, 1985), however, suggest further complications of extensional movements that do not as readily lend themselves to separation from those of the Cordilleran orogeny.

In much of the western part of the study region, Tertiary rocks characteristic of the Basin and Range province and older rocks are tilted upon crystalline basement rocks along low-angle faults. The Tertiary rocks mostly are weakly indurated clastic rocks and weakly to strongly indurated pyroclastic rocks, together rarely more than a kilometer thick. In contrast, the underlying crystalline rocks are foliated, lineated, and cataclastically deformed. A system of low-angle faults separates these strongly contrasting terranes, and local evidence typically shows that movement on these faults was in the Miocene or later. The amount of this movement is not conclusive; the strongest evidence is for movement in the order of a few kilometers, enough to accommodate the rotation of overlying rocks. However, evidence of an older generation of faulting is present where some pre-Oligocene rocks lie directly upon the basement. Commonly such evidence is only found through extensive systematic geologic mapping. Where such mapping was not undertaken, or where it accompanied by only a local study, the common resulting conclusion is of a single stage of gravity-generated detachment faults. In many other studies (Reynolds, 1980, 1982; Bykerk-Kauffman and Janeke, 1987), however, the multistaged development is recognized, with the Miocene detachment event being the last one. Even where no Paleozoic or Mesozoic cover rocks intervene between the brittlely deformed Tertiary deposits and ductily deformed basement rocks, the juxtaposition of these diverse styles of deformation must be viewed as enigmatic.

At the heart of this geologic enigma is a swarm of large gneiss-cored domes, also known as metamorphic core complexes. These domes occur singly or in small clusters in the western half of this southeastern extension of the Basin and Range province (fig. 3). Perhaps because of this apparent association, the domes were viewed by some workers (Davis, 1975) as being cogenetic with the mid- and late-Tertiary extensional structures. However, such domes occur far beyond the confines of the Basin and Range province, in the northwestern states and even in Canada; yet they do not occur outside of the Cordilleran orogenic belt. From this line of reasoning alone, at least the major components of the gneiss-cored domes must have been generated by early Tertiary time and were then modified or masked by later tectonic events. Part of the development of these domes, then, is germane to this synthesis but the later development, intertwined with extensional detachment faulting and with geothermal complications that are part of both these developments are largely avoided herein because it is too young.

About 300 km south of Bisbee and El Paso is the projected western end of the Appalachian-Ouachitan orogenic belt, of Permian age (fig. 3). This belt of folded and thrust-faulted rocks is well known in the northeastern part of the Big Bend area, occurs in a few windows in central Chihuahua (Bridges, 1962; Dyer, 1986), and has recently been identified in central Sonora (Poole and Madrid, 1986). At Big Bend, Tex., the foreland zone of the Cordilleran orogenic belt appears to underlie the older belt, although probably this is the result of Tertiary faulting. In general, the post-Ouachitan-deformation rocks form the pre-Cordilleran rocks, and the effects of the Cordilleran orogeny were superposed on those of the earlier Appalachian-Ouachitan orogeny. While this earlier orogenic event affected terranes south of the region under study, it doubtless left its mark on the sedimentary record of terranes to the north, although the remnants of mid- and upper Permian sedimentary rocks are sparse everywhere. Thus, the source of the sandstone in the Lower Permian (Leonardian?) Rainvalley Formation, which thickens southeastward to the sandstones of the Santa Rita Mountains[1] of Muela (1985) and Brown (1985), may reflect this early tectonic event.

A last consideration is the location of the study region between the large igneous provinces of the Sierra Madre in Mexico and the Datil volcanic field along the Arizona-New Mexico border. The Sierra Madre volcanics are believed to cap a major middle Tertiary batholith, and possibly the Datil field caps a smaller one. A subsurface tie between these igneous provinces is possible, but no investigations have yet been made that verify the subsurface tie. In our efforts to understand the tectonic changes across the southern states, we tend to forget that some variations may show up along a longitudinal section.

This brief review of various tectonic elements that demonstrably or speculatively occur near the El Paso-Las Vegas segment of the Cordilleran orogenic belt serves primarily to illustrate that structural complexities are to be expected in this, and perhaps in any, regional orogenic synthesis. Some of these complexities are likely to be reflected in multiple deformational events, or tectonic overprinting; others may reflect the diversity of tectonic events in neighboring regions whose interpretations are subject to question. The ensuing synthesis is thus a working hypothesis of the evolution of a segment of a major orogenic belt influenced by many kinds of tectonic features and processes, and that in turn has influenced the development of others. Despite the apparent structural complexity, a surprising degree of uniformity and a systematic change in structure, which is similar to the rest of the orogen, is identified.

CORDILLERAN OROGENIC BELT

Orogenic belts that formed during a single cycle of compressive deformation have in common many features of geometry and of timing, and the El Paso–Las Vegas segment of the Cordilleran belt is no exception. Some comments on the characteristic features of such an orogen will facilitate the description of the features in the study region.

Conventionally, in speaking of parts of an orogenic belt, the terms "autochthon" and "allochthon" are used to refer, respectively, to the passive and the active terranes. A proper application of these terms requires a knowledge of global plate movements, which is not usually available from the deformed rocks themselves. Those rocks only record the kind of stress, orientation of stress, and perhaps the duration of stress to which they responded. In the case of the Cordilleran orogenic belt, these terms were applied prematurely and erroneously, without a full appreciation of global tectonics. As now understood, the North American plate and the several northeast Pacific plates, chiefly the Cocos plate, were both mobile and were converging between Jurassic and Paleocene times and perhaps longer. Their collision along the western side of the North American craton resulted in the development of the Cordilleran orogen. Consequently, both the more deformed and generally overlying terrane, and the less deformed and commonly underlying terrane were active, or allochthonous. For the present purpose it is expedient to retain the entrenched improper application of the terms which equate "allochthon" with the more deformed and "autochthon" with the less deformed rocks. Furthermore, any disputes on whether the deformed belt to the southwest was pushed over the craton to the northeast or whether the craton was pushed under the deformed rocks is moot; both masses were actively converging, sometimes nearly head on, and at other times obliquely.

In plan view, the North American Cordilleran orogenic belt has some sharply delineated features along its eastern to northeastern or frontal margin, but generally gradational or poorly delineated features along its western to southwestern or core region. The orogenic belt is made up of several structural zones whose characteristics are more uniform along strike than they are between adjacent zones, across strike. Distinctions between a ductily deformed hinterland zone nearer the core of the belt and a more brittly deformed fold-and-thrust zone along the distal margin of the belt are commonly noted. In places, it is also practical to separate an intermediate zone, or several intermediate zones, from the others. These zones display physical and genetic continuity from one to the next. While the frontal line of the orogen is commonly linear (fig. 2), some segments of the line are lobate, probably reflecting such situations as a more rapid propagation of a central part of the stress field where rock strengths or overburden confinement changed more markedly than of the lateral parts. Offsets in the regularity of these tectonic zones (not to be confused with regions I–IV of fig. 2) may occur, along major strike-slip faults where transverse stresses were built up, such as may result from the presence of structural barriers or from diverse rates or duration of deformation between adjacent segments of the belt.

[1]These Santa Rita Mountains of the Sierra Alta chain of mountains of Chihuahua should not be confused with the Santa Rita Mountains southeast of Tucson.

The mildly deformed terrane of the foreland zone is in front of the orogenic belt, to the east or northeast and farther on the craton than the orogenic belt; in some studies, but not in this one, the zone is extended to include the fold-and-thrust belt (or zone). The foreland zone, as herein used, varies in width; it includes the Colorado Plateau and central Rocky Mountains (area II of fig. 2) but is much narrower north and south of this region (not shown in fig. 2). Its western or southwestern margin is sharply delineated by the frontal line; whereas, its eastern or northeastern margin is diffuse. Structural features of the foreland zone typically are scattered and less intensely developed than those in the orogenic belt itself. Genetically, these features are related to those of the belt but their geometric or physical ties to the structures of the orogenic belt are uncertain or unlikely.

Similar to the foreland zone is a terrane that lies beneath the eastern or northeastern part of the orogenic belt that I refer to as the subfold-and-thrust area. This area once lay beneath the fold-and-thrust zone but has been exhumed through deep erosion. In a sense, of course, it is continuous with the foreland zone (and hence, I view it as an "area" rather than a "zone"). It shares with the foreland zone the characteristics of having a mild deformational imprint, and it may share with that zone the uncertainty that specific structures were physically continuous with major structures of the orogenic belt.

While the distinction between the foreland zone and the subfold-and-thrust area seems arbitrary or unnecessary, if not impractical to demonstrate, the concept helps to explain some of the irregularities of the frontal line and of the particular distribution of various styles of deformation to be described in the following section.

Figure 4 shows a transverse view of a generic orogenic belt with many features that can be found across the El Paso–Las Vegas segment of the Cordilleran orogenic belt. The boundaries between the foreland zone, and its neighboring terranes are generally distinctive and sharp; that between the intermediate and hinterland zone is less sharp, except where the terranes are telescoped through faulting. The foreland zone and subfold-and-thrust boundary is gradational. The diagrammatic transverse view of figure 4, however, is not meant to imply that an orogenic belt ever had such an appearance at a given time. The effects of ongoing erosional processes and gradual transmission of compression across the belt were disregarded in the construction of figure 4 in order to illustrate the tectonic terminology that will be used in this report. Specific terranes exemplified in later sections of the report are indicated by the lettered inserts. Note the approximate position of the ductile to brittle zone isopleth, shown to rise toward the core area with its inferred high heat flow.

In such a composite generalized transverse view, the orogenic belt typically forms a blunt-tipped tectonic wedge, thinning east or northeast toward the craton. In the Canadian and Montana segments of the belt, structures of this frontal configuration are often referred to as sled-runner faults (fig. 4, A). Deeper into the orogenic belt, shingled thrust faults and snake-head structures (large drag folds or upper halves of faulted folds) are common (fig. 4, B), and in places form hydrocarbon traps. Across the intermediate zones of the belt, the interaction with late orogenic magmatic deposits increases, in places accompanied by mineral deposits. Folds and shingled thrust faults have thickened the allochthon. The youngest of the deformed rocks here are typically older than those near the frontal line, and they may have been derived from sources farther toward the hinterland. In the innermost or hinterland tectonic zone[2] of the Cordilleran belt (fig. 4, E and F) ductile deformation is the rule and metamorphic terranes are nearly pervasive. Such terranes are only exposed because of the large amount of late orogenic and post-orogenic uplift and ensuing erosion, implied by the gradual shift from A to F (fig. 4) from the surface to a 8–10 km depth. The sites A–F are analogs to locations to be reviewed across Chihuahua, New Mexico, and Arizona, and an even greater depth is recorded at sites farther to the southwest of the rocks located at site F.

In transverse profile, the boundaries between the tectonic zones are probably inclined, although in local studies this inclination may not be apparent. The directions of inclination probably converge, in part because of the uplift in the core of the ductile terrane that defines the hinterland zone. The net effect is to imply that a substantial factor in the northeastward development of these tectonic zones and their many separate nappes and plates has been gravitational. Perhaps that genetic basis for orogenic belts as a whole remains tenable only upon considering that the ultimate source of the regional force was derived through the rising of lighter weight magmatic material than its host rocks near the orogenic core. If this be the case, then the geologic controversy of a gravity versus push origin of orogenic belts may be relegated to a semantic arena.

This overview of some general features, both in plan and profile views, of the better known segments of the Cordilleran orogenic belt, composited from such classical sources as Price and Montjoy (1970), Rubey and Hubbert

[2]A zone of steeply inclined ductile-fabric features, common to many orogenic belts, is not here recognized; thus, hinterland zone, as used herein, is not the same as orogenic core.

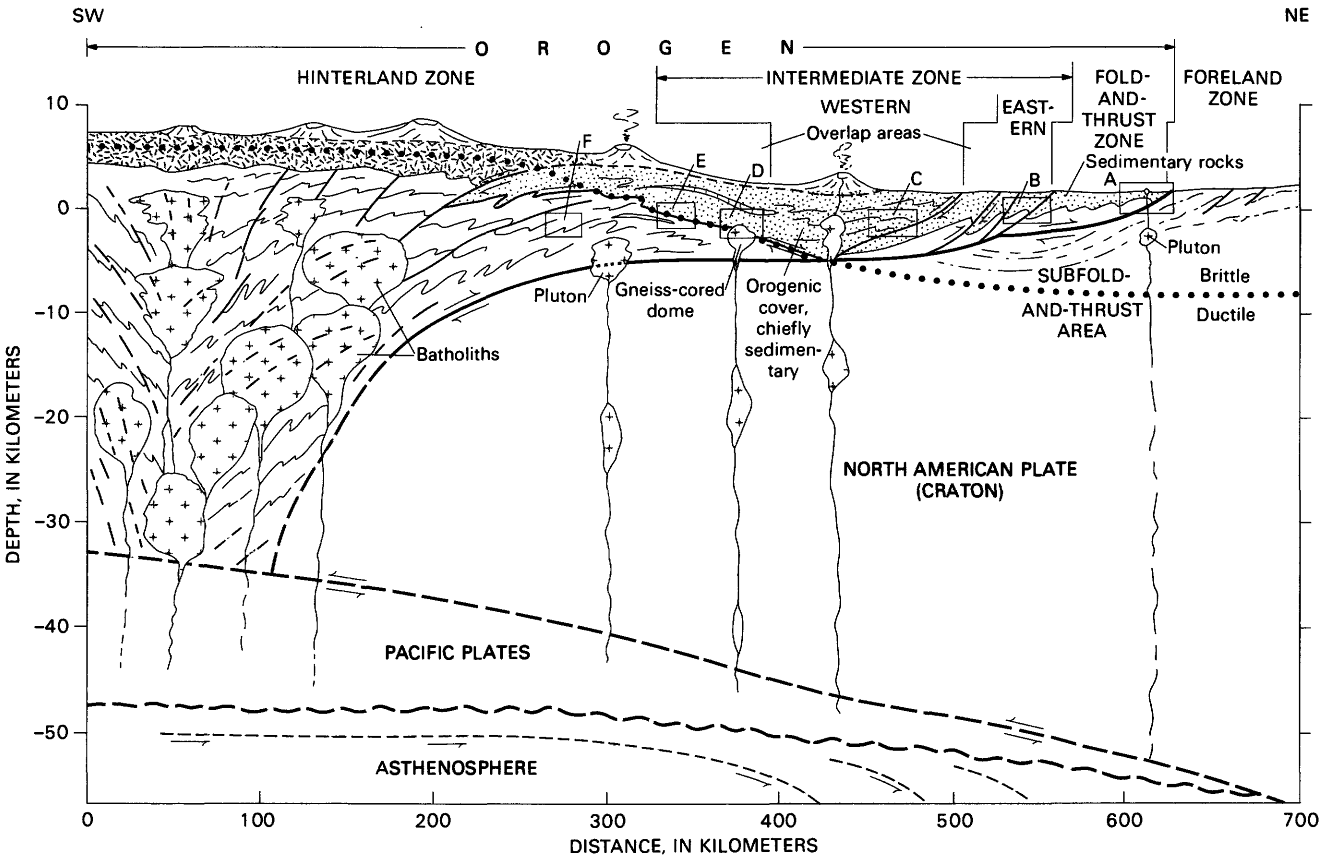


FIGURE 4.— Diagrammatic section across an orogenic belt, showing general geometric relationships between tectonic zones, compositing successive peak orogenic times in each zone, and omitting subsequent erosion. Sites A–F, referred to in text, simulate conditions found exposed at present along the southern transect of the Cordilleran orogenic belt.

(1959), and Mudge (1972), provides a starting point from which I will examine the geology of the El Paso–Las Vegas segment of the Cordilleran orogenic belt.

TEXAS-CALIFORNIA TRANSECT

STYLE OF DEFORMATION

The deformational elements common to each of the structural zones of other segments of the Cordilleran orogenic belt also are found in the El Paso–Las Vegas segment, but the distribution of these elements has been modified through a shear stress system along the southwestern side of the Colorado Plateau. Instead of the overriding allochthon colliding head-on with the overridden autochthon, the collision was oblique. As a consequence, a gap exists in the continuity of the fold-and-thrust zone (but not of the belt as a whole), and the other zones take on an extra intense deformation along the northern part of the allochthon.

In the following description on the styles of deformation, the zones will be reviewed in sequence from the fold-and-thrust zone to the hinterland. The foreland zone, which lies outside of the main part of the belt, is reviewed after the fold-and-thrust zone has been covered, in order to emphasize adequately the contrast between these zones. In general, the structural styles are gradational between zones; however, locally, the change in styles seems to be fairly abrupt. Where fairly abrupt change occurs there are usually indications that major basement faults or a change in mechanical properties of the deformed rocks have had an influence.

While systematic description often invites tedium, such effort may lead to exciting new insights on the nature of the subject, which has clearly been the case along the El Paso–Las Vegas segment of the orogenic belt. The concept of the propagation of a stress field through rocks and the resultant spread of an orogenic belt is well established. It follows, then, that each structural zone in an orogenic belt, except the leading one, is a composite of an earlier more frontal style of deformation and one or more styles of deformation developed under more hinterland-like conditions. For example, one might expect some remnants of a foreland style to occur in the fold-and-thrust zone, or some fold-and-thrust zone structures may be preserved among structures of the intermediate zones. This composite aspect of most zones in an orogenic belt may help to explain the local occurrence of structural anomalies. Examples of such remnants are believed to occur across this segment of the belt.

A point of clarification is needed on my use of the terms "segment" of the orogenic belt and "transect" in the following description. In reviewing some general features of the Cordilleran belt the label "El Paso–Las Vegas segment" was used as a lateral subdivision of the whole belt, just as one might refer to the Wyoming or Alberta segments. A "transect," crosses the structural grain of some terrane; it is a cross section or profile in an older parlance, perhaps without the connotation of the vertical dimension. A transect (line) cutting across the segment at right angles to the dominant structural grain would trend west-southwest, rather than west-northwest. However, more geologic information about Mesozoic to early Tertiary features is available along a line trending west-northwest, obliquely across the segment of the belt. Note that the orogenic belt, as a whole, trends north, not northwest, and so a transect that trends west-northwest is reasonably well oriented to show transverse changes. Therefore, I will follow such a transect and refer to it as the Texas-California transect, a trend that cuts through northern Chihuahua and southern New Mexico and Arizona and into eastern California.

FOLD-AND-THRUST ZONE

The fold-and-thrust zone is well developed almost everywhere along the northeastern side of the Cordilleran orogenic belt. It forms a broad terrane in Mexico, extending north to the El Paso area, and makes a narrower terrane across Utah, northeast of Las Vegas. However, between El Paso and Las Vegas it is absent or changes character. Typical fold-and-thrust zone structural features of the El Paso area (fig. 5) will be described in this section, and an explanation for the changes northwest of El Paso will be offered in a later section.

Structural features of the Sierra Juarez, just across the Rio Grande (Rio Bravo del Norte) from El Paso are typical of Chihuahuan fold-and-thrust terrane (pl. 1). Rocks of the sierra and its outlying hills are warped into a series of folds, many of which are closely associated with thrust faults whose continuity beneath the sierra is substantiated by structural repetition found in the Pemex Pozo Juarez No. 1 drill hole.

The Sierra Juarez are underlain mainly by Lower Cretaceous (Aptian and Albian) rocks but include one unit of Late Cretaceous age and also include some Eocene dikes (fig. 6). The Cretaceous sequence comprises alternating thick-bedded limestone and thin-bedded clastic units and marlstone. In the next range south of the Sierra Juarez, the Sierra de Salamayuca, these rocks are underlain by Neocomian clastic rocks and gypsum, possible Jurassic phyllite (Berg, 1969), and by Paleozoic rocks. Farther to the southeast, near the Big Bend area (fig. 3), the Cretaceous sequence includes the Upper Cretaceous Boquillas Formation as well as overlying units of Cenomanian to Maastrichtian age (Ojinaga to El Picacho Formations, respectively (fig. 6), of Kettenbrink (1984, p. 33). West of the Big Bend area this sequence is as much as 1.5 km thick, and it is folded. The Sierra Juarez area probably was covered also by these Cretaceous rocks, although they may have been a facies richer in clastic rocks than that to the southeast. Subsequent to early Tertiary uplift they were eroded. The Eocene dikes are andesitic to rhyolitic, and they cut folds and faults.

Large folds dominate the structural style of the Sierra Juarez. They trend N. 40–45° W., are some tens of kilometers long and about a kilometer in amplitude, and plunge gently (Wacker, 1972; Campuzano, 1973; Nodeland, 1977; work by Drewes and Dyer, in press; and this study, pl. 1). Smaller folds are common among the large ones; some merge with larger structures or merge with other small folds, to end in a homocline. Folds in the northeastern part of the area are generally tighter than those in the southwestern part. Fold axes plunge 5–20°, usually southeast, but some folds are doubly plunging. Most folds are inclined to the northeast and, inasmuch as they are genetically related to nearby southwest-dipping thrust faults, they are also vergent to the northeast. A fold with an anomalous southwest vergence is described below in connection with the Las Viboras Arroyo reverse fault or backthrust.

Folds of the Sierra Juarez are genetically related to thrust faults and tear faults, as illustrated by many disharmonic folds in thrust plates, or against (cut by) thrust faults (figs. 8, 9). Typically, clusters of folds end downward against a thrust fault following bedding planes in a shaley unit (Drewes and Dyer, in press). One fold ends against a strike-slip fault, probably as a disharmonic structure, and the strike-slip fault, in turn, terminates downward against a thrust fault (fig. 8). This group of structures is particularly instructive because the presence of a bedding-plane thrust fault might easily go unnoticed without close attention to local changes in attitude of beds. This thrust fault is also important because it dips toward the Pemex Pozo Juarez No. 1 well, in which unexpected stratigraphic repetitions were

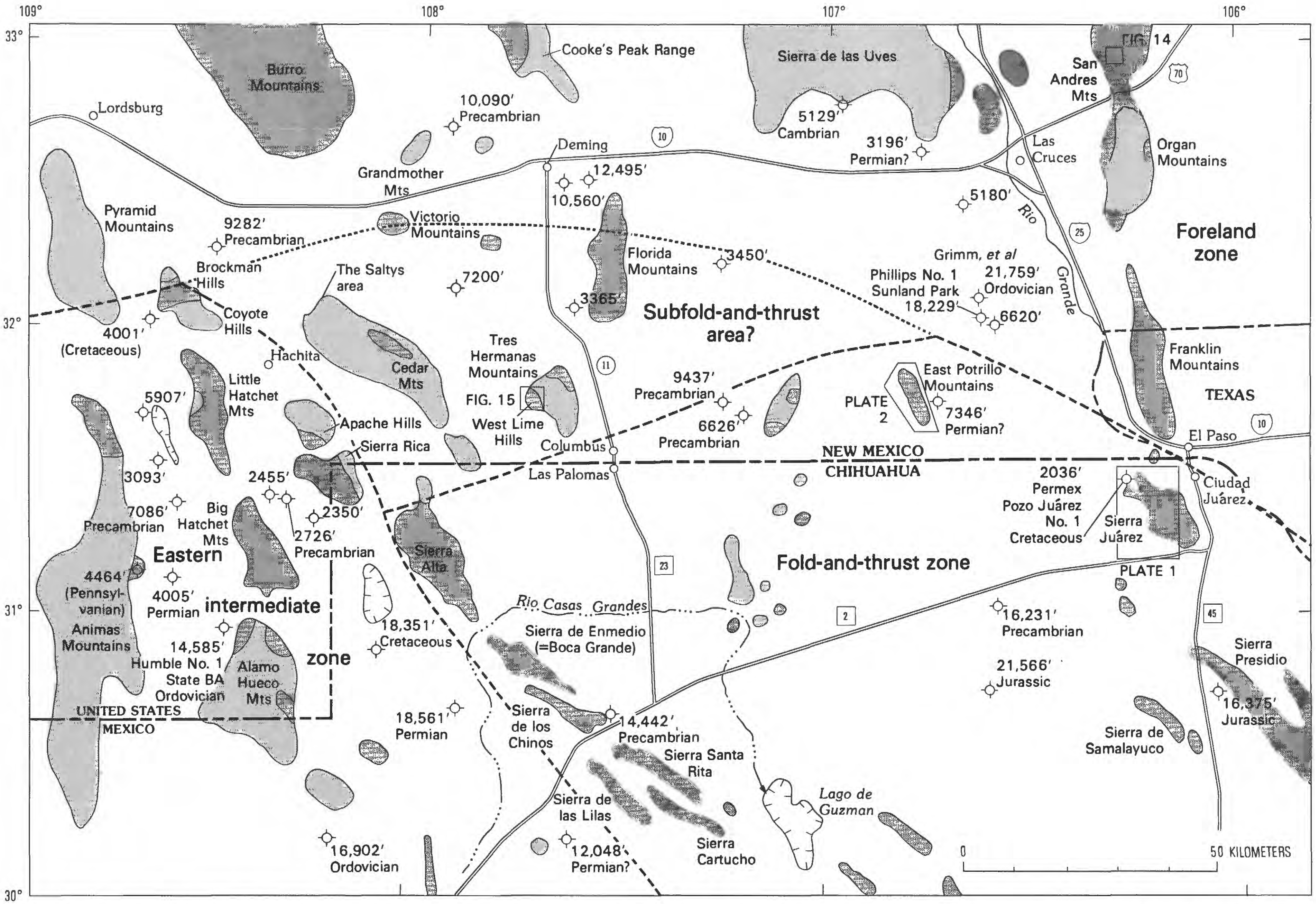


FIGURE 5.—Index map showing major outcrops of pre-Tertiary rocks exposed in the Cordilleran fold-and-thrust zone and some adjacent terranes (heavy shading) and Tertiary covering rocks (light shading). Key oil wells located, giving total depth, in feet, and age of rocks at bottom of hole where known (excluding a bottom in Tertiary igneous rocks, indicated in parentheses). (Some ranges in Mexico are known by various names on different maps.)

AGE		STAGE	FORMATION	DESCRIPTION	THICKNESS (m)
QUATERNARY	Holocene			Gravel and sand	Thin
	Pleistocene		Camp Rice Fm	Gravel and sand, high terrace	Thin
TERTIARY	Pliocene		Fort Hancock Fm	Gravel, sand, clay; bolson fill	Thick
	Miocene and Oligocene		Local names*	Mainly rhyolitic rocks; some andesitic lava	Thick
	Eocene		Campus Andesite	Hornblende andesite plugs, dikes	—
			Love Ranch Fm, upper part*	Conglomerate sandstone (to north)	300+
	Eocene and Paleocene(?)		Love Ranch Fm, lower part*	Sandstone, shale, conglomerate (to north)	300+
CRETACEOUS	Late Cretaceous	Maastrichtian	El Picacho Fm*	Shale and limestone (to southeast)	200–300?
		Campanian to Coniacian	San Carlos Fm*	Shale and limestone (to southeast)	500–1000?
		Turonian and Cenomanian	Ojinaga Fm*,	Shale and limestone (to southeast)	200–300?
			Boquillas Fm	Shale, thin-bedded limestone	10–265
	Early Cretaceous	Albian	Buda Limestone[1]	Thick-bedded, nodular	10–12
			Del Rio Fm[1]	Shale, thin nodular limestone	24–30
			Anapra Fm[1]	Sandstone and siltstone	67
			Mesilla Valley Fm	Shale, siltstone, thin limestone	42–55
			Muleros Fm	Nodular limestone, shale	21–38
			Smeltertown Fm	Shale, siltstone, thin limestone	44–140
			Del Norte Fm	Shale and thin limestone	15–27
			Finlay Limestone	Massive reef limestone	130–186
			Lagrima Fm	Shale, marlstone, reef limestone	208–239
			Benigno Fm	Massive reef limestone	206
			Cuchillo Fm	Limestone, sandstone, shale	278
		Aptian	Las Vigas Fm*	Limestone and siltstone	?
		Neocomian	Unnamed	Siltstone, sandstone, gypsum(s)	Thick
JURASSIC(?)			Unnamed	Sandstone, conglomerate, phyllite	Thick
PERMIAN	Early Permian		Hueco Limestone*	Light-gray, cherty (to east)	300–1000
			Epitaph Dolomite	Dark-gray (to west)	300–500
			Colina Limestone	Dark- and light-gray (to west)	300–500
PRE-PERMIAN			Older rocks*	Limestone, dolomite, quartzite, shale, and granite	—

[1] These formations are placed in the Cenomanian by Lovejoy (1980).

FIGURE 6.—Generalized stratigraphic column for the Sierra Juárez, Chihuahua, Mexico, and units present in nearby areas which may have been present, at or near the sierra, marked *. Unconformities shown by vertical ruled intervals. Since this report was prepared, a series of nannofossil-bearing marine rocks 3,000 feet thick, of Upper Cretaceous, Paleocene, and Eocene ages, were reported from a drill hole 70 km to the northeast (Sam Thompson, personal commun., 1988).

encountered. In figure 9 (see location on plate 1), the block northwest of the strike-slip fault (intruded by a dike) abef moved northeast relative to both the southeast block and to the lower thrust plate. The syncline does not cross either the thrust fault or tear fault within the area of outcrop. Likely, both the fold and the tear fault are disharmonic structures formed concurrently with the thrust fault; the alternate interpretation that three separate tectonic events acted successively unduly complicates the interpretation of geologic history.

Thrust faults in the Sierra Juárez mainly strike northwest and follow bedding planes (pl. 1; fig. 8; and Drewes and Dyer, in press). The faults are more abundant along the northeast flank of the range than along the crest or southwest flank (pl. 1). Along most of the thrust faults, older rocks are emplaced upon younger ones, but in some instances younger rocks are faulted upon older ones. In the few places where thrust faults are exposed within or adjacent to limestones, they are marked by thin sheets of sheared or brecciated rock that usually are recrystallized to secondary calcite strongly resembling the parent rock. Truncated beds and interrupted or repeated rock sequences commonly mark the concealed traces of thrust faults. The Pemex Pozo Juárez No. 1 well, at the northwest end of the range, penetrated a 3-fold repetition of some of the Cretaceous formations, apparently reflecting thrust faults like those shown in figure 8.

The relative direction of tectonic transport of upper plates was N. 40–55° E., much as has been described by Cordoba (1968). This direction is indicated by the vergence of most folds (pl. 1 and fig. 8) and by the

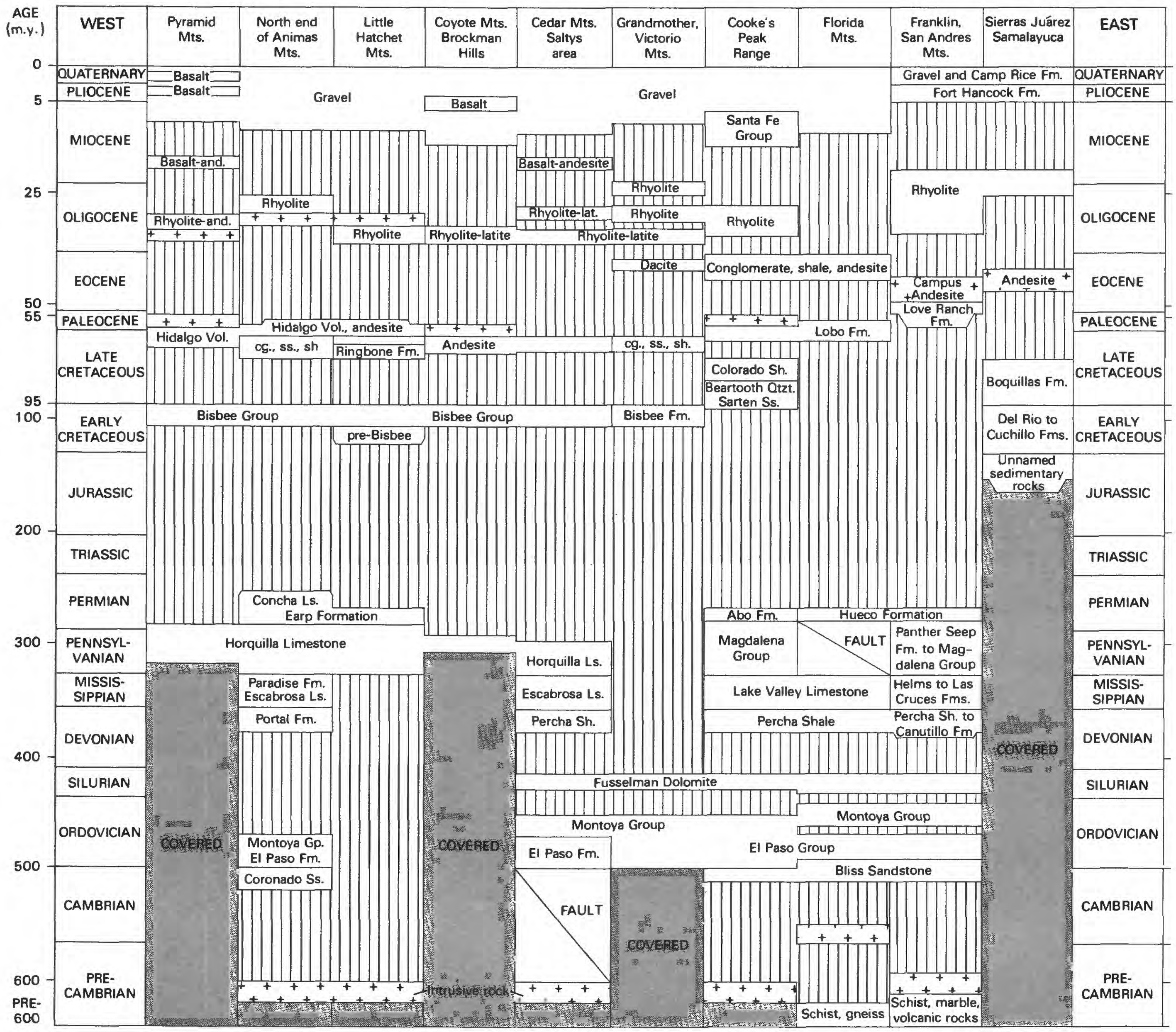


FIGURE 7.— Correlation diagram of rocks between El Paso, Texas, and southwestern New Mexico.

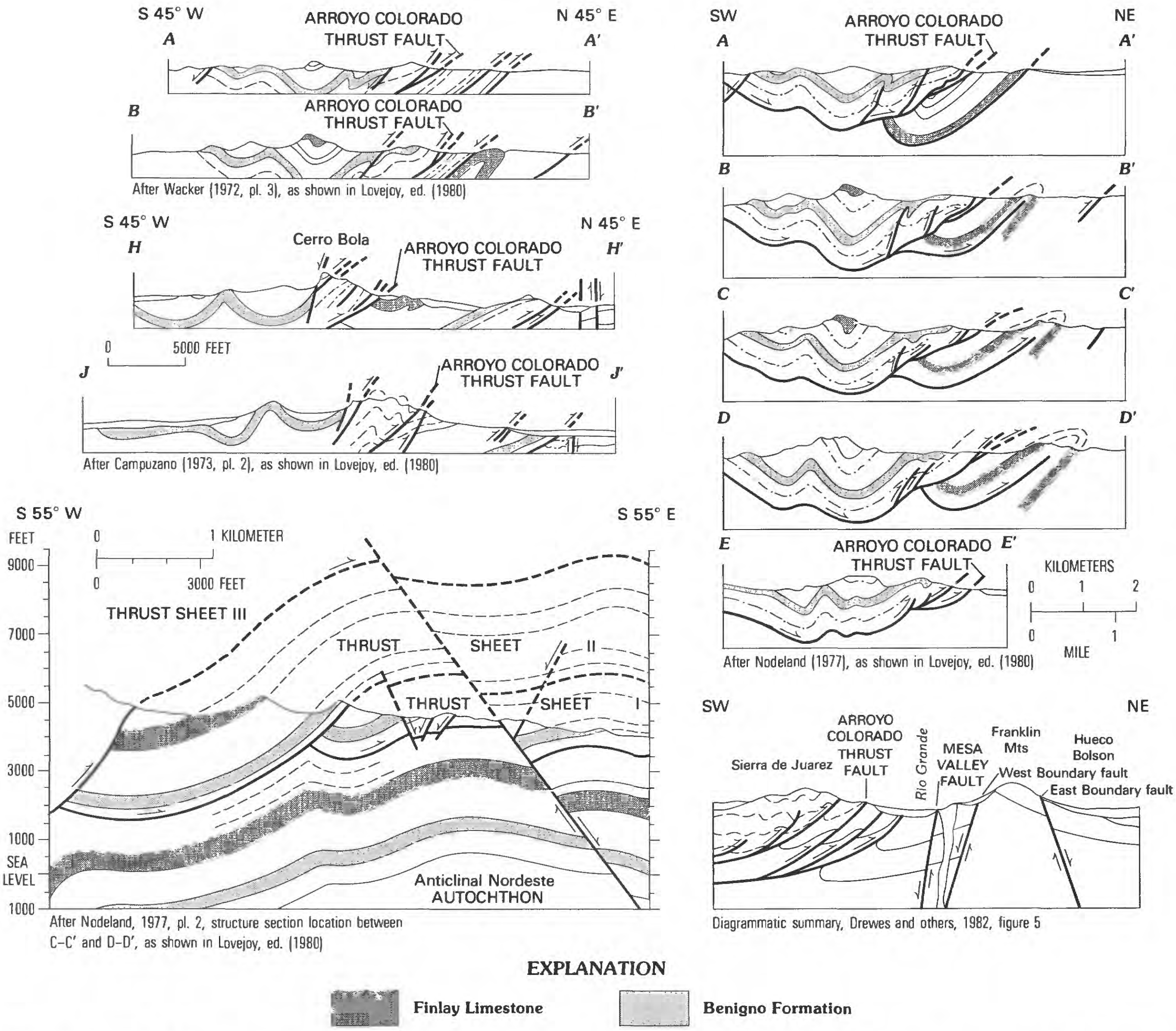


FIGURE 8.—Style of deformation in the Sierra Juárez, Mexico, as reported by many students, in Lovejoy (1980) and compiled by Drewes and others (1982, fig. 5).

orientation of strike-slip faults between thrust plates (fig. 8).

Some structures found in the fold-and-thrust zone can be misleading if considered out of context. The Las Viboras Arroyo fault is one such structure. Rather than being transported to the northeast, that fault was a site of southwest transport, as indicated by an adjacent fold overturned to the southwest (fig. 10). Structures, such as the Las Viboras Arroyo fault, are typical backward-directed features found in compressively deformed belts, and offer, at a small scale, an example of similar structures in the intermediate zones. The Las Viboras Arroyo back thrust probably terminates downward against the major underlying thrust fault that brings rocks low in the Cretaceous sequence (Cuchillo Formation) onto younger rocks (Anapra Formation, and, out of the plane of section of fig. 9, the Finlay Limestone, fig. 6). Although the critical juncture of faults is concealed, the disharmonic relations between structural units clearly show the presence of these faults and their relative importance. Furthermore, these relations are consistent with those of other folds and faults in the range. The Las Viboras Arroyo fold, if taken out of geologic context, could provide misleading evidence of regional transport direction. When viewed in context with the nearby structural features, it illustrates one method by which upper plate rocks are thickened and thereby gain the strength to further transmit a regional stress.

The age of deformation in the Sierra Juárez is Late Cretaceous to early Tertiary, typical of the Laramide phase of the Cordilleran orogeny. From local evidence, the deformation took place after deposition of the lowermost Upper Cretaceous Boquillas Formation (about 80 m.y. old) and before emplacement of a group of andesitic dikes, two of which have been dated by R.F. Marvin and others (written commun., 1985) as early Eocene (48.2±1.7 and 47.8±2.2 m.y.). This almost 32 m.y. age range can be substantially narrowed using the argument of the likely past presence of a substantial cover of Upper Cretaceous rocks, such as those west of the Big Bend area, that extend into the Maastrichtian Stage (as young as 63 m.y.). The Sierra Juárez folds could hardly have formed at the surface but required a cover of a few kilometers of rock. Relicts of potential cover rocks have been reported from an exploratory well recently completed northeast of the East Potrillo Mountains, in which 1,000 m of marine Maastrichtian, Paleocene, and Eocene beds were cut (Sam Thompson, personal commun., 1988). Likewise, across this segment of the orogenic belt the common association of andesitic rocks with a late stage of deformation favor the inference that the Sierra Juárez deformation occurred late during the available time range. Consequently, it is likely that most of the structures formed in the Sierra Juárez are Paleocene, or 55–63 m.y. old, and final response to stress may have lasted until about 48 m.y. ago.

The greater intensity of deformation affecting rocks on the northeast side of the Sierra Juárez relative to the southwest side is believed to reflect the proximity to the northeast side of a structural obstacle to the northeastward-moving thrust plates. Figure 11 illustrates the nature of this obstacle; the major northwest-trending fault (the Clint fault or Texas lineament) that separates the Sierra Juárez from the Franklin and Hueco Mountains to the northeast, had a vertical offset or component of movement, indicated by the exposure in these ranges of rocks as old as the Precambrian crystalline basement. A northeastward-moving plate

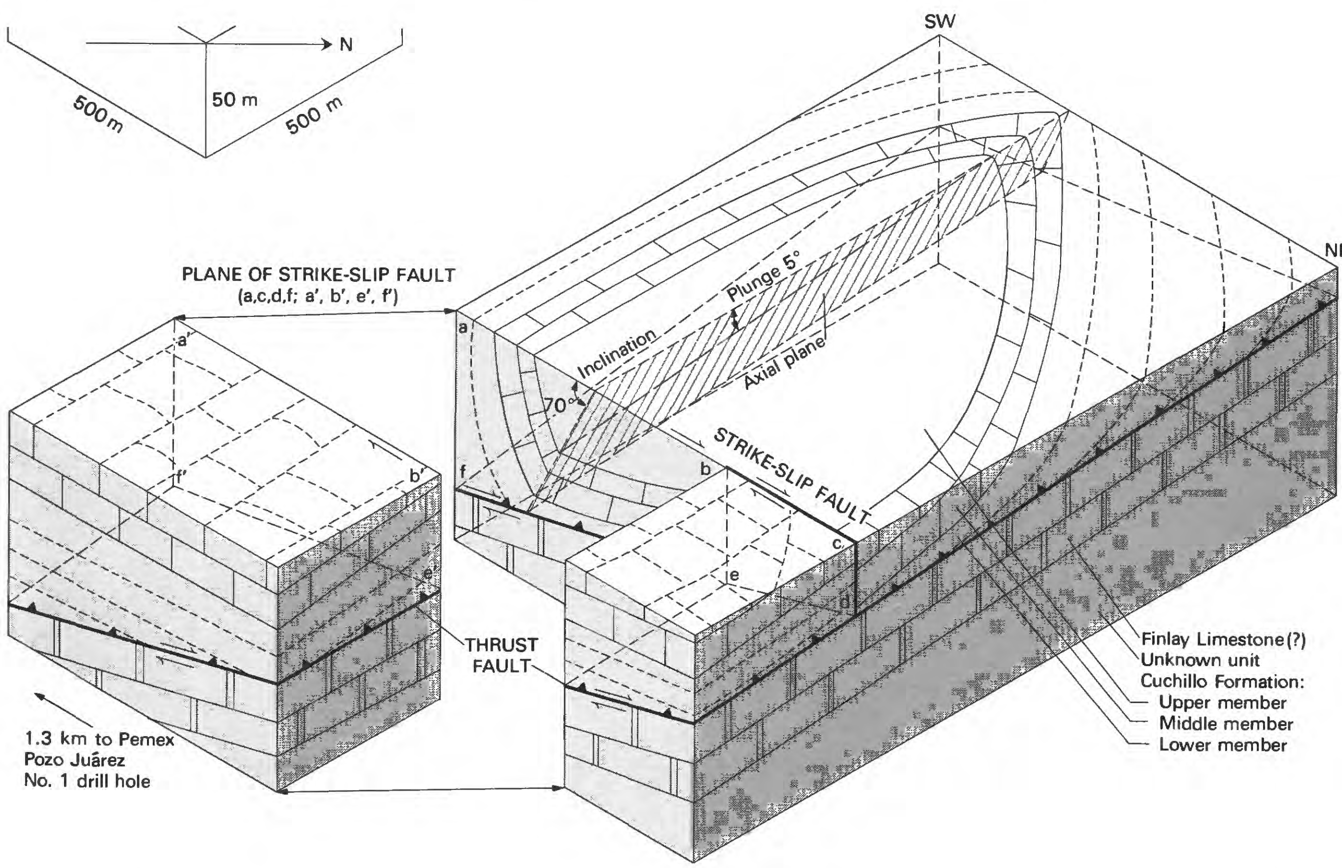


FIGURE 9.—Block diagram of disharmonic fold and strike-slip or tear fault, and their genetically related thrust fault, from an example in the fold-and-thrust zone, Sierra Juárez, Chiuhuahua, Mexico. Here the thrust faulting has brought the Cuchillo Formation (normally the third formation below the Finlay Limestone) above the Finlay. Effects of topography are eliminated and vertical exaggeration about 5×. Dike in plane of strike-slip fault is not shown.

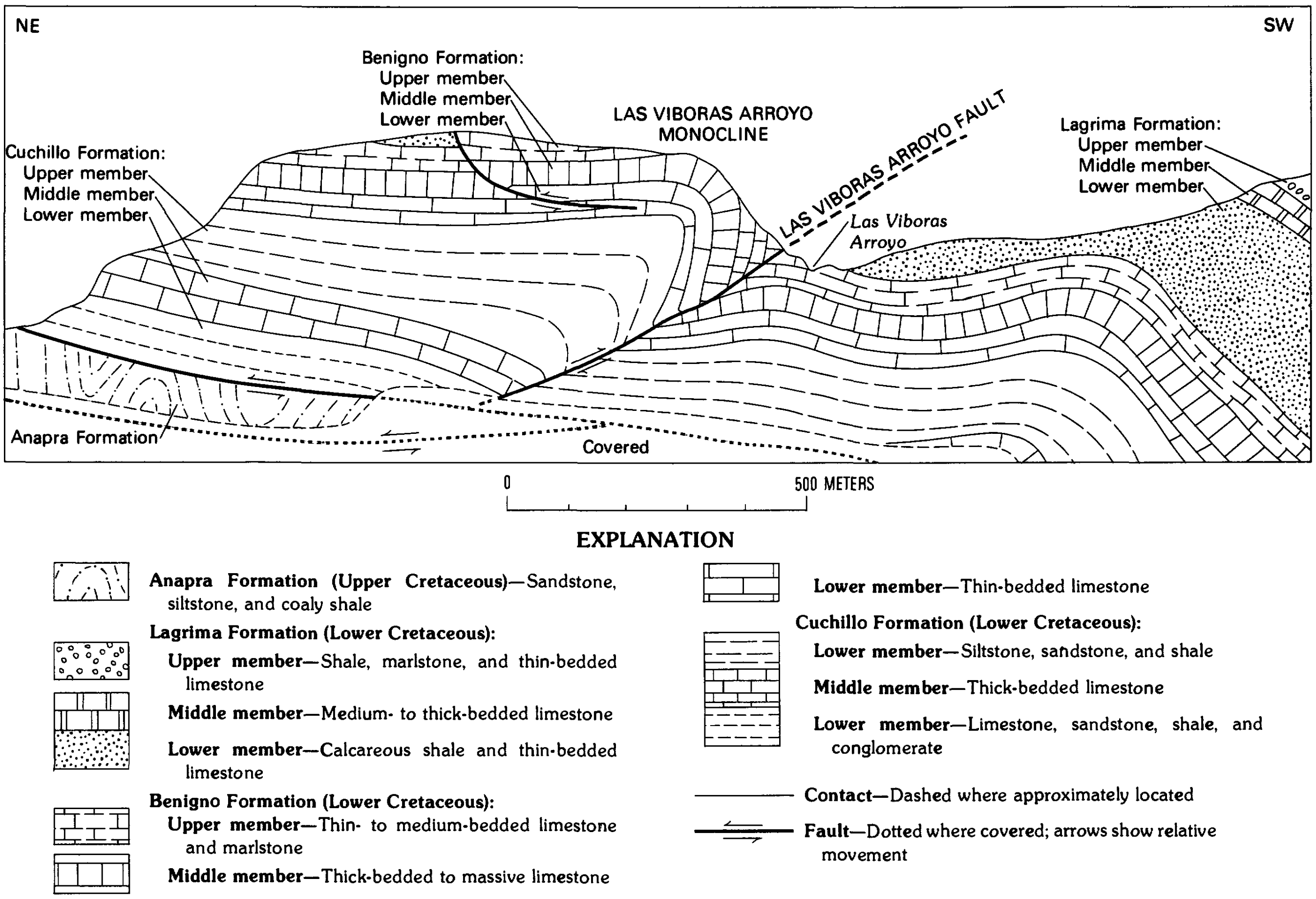


FIGURE 10.—Diagrammatic structure section of the Las Viboras Arroyo monocline and fault (a backthrust or reverse fault), Sierra Juárez, Chihuahua, Mexico. See geologic setting on plate 1. Large vertical exaggeration.

system, largely or wholly in Cretaceous rocks, encountering such basement rocks would likely be deflected upward, possibly developing a ramp structure, and probably leading to more intense deformation of the lower part of the plate system, as found in the Sierra Juárez. Deep erosion of the leading edge of this plate system, and basin-and-range block faulting and bolson sedimentation could subsequently generate the geologic relationship found today (fig. 11*C*).

Though much of northern Chihuahua is covered by young surficial deposits, structural features like those of the Sierra Juárez crop out in other ranges. Folds and thrust faults occur in Upper Cretaceous and possible Jurassic rocks of the Sierra de Samalyuca south of the Sierra Juárez (fig. 5) (Kettenbrink, 1984), some of which are mildly metamorphosed. West of the Sierra Juárez, across a broad bolson and extensive Miocene and Pliocene basalt field, folded and locally thrust-faulted Cretaceous and Permian rocks are reported in the Sierra Alta, also known as the Sierra de Palomas, (Phillips, 1986; Guthrie, 1987; Sivils, 1988) Sierras de Enmedio and de los Chinos (Brown, 1985) and Sierra Santa Rita (Campbell, 1984; Muela, 1985). These folds are upright or inclined to the southwest, and most are open structures like those on the southwest flank of the Sierra Juárez. With these folds are reverse faults or backthrusts having southwest-directed movement. The style of deformation in ranges west of the Sierra Alta changes markedly and is part of the eastern intermediate zone, described subsequently.

Rocks in the East Potrillo Mountains, New Mexico, (fig. 5) are deformed in a fold-and-thrust zone style comparable to that of the Sierra Juárez. The range, however, is so narrow that the criterion of a broad field of folds cannot be applied in a correlation of structural styles; only the characteristics of the structures themselves are used, and thus the correlation is provisional and awaits further study; perhaps reviews of seismic data from nearby areas will provide the confirming data. Low on the northeast flank of the range, the East Potrillo Mountains are underlain by Lower Permian limestone and dolomite and a disconformably overlying

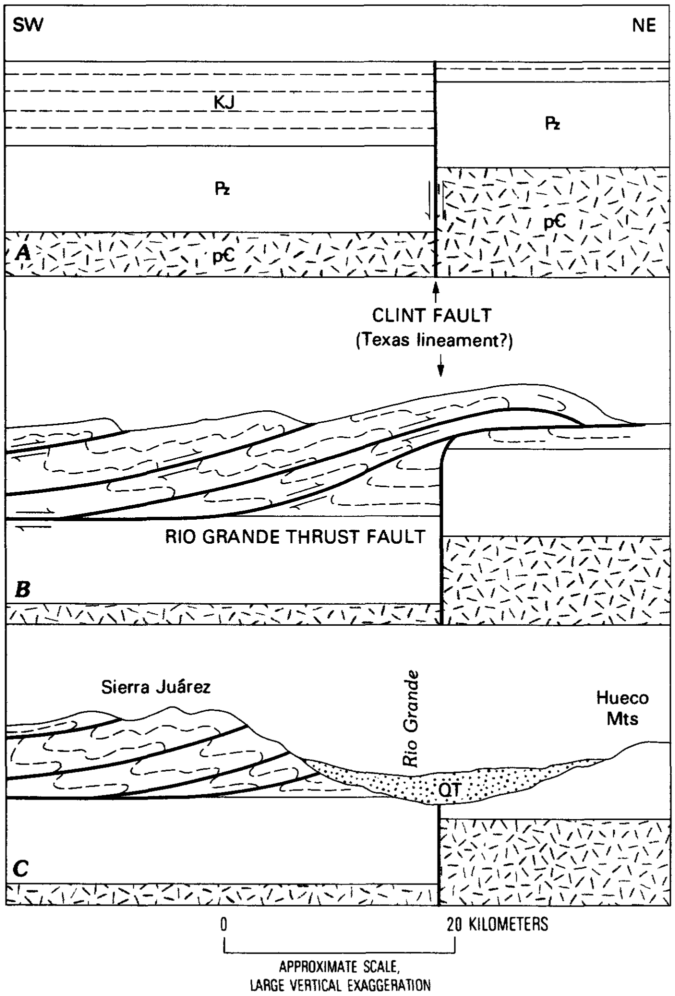


FIGURE 11.—Diagrammatic structure sections showing inferred development of the Rio Grande thrust fault near El Paso, Texas. A, Early Cretaceous; B, late Late Cretaceous; C, present. Arrows show relative movement of faults. QT, Quaternary and Tertiary rocks; KJ, Cretaceous and Jurassic rocks; Pz, Paleozoic rocks; Є, Cambrian rocks.

limestone and clastic rock sequence of Lower Cretaceous units that probably are correlative with the Cuchillo and Benigno Formations of the Sierra Juárez. The northern part of the southwest flank of the range is underlain by a structurally separate mass of the Lower Permian Epitaph Dolomite and Colina Limestone, formations resembling the Permian sequence to the west more than that to the east.

Geologic structures of the East Potrillo Mountains (pl. 2) increase in complexity to the northwest. Rocks of the southeastern third of the range dip homoclinally southwest (Powell, 1983). Those of the central part are slightly metamorphosed, and the Cretaceous rocks have a weakly developed cleavage and a northeast-trending lineation. Rocks of the northwestern third are unmetamorphosed, but part of the structurally separate mass, a thrust-fault plate or reverse-fault hanging-wall block, is silicified. The Cretaceous rocks onto which this mass was transported are folded and thrust faulted. Folds are of moderate or small size in both length and amplitude, are moderately tight asymmetric structures, in places discordant upon underlying thrust faults, and verge northeast. One of these thrust faults is itself folded. Crossbedding in sparse clastic beds in the overthrust mass of Permian rocks show that mass to be right side up, and therefore it cannot join the other body of Permian rock as a major overturned syncline.

The combined features of (1) rocks common to the deformed belt but not to the foreland zone, (2) scattered occurrence of cleavage implying local high-pressure environments, and (3) disharmonic folds and several thrust faults suggest that the East Potrillo Mountains have a stronger tectonic affinity to the fold-and-thrust zone of the orogenic belt than to the less deformed foreland zone. These rocks probably are part of a relatively thin plate of folded and thrust-faulted rocks separated by a décollement fault from much less deformed underlying rocks. They also are probably just southwest of a northwest-trending fault. Together, these structures could provide the tectonic setting for the development of cleavage and lineation and for the presence of the overthrust mass of Permian rocks through providing a stoss-side zone of extra high pressure. This condition will be reviewed in greater detail in the section of this report on the structural features of the northeast margin.

In northwest Chihuahua, a long chain of mountains, which I will refer to as the Sierra Alta chain, is broken into separately named short segments, stretching from the Sierra Alta in the northwest to the Sierra Cartucho in the southeast. This mountain chain seems to form the northwesternmost range of the fold-and-thrust zone (fig. 5). Most of the Cretaceous and older rocks between the Sierra Juárez and the Sierra Alta chain are covered; the few sites at which older rocks are exposed, northeast and southeast of Columbus, N. Mex., contain too few structures to define their tectonic style. The terrane generally north of Columbus has somewhat different structural features than those of the fold-and-thrust zone and will be described in the next section as part of a subfold-and-thrust area.

The Sierra Alta chain is underlain mainly by Pennsylvanian to Permian limestone, dolomite, and sandstone, by Lower Cretaceous rocks correlative with the Cuchillo Formation of the ranges to the east, and by the Bisbee Group of the ranges to the west. These rocks are folded in long northwest-trending open structures, mainly asymmetric to the southwest but locally inclined to the northeast. Some of the low-angle faults are northeast directed; these are probably the larger faults inasmuch as they emplace Paleozoic over Cretaceous rocks and have northeast-vergent folds. However, a few, apparently small, low-angle faults are found in most of the separate short ranges, and many of these faults record southwest-directed tectonic transport. Several of the individual short ranges underlain by diverse formations are separated from one another by faults (or valleys inferred to be fault controlled) trending subparallel to the axis of the Sierra Alta chain and branching around lozenge-shaped blocks, as if they were strike-slip structures. Some smaller hills along the southwest side of the chain are underlain by rhyolitic porphyry plugs that may be aligned with similar bodies in the Sierra Rica (fig. 5) on the international border 30 km northwest of the Sierra Alta chain. Further details of this region are available from a series of theses at the University of Texas at El Paso, that include, from southwest to northeast, work by Campbell (1984), Muela (1985), Brown (1985), Guthrie (1987), Phillips (1986), and Sivils (1988).

Structures of the Sierra Alta chain resemble those of the southwest side of the Sierra Juárez more than those in ranges to the west, with the exception that southwest-directed transport is more widespread (or extended farther along strike), and that the folds are generally more open than those of the Sierra Juárez (pl. 1). These changes may occur because, to the southwest, the level of the transmission of most of the stress lay at a lower stratigraphic level, near the base of a thick thrust sheet; whereas, farther east in the Sierra Juárez the stress was at a higher stratigraphic level (fig. 12). The southwest-directed structures are viewed as back thrusts, such as the small ones (pl. 1) found in the Sierra Juárez, and such as others to be described in the eastern intermediate zone later in this report. In both places, the back thrusts also are largest along the southwestern side of a local fault block, the northeast side of which contains mainly large thrust faults. For example, in the northeastern part of the Sierra Rica–Little Hatchet–Big Hatchet group of ranges are several large northeast-directed thrust faults (Zeller, 1960, 1970) and in the southwestern part of the ranges are mostly small southwest-directed back thrusts (Zeller, 1960; Drewes, 1988).

FORELAND ZONE AND SUBFOLD-AND-THRUST AREA

The tectonic transition terrane between the pervasively deformed fold-and-thrust zone and the undeformed autochthon is referred to as the foreland zone (fig. 4). Foreland zone structural features are isolated, commonly a single syncline and anticline fold-pair, either with or without a related fault, between which rocks were undeformed during the same tectonic event as when rocks were deformed in the main part of the orogenic belt. Major faults commonly are axial plane faults to anticlines, which extend down-dip probably to more ductile crustal level rather than flattening out to remain in the more brittle level. For this reason, these faults probably do not join the frontal fault or the décollement fault in front of, or beneath, the fold-and-thrust zone. With improved knowledge of deep

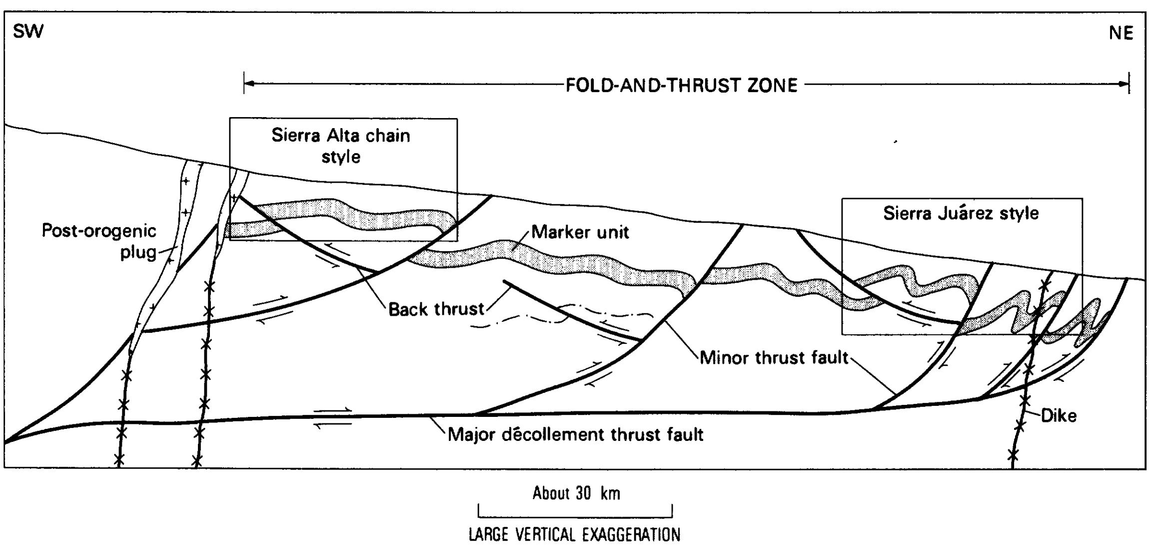


FIGURE 12.—Diagrammatic structure section across the fold-and-thrust zone of northeastern Chihuahua, Mexico, showing inferred changes in degree of folding between thinner distal (NE) and thicker proximal (SW) parts. Arrows show relative movement of faults.

structural configurations, this distinction may be refined or modified; however, the anticipated deep projection of faults is still an uncertain feature.

The term "subfold-and-thrust area" is used to refer to that terrane underlying the fold-and-thrust zone, or once having underlain it, but now exposed by deep erosion (fig. 13). The subfold-and-thrust area, like the foreland zone, has isolated structures that were formed in response to mild compressional stress relative to the more intense stress that produced the structural features of the fold-and-thrust zone and the more interior zones of the orogenic belt. Also like the isolated structural features of the foreland zone, there is no indication of physical continuity between the minor structures of the subfold-and-thrust area and the structures of a major overlying (or once overlying) plate of the fold-and-thrust zone. However, unlike the structures of the foreland zone, those of the subfold-and-thrust area were influenced by the northwest-trending strike-slip system of basement faults, so characteristic of the Texas-California transect of the orogenic belt. Folds and thrust faults near these northwest-trending basement faults are abundant, small in amplitude or apparent offset, complexly related to one another, and associated with signs of oblique-slip or strike-slip movement. Possibly, these locally complex structures lay beneath oblique-slip ramps in the once overlying thrust plate. In general, then, these scattered signs of complex structures localized along older (pre-orogenic) faults are believed to result from the transmission of a minor amount of regional compressional stress into rocks beneath a major thrust plate, as well as into the foreland zone, northeast of this thrust plate.

Some field examples of structures of the foreland zone and of the proposed subfold-and-thrust area are described here. In a later section of the report I review these structures in context with the variety of other features characteristic of the northeastern border of the deformed or mobile belt.

In the southern part of the San Andres Mountains (fig. 5), the Bear Peak fold and fault is a prime example of a foreland zone structure (Seager, 1981). Whereas rocks in the adjacent areas are flat-lying to homoclinally gently dipping, unfolded, and cut only by late Tertiary faults, those of the Bear Peak area are warped in an anticline-syncline pair about 10 km in length and 2 km in amplitude (fig. 14). The axial plane of the anticline dips moderately southwest and is followed by a thrust or reverse fault that brings Precambrian granite up against Paleozoic rocks. This fault does not flatten downward and, indeed, may actually steepen, so that it is unlikely to join the fold-and-thrust zone to the southwest. Orientation of these structures indicate that they were formed in response to northeast-southwest-oriented compressive stress.

Besides being an excellent example of a foreland zone structure, the Bear Peak fold-and-fault area offers a means of dating the time of deformation when the anticline and syncline were formed. In the axial area of the syncline can be found the Love Ranch Formation, which was deposited from late Paleocene to early Eocene. During the time of deposition of the Love Ranch Formation, deformation began, for within the Love Ranch Formation is an angular unconformity between the lower member and the upper member (Seager, 1981). The lower member of the formation is strongly folded in the core of the syncline, while the upper member, composed of coarse conglomerate derived from the adjacent anticline, is only weakly folded across the syncline core. This is demonstrated in figure 14, section *A–A′*; the upper member of the Love Ranch Formation is projected into the plane of the structure section from just north of the mapped area.

Such structures as the Bear Peak fold and fault are part of the foreland zone. As part of the foreland zone, they are physically separated from the structures of the other zones, but they still can be linked to the formation of the main part of the belt by the time of formation of the structures, during late Paleocene and early Eocene, and by the direction of stress that produced the structures, namely, northeast-directed stress.

The structural features of the Florida, Victorio, and Cedar Mountains, and possibly also the Tres Hermanas Mountains, 30–50 km southeast to southwest of Deming, New Mexico (fig. 5), have some characteristics of the Bear Peak fold-and-fault area of the foreland zone, but also the complications of polyphase deformation and involvement of the northwest-trending basement faults typical of this transect of the orogenic belt. Thrust faults are present in the ranges, but their significance is unclear. If not actually retaining remnants of a major thrust plate, these areas probably lay beneath such a plate, which subsequently was eroded away southward to about the northwest ends of the Sierra Alta and the East Potrillo Mountains (fig. 5). Other ranges, which lie north of Deming, have exposures of pre-orogenic rocks that were not much deformed in Late Cretaceous or Early Tertiary time; likely they are of the foreland zone. The rest of the ranges near Deming have only post-orogenic volcanic rocks, and thus help little. Most likely the main faults of the Florida and Victorio Mountains are low level parts of structures whose upper levels were reactivated as bounding structures of a plate system.

The Florida Mountains are composed of Cambrian granitic rock, a Paleozoic sequence, a Cretaceous and Tertiary orogenic conglomerate, and Eocene and younger volcanic and sedimentary deposits. The rock

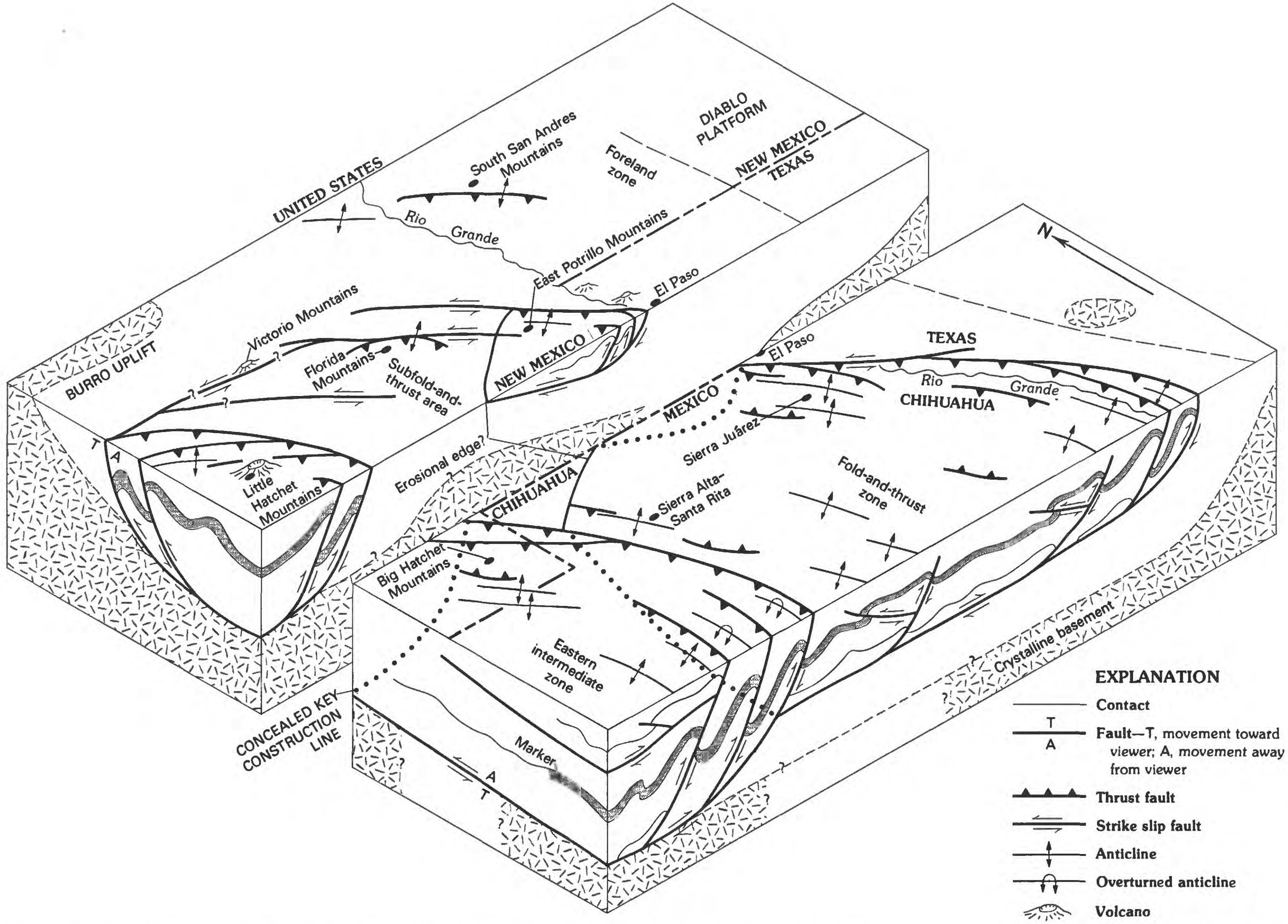


FIGURE 13.—Block diagram (pulled apart) showing variations of tectonic styles between the fold-and-thrust zone and adjacent terranes (not drawn to scale and with vertical exaggeration). Erosional edge subsequently concealed by Tertiary and Quaternary deposits. See figure 5 for map view and location of features.

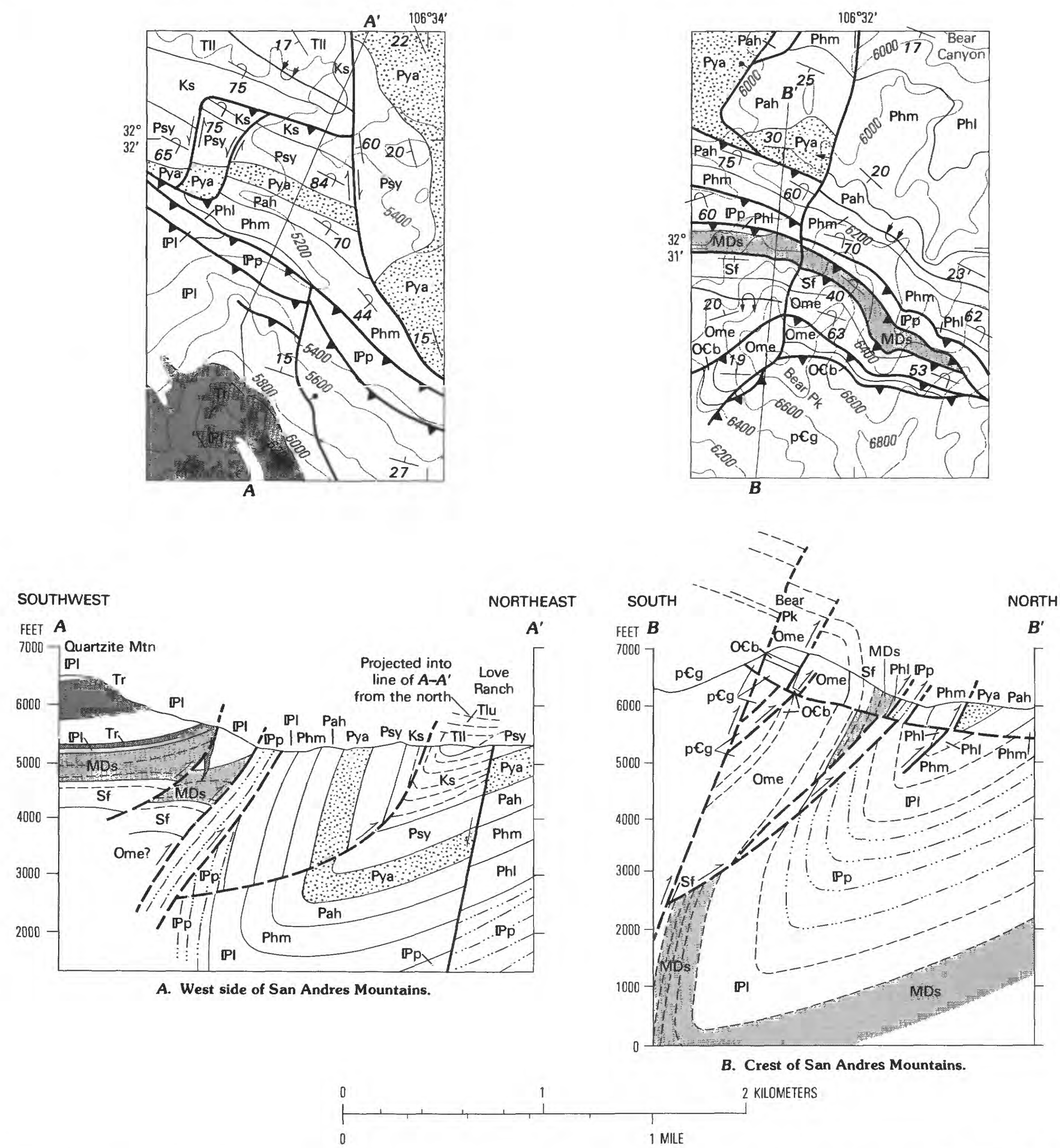
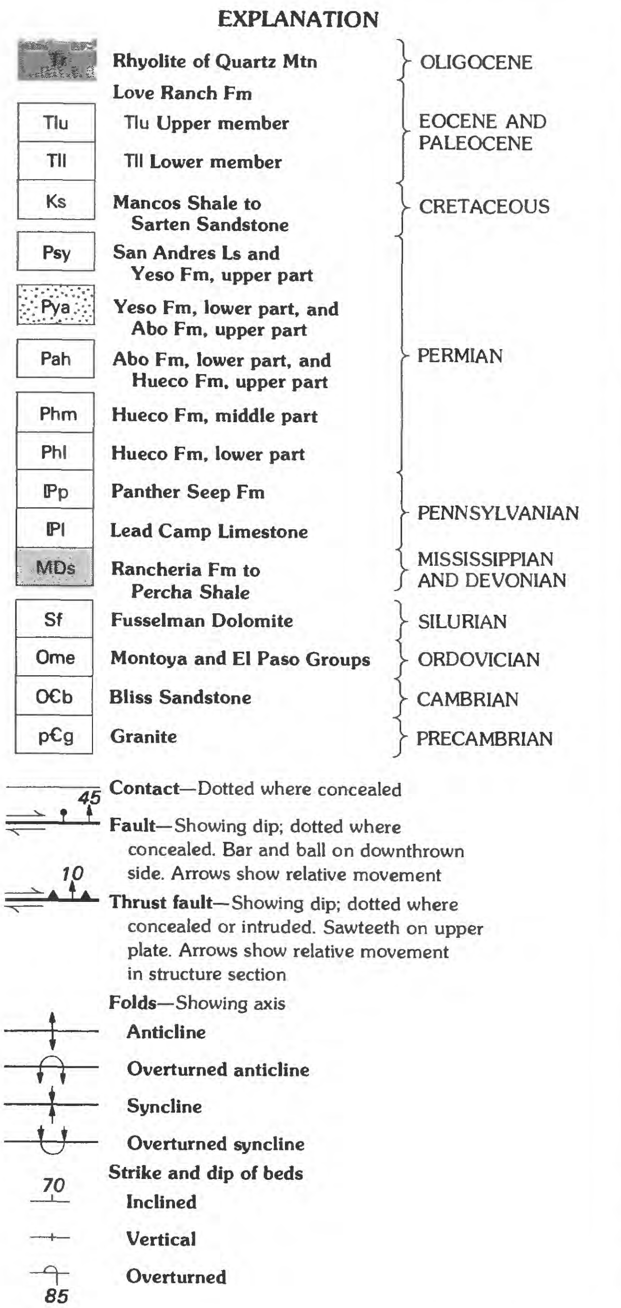


FIGURE 14 (Above and facing page).— Geologic maps and structure sections, showing a foreland zone structure, from parts of the Bear Peak fold and fault, southern San Andres Mountains, northeast of Las Cruces, New Mexico, after Seager (1981), compiled by Drewes and others (1982, fig. 11).

units and their structures have been variously interpreted during early reconnaissance and topical studies by Darton (1916), Corbitt and Woodward (1970), and Brown (1985) and are shown in greatest detail by Clemons and Brown (1983) and Clemons (1985). A major fault zone strikes northwest across the mountains, dips moderately southwest, and records recurrent thrust- and reverse-fault movement. Arguments may also be made from



detailed maps of these studies that strike-slip or normal fault movement may have occurred on this fault zone. In general, tectonic transport was northeasterly, although the detailed features on these maps and their structure sections show diverse (conflicting?) transport directions ranging from northwesterly to southeasterly. Likewise, the general impression from map relations and texts suggest the age of deformation to be Late Cretaceous or Paleocene, although a few detailed map features leave some uncertainty. For example, the critical orogenic conglomerate is shown to lie unconformably across a thrust fault at one site, but at another it wraps around a hill of Precambrian granite in such a way that a thrust-fault contact is far more probable than the depositional contact shown (Clemons and Brown, 1983). In structure sections the authors show an eastward overturned anticline-syncline pair that to the west is cut by a major thrust fault; yet the lines of intersection of crossing structure sections are markedly different from one another, so that details of interpretation from these studies deserve further scrutiny.

Deformation of the Florida Mountains is, in any case, complex, in the sense that several kinds of movement occurred at different times, that the direction and age of deformation was of the Cordilleran orogeny, and that northwest-trending basement faults influenced some of the thrust faulting. Of several options of general interpretation that are viable, I favor one of relatively modest amount of response to compressive stress of rocks along an old, reactivated basement fault, and possibly beneath a now-eroded fold-and-thrust zone décollement plate composed chiefly of Permian and Cretaceous rocks. The total impression is one of a minor response to northeast-directed regional stress which reactivated a basement flaw as an oblique-slip fault. Perhaps this fault was the base of a ramp structure at the foreland margin of the now-eroded fold-and-thrust zone. The scattered signs of compressive deformation to the northeast of the Florida Mountains is distinctly milder or less complex.

In the Victorio Mountains, 35 km west-southwest of Deming (fig. 5), Cretaceous and Paleozoic rocks are cut by an east-trending, steeply southward-dipping, major fault. Eocene volcanic rocks cap the mountains. An east-trending fault, inferred to be a reverse fault (Kottlowski, 1960, fig. 13) or a thrust fault (Corbitt and Woodward, 1970, fig. 6), has more recently been viewed as a strike-slip fault (Thorman and Drewes, 1980). Evidence of compressional deformation is found only very near this fault, probably reflecting a local stress field. Any thrust plate that reached this far north in central southern New Mexico would probably have overlain the area of the present day Victorios, but it has since been eroded. Complexly deformed rocks of this orogenic belt are not known north of the Victorio Mountains.

Rocks of the Tres Hermanas Mountains, 40 km south of Deming (fig. 5), show evidence of mild compressive deformation probably related to the Cordilleran orogeny. The mountains are underlain mainly by a quartz monzonite stock, and by volcanic rocks of middle Tertiary age (Balk, 1962). Paleozoic rocks, some strongly contact

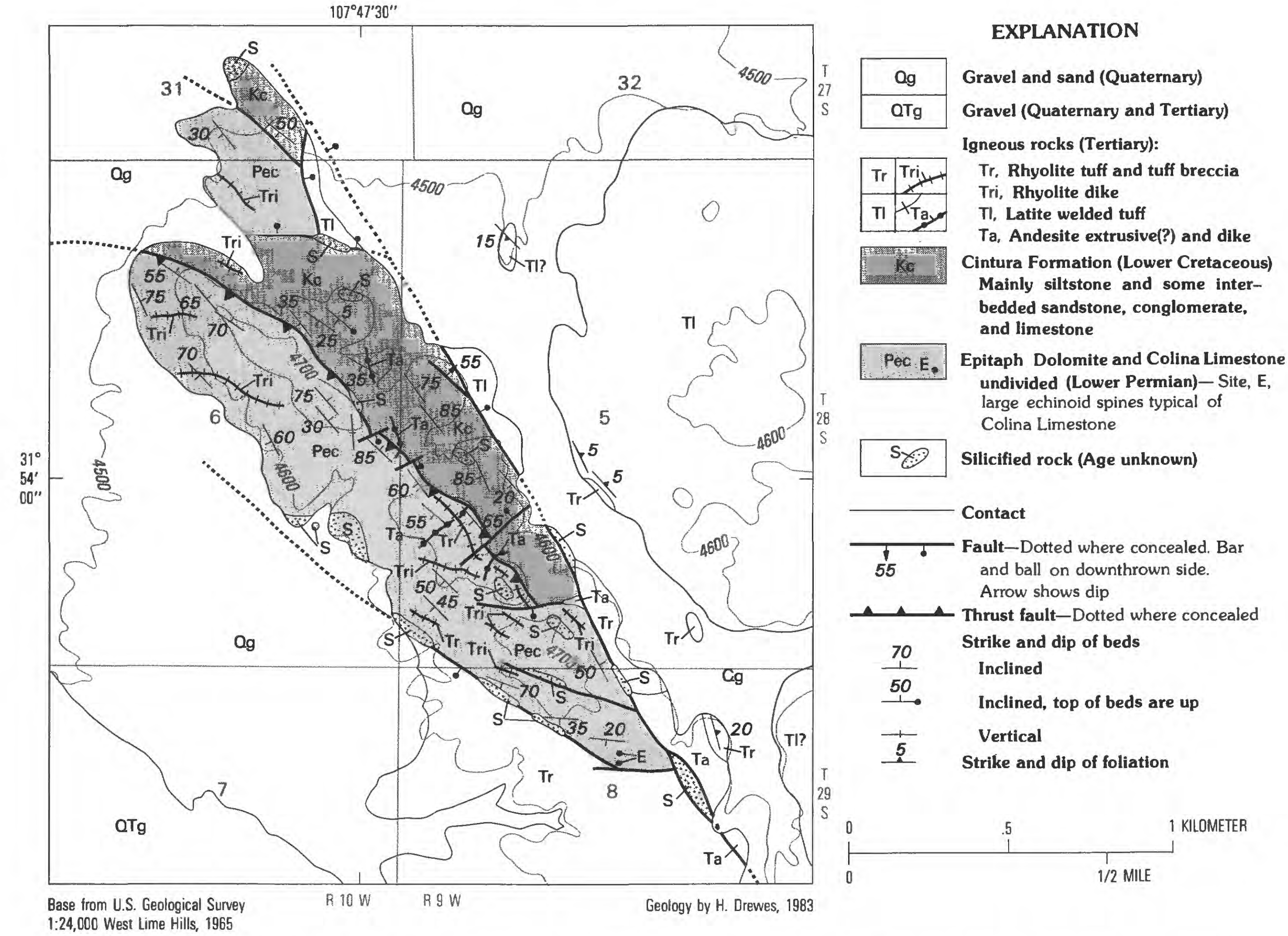


FIGURE 15.— Geologic map of the West Lime Hills area of the Tres Hermanas Mountains, New Mexico, showing Permian strata thrust faulted on Lower Cretaceous strata.

metamorphosed, occur along the north flank of the mountains, and Paleozoic and Mesozoic rocks lie in a northwest-trending fault block in a part of the western foothills known as the West Lime Hills (fig. 15).

Remapping of the West Lime Hills area of the Tres Hermanas Mountains by me (fig. 15) shows Lower Permian rocks thrust over Lower Cretaceous rocks. Rocks along the crest of the hills previously assigned to an unnamed Lower Cretaceous massive limestone (Balk, 1962) are herein assigned to the undifferentiated Lower Permian Epitaph Dolomite and Colina Limestone on the basis of lithologic similarity to these formations in nearby ranges and on the basis of euomphalid gastropods indicative of the Colina. Similar conclusions were independently reached by Sam Thompson on the basis of middle Leonardian conodonts collected by Drew Selby and identified by Tim Carr (Leonard, 1982, p. 19). The thrust fault juxtaposes older rock upon younger along a low-angle to moderately dipping surface partially concealed by colluvium. Furthermore, right-side-up sandstone beds in the Lower Cretaceous Cintura Formation of the Bisbee Group are truncated by the thrust plate. Two steep northwest-trending normal faults confine these older rocks and structures between blocks of Tertiary volcanic rocks. The trend of these normal faults resembles that of reactivated parts of larger strike-slip fault zones, such as those that occur in nearby ranges (Florida, Victorio, and Cedar Mountains). Thus, the thrust faulting may be a local feature with little tectonic transport. The Tres Hermanas situation illustrates the need for careful reviewing of local geologic maps used in regional compilations.

The close spatial and possible genetic association of major northwest-trending strike-slip faults and minor thrust faults is well illustrated in the Cedar Mountains, 45 km southwest of Deming. Most of this range is made up of middle Tertiary rhyolite, but low hills on their northeast side, known on some older maps as the Klondike Hills, are chiefly composed of Paleozoic rocks and some fault slices of Precambrian granite and Lower Cretaceous Glance Conglomerate (Thorman and Drewes, 1981). The major northwest-trending Cedar Mountain fault separates a northeastern terrane, mainly Mississippian Escabrosa Limestone, from a southwestern terrane, chiefly thrust-faulted Ordovician and Silurian formations. Strike-slip movement along the Cedar Mountain fault is indicated by slivers of rock unlike those of adjacent terranes along the fault and its branches, especially, the wedge of Precambrian granite faulted among Ordovician to Mississippian formations that is otherwise difficult to explain. The association of northwest-trending reactivated basement faults with signs of minor compressive deformation are the basis for placing the Cedar Mountains within the subfold-and-thrust area.

In summary, the structural features of the Florida, Victorio, Tres Hermanas, and Cedar Mountains, all of the theorized subfold-and-thrust area, show the influence of strike-slip faults, probably part of a basement fault system that generally trends northwest and has reactivated segments. Near these structures small folds and thrust faults show the same age of deformation and direction of tectonic transport found on structures in the ranges to the southeast, south, and southwest where pre-orogenic rocks are more extensively exposed. The structural features of these four mountain ranges thus resemble those of the foreland zone to the northeast and east, except for the involvement of the local folds and thrust faults with the northwest-trending basement fault system, a characteristic feature of the craton margin elsewhere in this orogenic segment and an integral part of the orogenic belt structures.

When viewed by themselves, the structural features of the Florida, Victorio, Tres Hermanas, and Cedar Mountains show little difference in style from those of the foreland zone. But that little difference in structural style takes on a greater significance when these structural features are viewed in a more regional context. For instance, the four mountain ranges lie just northwest of the enigmatic line north of the East Potrillo Mountains along which the fold-and-thrust belt appears to end. Also, the significance of the special interaction of local folds and thrust faults with the northwest-trending, reactivated, basement faults is apparent only after studying their distribution over a broad region from Texas to California. Likewise, the reason for finding that the shingled thrust faults characteristic of the next tectonic zone to the west extend north to a position in the Brockman Hills, north of the latitude of the Cedar Mountains, is appreciated only after recognition of the existence of district zones of various structural styles over a broad region. In this broader context, then, I propose that the rocks of the four mountain ranges were once overlain by a major thrust plate, probably consisting mainly of Cretaceous rocks, deformed in the style of the fold-and-thrust zone. This plate probably extended north about to the position of the north end of the Florida Mountains and the Victorio Mountains (fig. 5). Minor compressional responses occurred in the largely older rocks of the autochthon, particularly where the pre-orogenic basement structures produced local interferences to the minor stress transmitted into the upper part of the autochthon, and resulted in an assortment of minor strike-slip, oblique-slip, and reverse- to thrust-fault (up-dip) deformational responses.

It follows then that the fold-and-thrust zone may have extended northwest about to a line extended between Deming and the Brockman Hills (fig. 5). Along this line there probably was a strike-slip fault at the level of the allochthon, but the line may, in places, indicate segments of the basement fault system within the autochthon. Where the allochthon was fairly thin, as in the fold-and-thrust zone, erosion stripped back the plate and left the enigmatic subfold-and-thrust area, but to the west, where the allochthon became thicker in the intermediate zones (to be described in the following sections) that northern edge has some remnants. In a later section of this report the topic of this northern edge and the truncation of some tectonic zones will be reviewed systematically.

EASTERN INTERMEDIATE ZONE

The terrane west and southwest of a part of the subfold-and-thrust area and fold-and-thrust zone is deformed in a style sufficiently different from the eastern terranes to be given separate consideration as an intermediate terrane between the distal and proximal end members. Put more colloquially, it is intermediate between the "crumpled rug" style of the fold-and-thrust zone and the "putty" style of hinterland zone. Indeed, the intermediate terrane of this transect comprises two distinct parts that herein are referred to as the eastern intermediate and western intermediate zones. The eastern intermediate zone is made up of a single major plate without much (or any?) involvement of crystalline basement rocks; the western intermediate zone has multiple major plates and substantial basement involvement. In both of these intermediate zones, thrust faults are more abundant relative to folds than is the case in the fold-and-thrust zone, and also folds mostly are smaller. Tectonic associations with pre-orogenic to early orogenic foreland basin deposits and with late orogenic to early post-orogenic igneous rocks are more conspicuous and widespread, and the deformation is slightly older and more complex than it is in the fold-and-thrust zone.

The eastern intermediate zone extends essentially across the breadth of Hidalgo County, New Mexico, the county occupying the southwestern corner, or "bootheel" of the State. Mountains are more extensive and closer together there than to the east but, except for the northern 30–40 km of the deformed terrane, Tertiary volcanics cover much of the folded and thrust-faulted older rocks. Of particular interest are the Sierra Rica, Little Hatchet, Big Hatchet, and northern part of the Animas Mountains, as well as the Brockman Hills (fig. 16).

The boundary between the fold-and-thrust and eastern intermediate zones is covered mainly by the surficial deposits of the broad basins (bolsons) of the Hachita area and Playa de los Moscos, where the boundary probably is a major southwest-dipping thrust (or reverse?) fault. Thrust faults in the Boca Grande area southeast of Sierra Alta (fig. 5) mapped by Guthrie (1987) may splay from this major structure. The presence of a plate of Paleozoic rocks upon Cretaceous rocks and of aligned rhyolite porphyry plugs, sills, and dikes of Cretaceous or Tertiary age may mark the proximity of such a major northwest-trending fault in the Apache Hills and Sierra Rica (fig. 16) and Sierra de las Lilas (fig. 5); the next ranges northeast and southwest of this concealed fault have no such intrusive rocks, although the Little Hatchet Mountains do have assorted late orogenic granitic rocks.

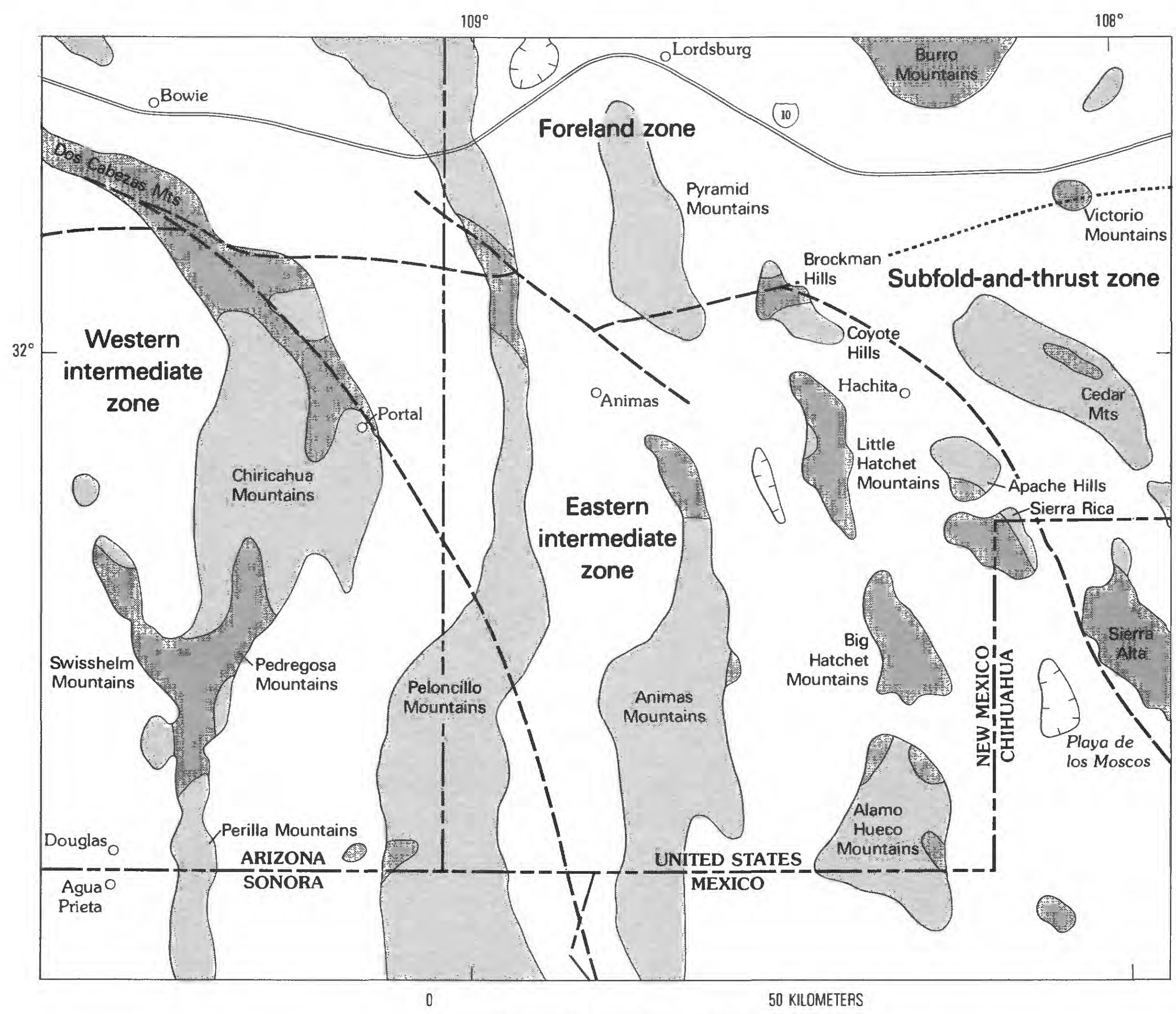


FIGURE 16.—Index map of the eastern intermediate zone, showing major outcrop areas of pre-Tertiary rocks (heavy shading) and Tertiary covering rocks (light shading), borders between tectonic terranes (heavy dashed lines), and border between foreland and subfold-and-thrust area.

The Sierra Rica, straddling the United States–Mexico border, may represent the eastern intermediate terrane nearest the northeastern edge of the zone. They are underlain chiefly by southwest-dipping Lower Cretaceous Bisbee Group, intruded along the eastern margin by the Cretaceous or Tertiary rhyolite porphyry and overridden, along their west flank, by a thrust plate mainly of Paleozoic rocks (Zeller, 1960, and by Van der Spuy, 1970). Zeller showed the Bisbee to comprise a series of imbricate thrust platelets, a type of structure he has mapped in both the Little Hatchet Mountains (Zeller, 1970) and on the southwest side of the Big Hatchet Mountains (Zeller, 1960). Van der Spuy viewed the Bisbee of the Sierra Rica as being essentially an unfaulted very thick Bisbee sequence. The belt of Bisbee Group at issue has thick shaley units in which thrust faults are readily concealed. Provisionally, I accept Zeller's interpretation, for he had the greater experience with Lower Cretaceous rocks and with field mapping in structurally complex areas; also my own mapping in the area closely resembles Zeller's work.

Along the west side of the Sierra Rica and at Doyle Peak, rocks of Precambrian to Permian age are thrust faulted northeast over the upper part of the Bisbee Group (locally called the Mojado Formation) (fig. 17). The Precambrian is represented by a small fault block of strongly shattered and altered granite resembling granite of this age in nearby ranges, but here not dated. This block of granite is separated from Paleozoic rocks by vertical faults. All these rocks and vertical faults are underlain by a thrust fault and its splay faults, which dip 10°–30° southwest. Because of the limited extent of this granite block and because of its vertical contacts, it is believed to have been faulted into the Paleozoic rocks before, rather than during thrust faulting. Therefore, this granite block is not viewed as evidence of basement rock involvement in the Cordilleran deformation, as that concept is normally applied. Folds in and under the thrust plate are of moderate size (less than 1 km long) and are asymmetric to overturned northeast. Inasmuch as the folds are closely associated with the thrust fault and its splays, they have a northeast vergence. Therefore, the Paleozoic extends beneath the basin gravels to the southwest. Farther to the southwest and northwest, along the northeast flanks of the Big and Little Hatchet Mountains are other thrust plates, the lower of which may be a continuation of the Paleozoic plate of the Sierra Rica and Doyle Peak.

The Big Hatchet and Little Hatchet Mountains make up the central part of a mountain chain extending from the Brockman Hills in the north to the Alamo Hueco Mountains in the south (fig. 16). The Hatchet Mountains are underlain mainly by Paleozoic and Mesozoic rocks; Precambrian granite is exposed in the gap between them. Late Cretaceous or Paleocene stocks intrude the Little Hatchet rocks (Lasky, 1947; Zeller, 1960, 1965, and 1970; Thompson and Jacka, 1981; Drewes, 1988, 1989, 1990a). Rocks of both mountain ranges have been deformed by northeast-southwest-oriented compressional deformation, of about Paleocene age, with some noteworthy changes in deformational style from north to south, illustrated in figure 18.

Cretaceous rocks of the northern half of the Little Hatchet Mountains are capped by the andesitic Hidalgo Volcanics that are latest Cretaceous (69.6 m.y.) in age (Marvin and others, 1978, p. 244, sample A). These occur in northwest-trending shingled fault plates aligned with the shingled thrust plates of the Sierra Rica. Stocks of quartz monzonite to diorite composition are probably cogenetic with the andesitic volcanic rocks, and are associated with typical late Cordilleran (Laramide) base-metal-vein mineralization.

Rocks of the southern half of the Little Hatchet Mountains, the northern part of the Big Hatchet Mountains, and the gap between them are cut by a series of northwest-trending strike-slip faults, possibly having left slip. Such movement is indicated on the northeastern faults of the northwest-trending set by left-lateral offset of the Bliss Sandstone. This formation lies on the east flank of the mountains 2 mi south of Hatchet Gap, and it must lie concealed beneath alluvium west of the hills in the gap itself. However, the same argument may be offered for an opposite movement sense on the southwestern fault of this set. Zeller (1965) offered other evidence for right slip, based on offset in limestone reef facies and on changes in attitude of some beds near the fault. However, since this fault dies out a few miles into the range, it may not be a strike-slip fault at all but a scissors fault (pivotal fault of Zeller), an option he also considered, and one which I favor in this instance.

These faults are at least partly older than some of the thrust faults of the northern Big Hatchet Mountains; in addition, some thrust faults are cut by other strike-slip faults. An Eocene granite (45 m.y.) cuts the northern of these later strike-slip faults. Apparently then, the Cordilleran tectonic events include an earlier thrust-faulting phase and a later strike-slip faulting phase, both of about Paleocene age. Furthermore, the shingled thrust plates appear to project downdip beneath older rocks of the southern Little Hatchets and likely also beneath the Big Hatchets.

Rocks of much of the Big Hatchet Mountains are gently folded limestone of Pennsylvanian and Permian ages (Zeller, 1965; 1975). Along their southwestern flank, however, the rocks are cut by northeast-dipping thrust faults separating platelets of tightly folded Permian rocks. These folds are of small or moderate (0.5–2.0 km) length and of small (0.1–0.5 km) amplitude and are

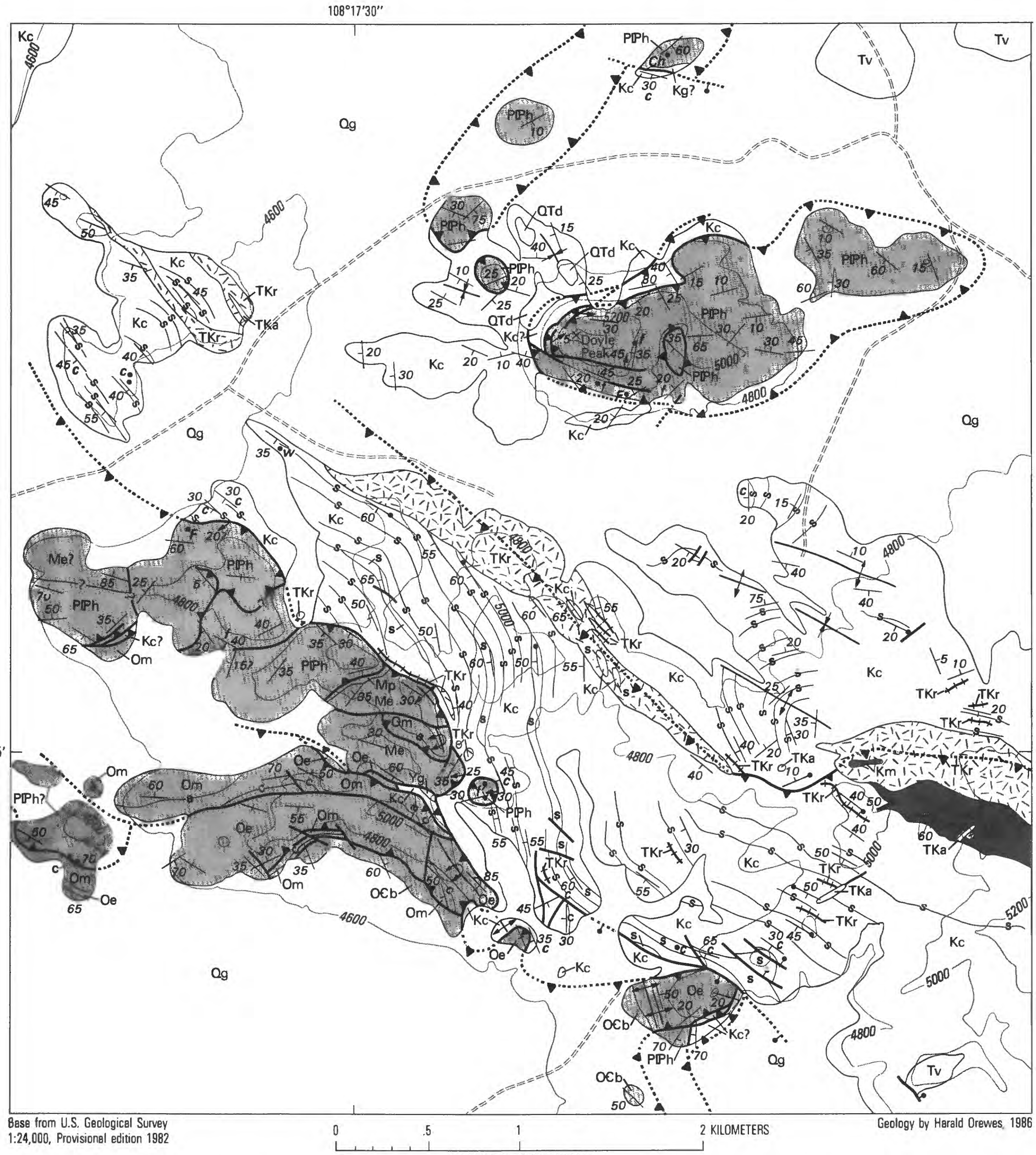


FIGURE 17 (Above and facing page).— Geologic map of a part of the Sierra Rica, New Mexico, showing Paleozoic rocks thrust over Cretaceous rocks.

EXPLANATION

Qg **Gravel and sand (Quaternary)**—Alluvium on pediments and on stream terraces

QTd **Block detritus (Quaternary or Pliocene)**—Landslide deposit

Tv **Volcanic rock (Tertiary)**—Rhyolite welded tuff to northeast; andesite to southeast

TKa **Andesite (Tertiary or Cretaceous)**—Aphanitic or rarely porphyritic

TKr **Rhyolite or latite (Tertiary or Cretaceous)**—Porphyritic (feldspar, quartz and biotite phenocrysts) or aphanitic

Bisbee Group (Lower Cretaceous)

Kc **Cintura Formation (equivalent to Mojado Formation)**—Shale, sandstone, and some thin beds of conglomerate and limestone

s **Sandstone marker bed**—Locally quartzitic

c **Site of clam fossils**

w **Site of silicified wood**

Km **Mural Limestone (equivalent to U-Bar Limestone)**

oy **Site of oyster fossils**

Kg **Glance Conglomerate**—Limestone pebbles

Horquilla Limestone (Lower Permian and Pennsylvanian)—Mostly medium-bedded, cherty, light-gray; some reddish-gray siltstone in upper part

F **Site of large fusulines, probably Permian**

f **Site of small fusulines, Permian or Pennsylvanian**

Ch **Site of *Chaetetes*; Pennsylvanian**

PIPh **Paradise Formation (Upper Mississippian)**—Shale

Me **Escabrosa Limestone (Mississippian)**—Thick-bedded crinoidal limestone with large loaf-shaped chert pods

Om **Montoya Group undivided (Upper and Middle Ordovician)**—Mostly brownish-gray cherty dolomite

a **Base of highly cherty dark-grayish-brown Aleman Formation (Upper Ordovician)**

c **Base of Cable Canyon Sandstone (Middle Ordovician)**—Round quartz grains in dolomite

Oe **El Paso Limestone (Lower Ordovician)**—Light-gray, crinkly bedded cherty limestone and dolomitic limestone

OCb **Bliss Sandstone (Lower Ordovician and Upper Cambrian)**—Quartzite and sandstone

Yg **Granite (Middle Proterozoic)**—Coarse-grained, reddish-gray, altered

Contact—Dotted where concealed

45 **Normal fault**—Showing dip; dotted where concealed. Bar and ball on downthrown side

10 **Thrust fault**—Showing dip; dotted where concealed or intruded. Sawteeth on upper plate

Folds—Showing axis

Anticline

Overturned anticline

Syncline

Strike and dip of beds

70 **Inclined, direction of top beds unknown**

40 **Inclined, direction of top beds known**

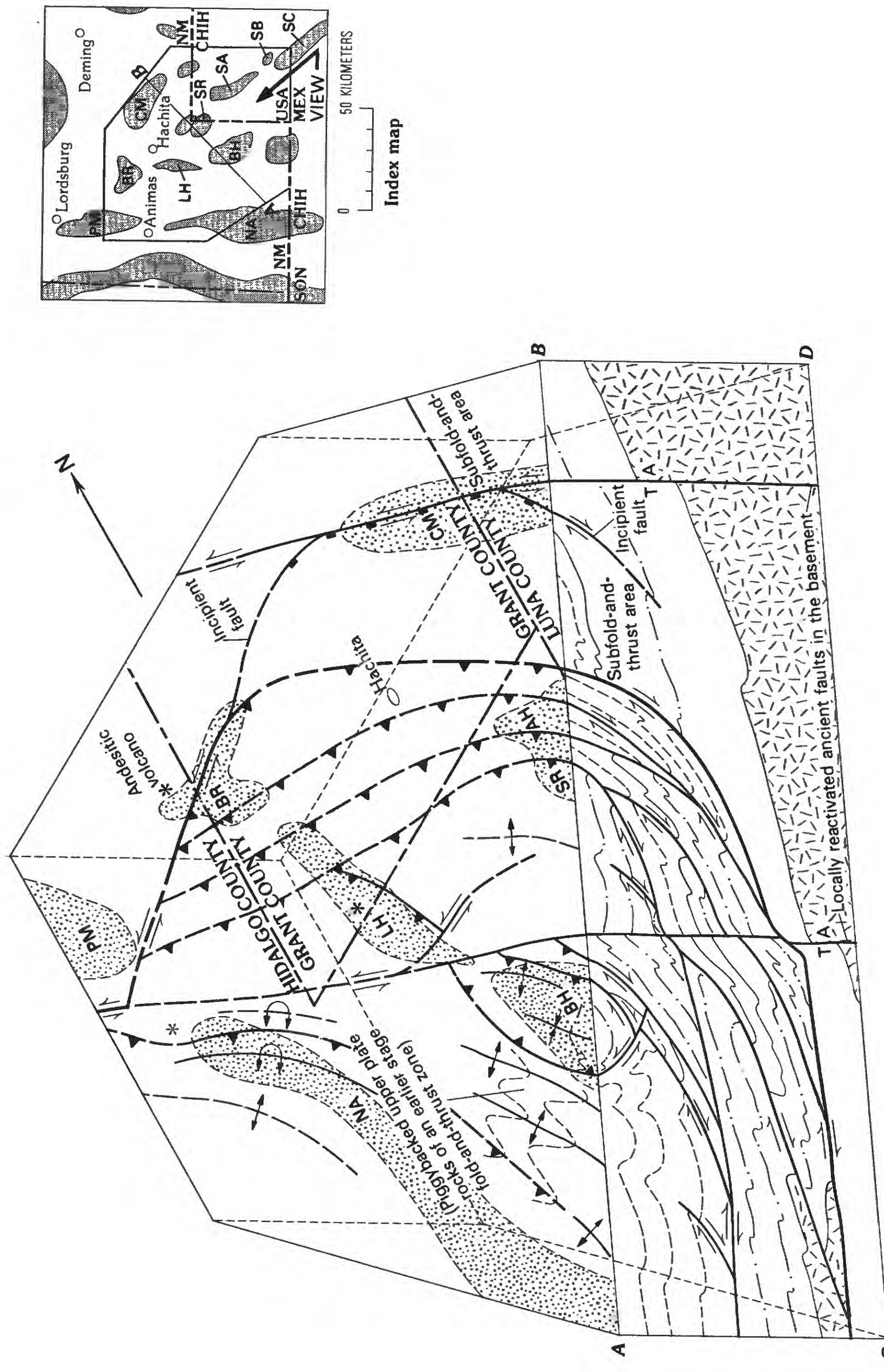
Vertical

85 **Overturned**

inclined toward the southwest. The terrane of shingled platelets is faulted over Lower Cretaceous rocks in the U-Bar Ridge foothills southwest of the Big Hatchets. These rocks are also folded, but the folds are large in amplitude (2–3 km) and length (6–10 km), a style that stands in marked contrast to the folds and faults of the southwestern flank of the Big Hatchets. Rather, the deformational style of the U-Bar Ridge foothills is characteristic of the fold-and-thrust zone in the Sierra Juarez area, for example. These large folds extend southward into the northern part of the Alamo Hueco Mountains (Zeller, 1959) (fig. 16), where an anticline was drilled to a total depth of 14,585 ft. The drill hole encountered the Lower Ordovician El Paso Formation (fig. 5) before reentering the Mississippian Escabrosa Limestone (Zeller, 1965, p. 118) at a depth of 14,120 ft.

The Brockman Hills lie along the northern part of the eastern intermediate zone. They are underlain mainly by Lower Cretaceous Bisbee Group rocks warped into folds with an anomalous east-west trend. To the northwest these rocks are thrust over a thin plate of Paleozoic rocks that in turn is thrust over andesitic rocks of the Hidalgo Volcanics (Corbitt and others, 1977; Thorman, 1977a). Fold orientation here indicates a northerly transport direction, and thrust faulting here, as in the Little Hatchet Mountains, postdates latest Cretaceous(?) andesite, and thus probably is Paleocene. Thrust faulting was reportedly also encountered in deep drill holes northeast of the Brockman Hills (sites not shown in fig. 16), but details of these finds are still unavailable.

A final observation should be made on the structural features of the northern Animas Mountains, before offering an interpretation of the eastern intermediate tectonic zone. This part of the range is underlain by Mesozoic and older rocks that are primarily thrust faulted and locally also folded (Drewes, 1986). Most thrust faults in the Paleozoic sequence bring younger rocks over older ones, and a major thrust fault emplaces Paleozoic rocks upon folded Lower Cretaceous Bisbee Group on the northeast flank of the range, as previously recognized by Soule (1972). However, more recent mapping fails to verify the overthrusting of the Paleozoic terrane by a plate of Precambrian granite, as mapped by Soule. Additionally, a capping thrust plate of overturned Pennsylvanian and Permian Horquilla Limestone must be a remnant of another major plate now largely eroded (Drewes, 1986). Recognition of the inversion of the succession was established by an inverted fusuline sequence (identification arranged by Sam Thompson, personal commun., 1975) and corroborated by inverted growth positions of *Chaetetes milliporaceous* sponge fossils.


Lordsburg
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Index map
N
A
B
C
D
T
Incipient fault
Subfold-and-thrust area
Andesitic volcano
HIDALGO COUNTY
GRANT COUNTY
LUNA COUNTY
NA upper plate
(Piggybacked rocks of an earlier stage fold-and-thrust zone)
Locally reactivated ancient faults in the basement

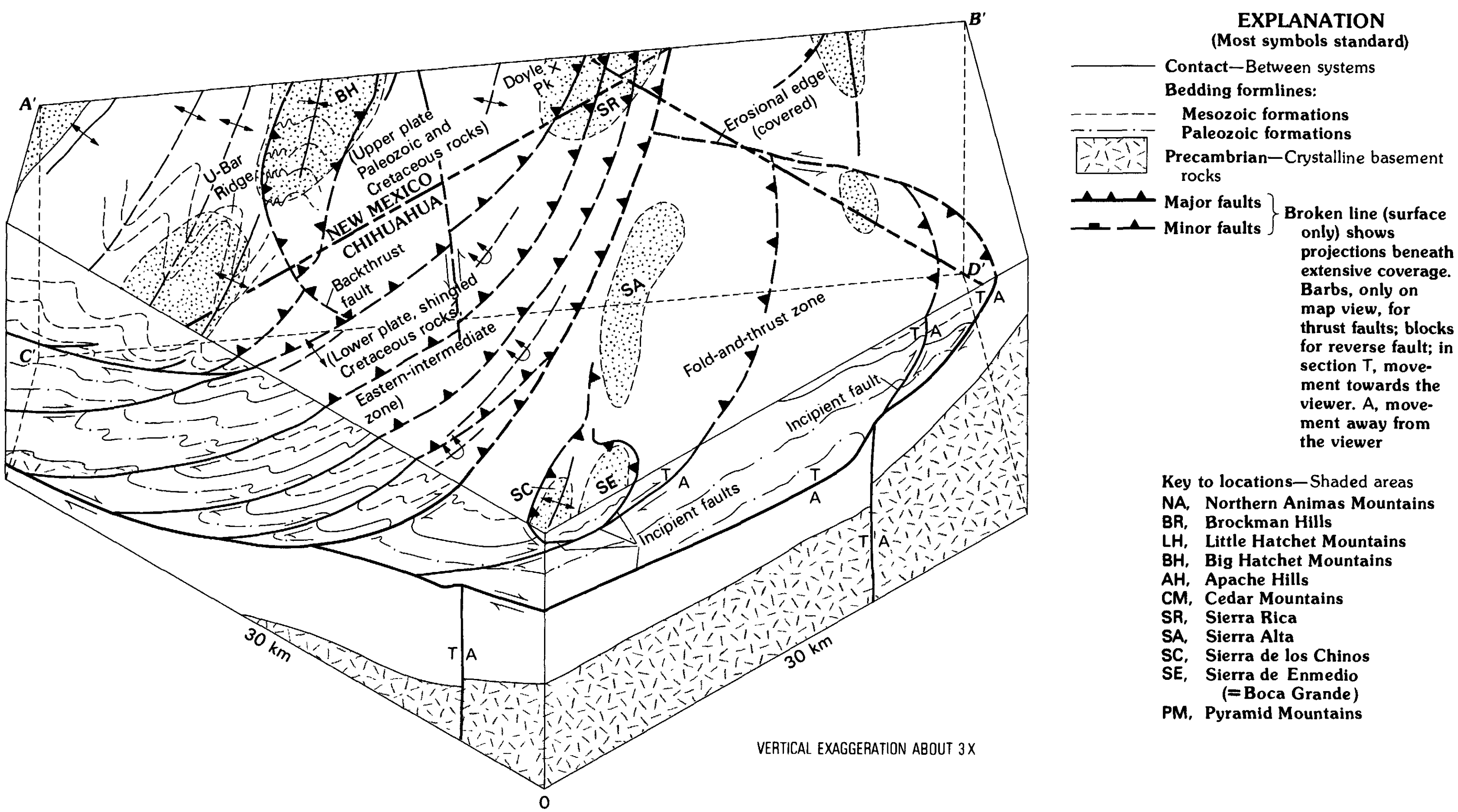


FIGURE 18.—Block diagram (pull apart) showing relations among structural features formed during the Cordilleran orogeny in northwestern Chihuahua and southwestern New Mexico. Mountains (shaded) are identified in key. Incipient faults, shown with minimal offset, develop slightly later than the others.

An interpretation of the structural features of the eastern intermediate zone is shown in figure 18. The essential features are a major structural break to the northeast, separating a thinner plate system of the fold-and-thrust zone from a thicker plate system of the eastern intermediate zone. This change in thickness is marked by a shift of a main décollement position from the incompetent Permian rocks in the east to a position at the base of the Paleozoic sequence in the west. Northwest of Hachita this major structural break merges with a tear-fault and thrust-fault system lying along the north side of the Brockman Hills, and thence trending west beneath the volcanic cover of the southern part of the Pyramid Mountains. A splay of the (older?) tear fault may extend eastward to the Victorio Mountains (fig. 16). Within the eastern intermediate zone, a northeastern belt contains shingled thrust plates of Cretaceous rocks, including both Upper Cretaceous Ringbone Formation that may be a foreland basin deposit of an earlier stage of the orogeny than that represented here, and the northeastern belt contains the late orogenic andesitic Hidalgo Volcanics. These thrust plates are overthrust by other plates of Paleozoic rocks. Remnants of a fold-and-thrust zone style of deformation are preserved alongside those reflecting more intense (deeper seated?) deformation of eastern intermediate zone folding. Reactivation of segments of some faults, believed to be part of the basement flaw system, caused the offset of thrust faults toward the end of the Cordilleran orogeny.

The construction of the block diagram of figure 18 incorporates several assumptions and compromises. While the focus was intended to show conditions at the close of the orogenic deformation in the eastern intermediate zone, about 60 m.y. ago, some younger deformation to the northeast that formed only slightly later is shown as "incipient structures." Probably some mild deformational responses to the waning time of orogenic deformation also continued into the 60–55 m.y. period within the eastern intermediate zone. Also, some deformation shown, such as the folding of the U-Bar Ridge, was formed earlier than most of the other structures of this zone. Additionally, there is little control on the amount of tectonic transport that has taken place in the eastern intermediate zone. An amount of 50 km was assumed in this construction but 30 or 80 km could also have been used. Finally, the diagram does attempt to show some key geologic relations, such as: a likely interaction between an incipient thrust fault of Paleozoic rock with an older basement flaw of the Cedar Mountains; a plate of Paleozoic rock over shingled Mesozoic rock in the Sierra Ricas; the backthrusts of the southwestern Big Hatchet Mountains; and the overturned plate in the northern Animas Mountains. However, the diagram does not account for other features, for example: the presence of some Precambrian granite in plate position in the Sierra Rica is unaccounted for; the presence of little-disturbed rock in the Big Hatchet Mountains and the more deformed rocks to the northeast and northwest are not shown at this scale; and the northwest-trending basement faults are underemphasized. Furthermore, in the interests of interpreting structural relationships, I have paid insufficient attention to the likely past distribution of foreland basin deposits and early post-orogenic volcanic deposits now largely eroded. The impact of their past distribution and their genetic ties to the orogenic development will be discussed later in this report.

A final tectonic enigma: occurring at four sites in the southwestern Big Hatchet Mountains (Drewes, 1990a, 1990b) and at one place in the northern Animas Mountains (Drewes, 1986) is the presence of small plugs of gypsum. While some Permian rocks of the southwest have gypsum crystal casts, none of the rocks near gypsum plugs shows such features. Consequently, the suspicion is that the plugs have flowed upward from other rock sources than those known at the surface. Perhaps the Jurassic rocks, known to be gypsiferous in much of central Chihuahua, and known to extend at least to northern Chihuahua and possibly into south-central New Mexico, have a wider distribution in the subsurface. This intriguing, albeit remote, possibility also was not portrayed in figure 18. This topic will be brought up again in more detail in the next section, in connection with a site near Tucson.

Rocks of Paleozoic and Mesozoic age exposed in the next ranges to the west of the Animas Mountains, the Peloncillo Mountains along the New Mexico–Arizona border and in that part of the Chiricahua Mountains lying northeast of the Apache Pass fault zone are also folded and thrust faulted in the tectonic style of the ranges of the "bootheel" of southwestern New Mexico (fig. 19). Shingled thrust plates, tight folds, and no substantial involvement of basement rocks are the characteristic features of the Peloncillo and northeastern Chiricahua Mountains. Late orogenic volcanic rocks are more voluminous than to the east and are related to specific centers, cored by genetically related granitoid stocks along the northern margin of the deformed terrane. These stocks likely are exposed by the deepest level of erosion of that terrane. In contrast, the next few ranges southwest of the Apache Pass fault zone, such as the Dragoon Mountains, show characteristics sufficiently different from the eastern intermediate zone to group them separately as the western intermediate tectonic zone.

WESTERN INTERMEDIATE ZONE

The western intermediate tectonic zone is characterized by a tectonic style and history of greater complexity than that of the eastern intermediate zone. It appears to mark the deepest level of dominantly brittle deformation, distinctive from the more ductile style of deformation of the hinterland tectonic zone. In addition to shingled thrusts, the allochthon contains some major thrusts that repeat thick sequences of rock. Some of the overriding plates contain crystalline basement rock. Thrust faults emplacing younger over older rocks are as common as those bringing older over younger ones. Backthrusts continue to be interspersed with thrust faults and may be larger structures than those toward the east. Folds among the shingled thrust plates remain mostly small tight structures markedly inclined toward the genetically related nearby thrust faults or backthrusts; however, some folds are large, strongly inclined or recumbent structures. Orogenic volcanic rocks probably were more voluminous, resulting in a thicker and more widespread—perhaps continuous—cover of andesitic rocks. The volcanic roots of this andesitic cover were penetrated by many granitic to dioritic plutons. Associated with some of these orogenic (to early postorogenic) igneous centers are rhyolitic ash-flow sheets that spread from calderas. Metamorphic Phanerozoic rocks occur at several places in the eastern part of this zone and are widespread (or semiregional) in the western part. While greenschist grade of metamorphism is common, amphibolite-grade rocks are also present locally. In many ranges, several Cretaceous and Paleocene stages of tectonic development are recorded. The combined effects of the characteristic tectonic style and of the more involved geologic history of the western intermediate zone suggest that they were developed closer to the orogenic core, perhaps beneath a thicker cover, and certainly over a longer span of time.

Although some of these tectonic features of the western intermediate zone sound markedly different from those of the eastern intermediate zone, others are gradational or incremental. The eastern boundary of the western intermediate zone is taken, somewhat arbitrarily, at the first (or easternmost) regional occurrence of crystalline basement rocks in thrust plates, local occurrence of amphibolite-grade metamorphism of Paleozoic rocks, and a widespread polyphase deformation. The eastern boundary of the western intermediate tectonic zone appears sharply developed and steep only because of its local control by the Apache Pass fault zone, a northwest-trending reactivated basement flaw (Drewes, 1981a, 1981b, 1982a). Likely, this boundary flattens as it diverges from the fault zone to the south beneath the extensive cover of younger rocks of southeasternmost Arizona and adjacent Mexico (Drewes, 1985a, 1985b; Drewes and others, 1988).

In the following description of the style of deformation of the western intermediate zone, features of selected ranges will be presented from east to west (beginning with the northwestern part of the Chiricahua Mountains near the Apache Pass fault zone (fig. 19)). Many of the essential geologic features are shown by Wilson and others (1969); others represent new data. Thrust faults cut Mesozoic and Paleozoic rocks on both sides of the Apache Pass fault zone, and folds are both small and uncommon features. These faults consist of mixed older-over-younger and younger-over-older thrust faults, with the upper plate transported east-northeastward oblique to the trace of the Apache Pass fault zone. The rocks southwest of the zone include a plate of Precambrian granodiorite or granite between plates of Lower Cretaceous Bisbee Group. In addition, the basal formation of the Bisbee, the Glance Conglomerate, is thick and coarser grained to the southwest of the Apache Pass fault zone, and in the adjacent Dos Cabezas Mountains within the zone; whereas, the conglomerate is thin and finer grained northeast of the zone (Sabins, 1957a; Drewes, 1980, 1981b, 1982b, 1984, 1985a, and 1985b). The combination of having a plate of Precambrian crystalline rock among plates of Cretaceous sedimentary rocks, and the juxtaposition of diverse facies of the Cretaceous rocks is indicative of large amounts of tectonic transport along thrust faults, some of which must be rooted at deeper structural levels than those occurring farther east in the Cordilleran belt.

Cambrian and Devonian argillaceous rocks along a several kilometer stretch of the Apache Pass fault zone near old Fort Bowie (Drewes, 1984) contain the easternmost known effects of amphibolite-grade metamorphism. The rocks contain large porphyroblasts of andalusite after kyanite and of biotite after staurolite, as well as chloritoid, tourmaline, and graphite(?). These metamorphic rocks lie along the Apache Pass fault zone; there are no plutons near them. Plutons that appear along other parts of this fault zone have not generated such metamorphism in their contact zones. This mineral assemblage suggests an initial response to high pressure, followed by a second response to another period of high pressure or temperature. A pressure of an 8- to 10-km-thick cover would lead to the development of the initial assemblage. The local occurrence of these minerals along the Apache Pass fault zone suggests that the cover thickness was 8 km or a bit less, and that pressure enhancement took place on the stoss side of a ramp against which a thick thrust plate moved onto a raised basement mass northeast of the Apache Pass fault zone.

The northern Chiricahua Mountains provide the easternmost evidence of two stages of Cordilleran defor-

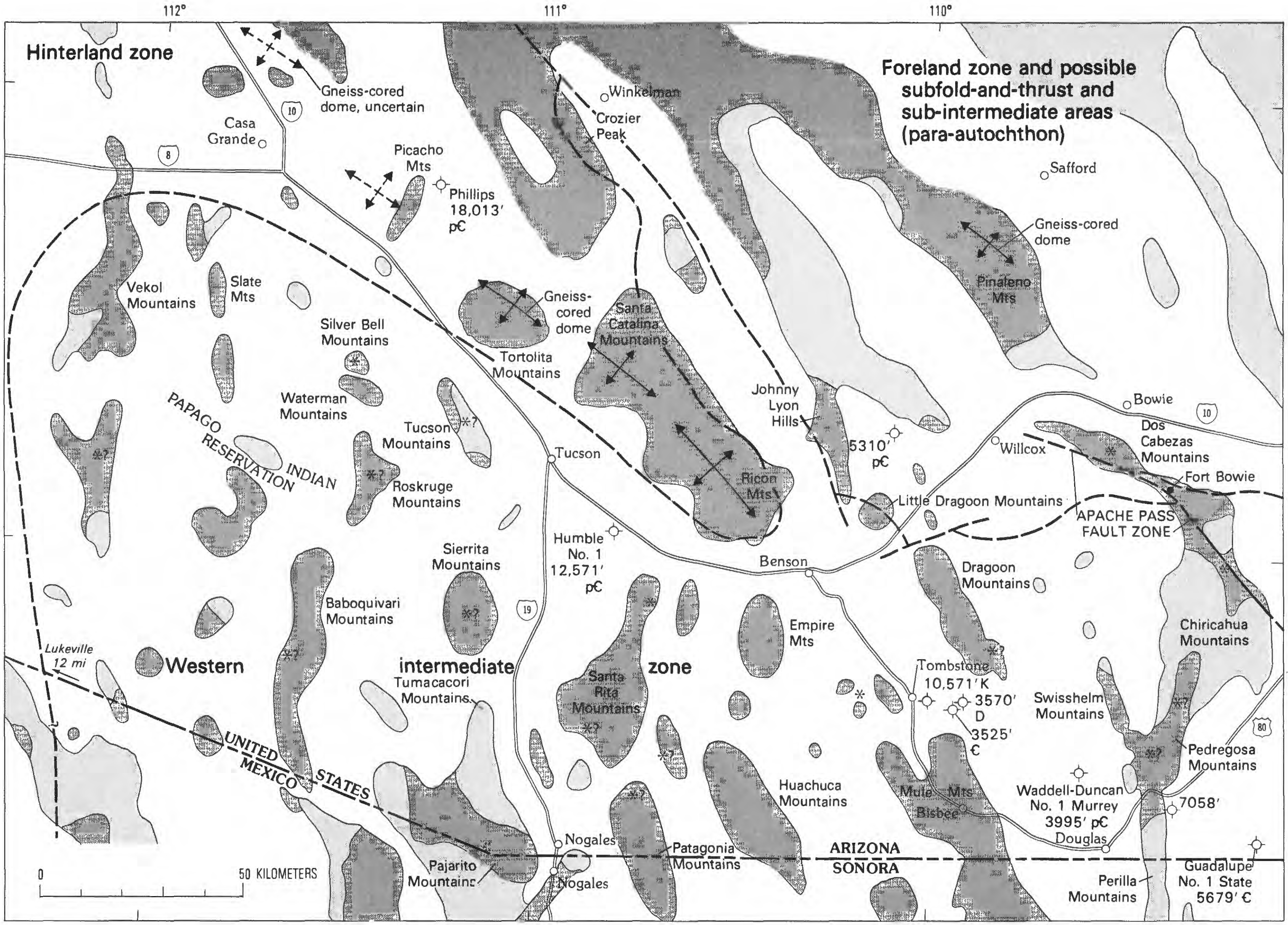


FIGURE 19.—Index map of the western intermediate zone, showing major outcrops of pre-Tertiary rocks (heavy shading) and Tertiary covering rocks (light shading). Key oil wells located, giving total depth, in feet, and age of rock at bottom of hole (K, Cretaceous; D, Devonian; €, Cambrian; p€, Precambrian). Volcanic centers of orogenic age shown, *.

mation. Thrust-faulted Paleozoic and Mesozoic rocks occur on both sides of the Apache Pass fault zone and its few branch faults, such as the Emigrant Canyon fault (Sabins, 1957a; Drewes, 1982a). During the first stage, thrust faulting resulted in alternating platelets of Paleozoic and Mesozoic rocks that are nearly upended where they cross Emigrant Canyon. During the second stage these platelets underwent 2 km of left slip along the Emigrant Canyon fault before the emplacement of an Oligocene granite stock. Apparently, most of the compressive deformation occurred early, perhaps during the Late Cretaceous, and minor deformation occurred during the second phase of compression, perhaps during the Paleocene. More constrained ages on similar movement phases are available in the Santa Rita[3] Mountains (fig. 19), as presented by Drewes (1972b) and are summarized in a later section of this report on the "Age of Deformation."

Paleozoic and Mesozoic rocks of the Pedregosa and Swisshelm Mountains show other aspects of the style of deformation of the western intermediate zone, although the intensity of deformation is generally less than it is in the northern Chiricahua Mountains near the northeast margin of the mobile belt. At several places in all three mountain ranges, Paleozoic rocks are thrust faulted over Mesozoic rocks (Epis, 1956; Cooper, 1959; Drewes, 1980; and Drewes and Brooks, 1988). Of particular interest are the indications of polyphase deformation given by diversely oriented small folds and the presence of large (late phase) backthrusts. Also, the Pedregosas have the easternmost occurrence of dated Late Cretaceous rhyolite ash-flow tuffs and of voluminous andesitic lava flows.

The Deep Well Canyon area of the southern Pedregosa Mountains shows essentially a western terrane of Late Cretaceous or early Tertiary andesitic lava flows overlain by an eastern terrane of Lower Cretaceous Bisbee Group rocks (fig. 20). These terranes are separated by a fault that dips about 45° northeast and that probably had two phases of movement. The Bisbee Group is deformed by two sets of folds: a more abundant northwestward-trending set inclined steeply to the northeast that is probably related to the backthrust development, and a less abundant northeast-trending set that is recumbent and may be related to subhorizontal adjustments on a late-phase strike-slip movement on the same backthrust fault beneath the Bisbee. Both deformations took place during the interval between the deposition of the thin sheet of rhyolite-welded tuff in the andesite (sample taken from at a site 1 km west of the southwestern corner of the area of figure 20) dated as 73.6±2.6 m.y. from K-Ar on biotite, and the emplacement of the rhyolite plug shown at the northern side of figure 20 (sample taken from a site 1 km from the northwest corner of the plug) dated at 22.5±2.0 m.y. from fission track on zircon (Drewes and Brooks, 1988).

In the Swisshelm Mountains a reverse fault or the steep part of a backthrust cuts a large north-northwest-trending overturned syncline (fig. 21). Because the overturning of the fold is most pronounced where the fault closely parallels the fold, it is believed to have been overturned during faulting. These structures appear to involve only the rocks of an upper major thrust plate; similar complications are not known in the lower plate comprising essentially the same stratigraphic sequence plus the Late Cretaceous and early Tertiary andesitic lava flows (Cooper, 1959; Drewes, 1990c). This deformation predates the emplacement of an Oligocene granite stock low on the west flank of Swisshelm Mountain. The features of both the Swisshelm and Pedregosa Mountains are relatively minor ones and characterize some exceptions to the dominant tectonic style of the zone, to be discussed next.

Structures of the Dragoon Mountains are typical of the tectonic style of the western intermediate zone (fig. 19). These mountains are underlain by multiply deformed Precambrian, Paleozoic, and Mesozoic rocks that are intruded by granitic stocks of Jurassic to Miocene ages. Deformation is strongest to the north and weakens to the south, a characteristic of many of the linear ranges of the allochthon. Deformation in other small groups of hills north of the Dragoons is markedly weaker. In style, the deformation of the hills north of the Dragoon Mountains resembles that of the foreland zone; therefore, this group of hills may be viewed as para-autochthonous (Drewes, 1980, 1981a).

The structurally most complex part of the Dragoon Mountains northwest of the Stronghold stock of Miocene age is described below in terms of its three major thrust plates (fig. 22).

The lowest major thrust plate, LP, constitutes the northeastern part of the mountains and is made up of upended, metamorphosed, thrust platelets of upper Paleozoic and Lower Cretaceous rocks. Commonly, the Paleozoic carbonate rocks are recrystallized; marble occurs in some places, and the shales are argillic to phyllitic. Conglomerate clasts are stretched; some rocks have a cleavage. The stratigraphic sequences in these platelets face west, that is, they go up-section to the west. The Lower Cretaceous rocks of plate LP include a basal conglomerate 300 m thick that is rich in cobbles and locally include boulders and landslide masses.

[3]Here and hereafter the Santa Rita Mountains refers to the mountains of that name near Tucson, Arizona, and not the mountains in Chihuahua, Mexico.

The middle major thrust plate, MP, comprises the northwestern part of the Dragoon Mountains and is made up of upended, metamorphosed, thrust platelets of Precambrian basement rock and lower Paleozoic rock. The Paleozoic sequences of these platelets face east. Kyanite occurs in some of the Precambrian phyllite; as kyanite is not commonly found in this rock, the low-amphibolite grade of metamorphism indicated by this mineral probably reflects the local metamorphic conditions related to Cordilleran tectonism, rather than metamorphism that occurred during the Precambrian.

The upper major thrust plate, UP, makes up some of the southern part of the area, especially in Fourr Canyon and on Mount Glenn. It consists of subhorizontal to gently dipping unmetamorphosed upper Paleozoic, and Upper and Lower Cretaceous rocks. The basal conglomerate of the Lower Cretaceous rocks is a 2- to 3-m-thick pebble- to cobble-bearing unit. The Upper Cretaceous rocks are distinguished by their volcaniclastic beds. At the head of Jordan Canyon the thrust fault benaeath plate UP merges with, or grades into, a steeply southeast-dipping left-lateral strike-slip fault.

The telescoping of sedimentary facies and metamorphic facies is equally impressive. Thick coarse-grained Glance Conglomerate of plate LP occurs but a kilometer from the thin fine-grained Glance Conglomerate of plate UP. While conglomerate thickness and coarseness may change in facies over a fairly short distance, the change from 3 m of pebble conglomerate to 300 m of cobble conglomerate and coarser deposits

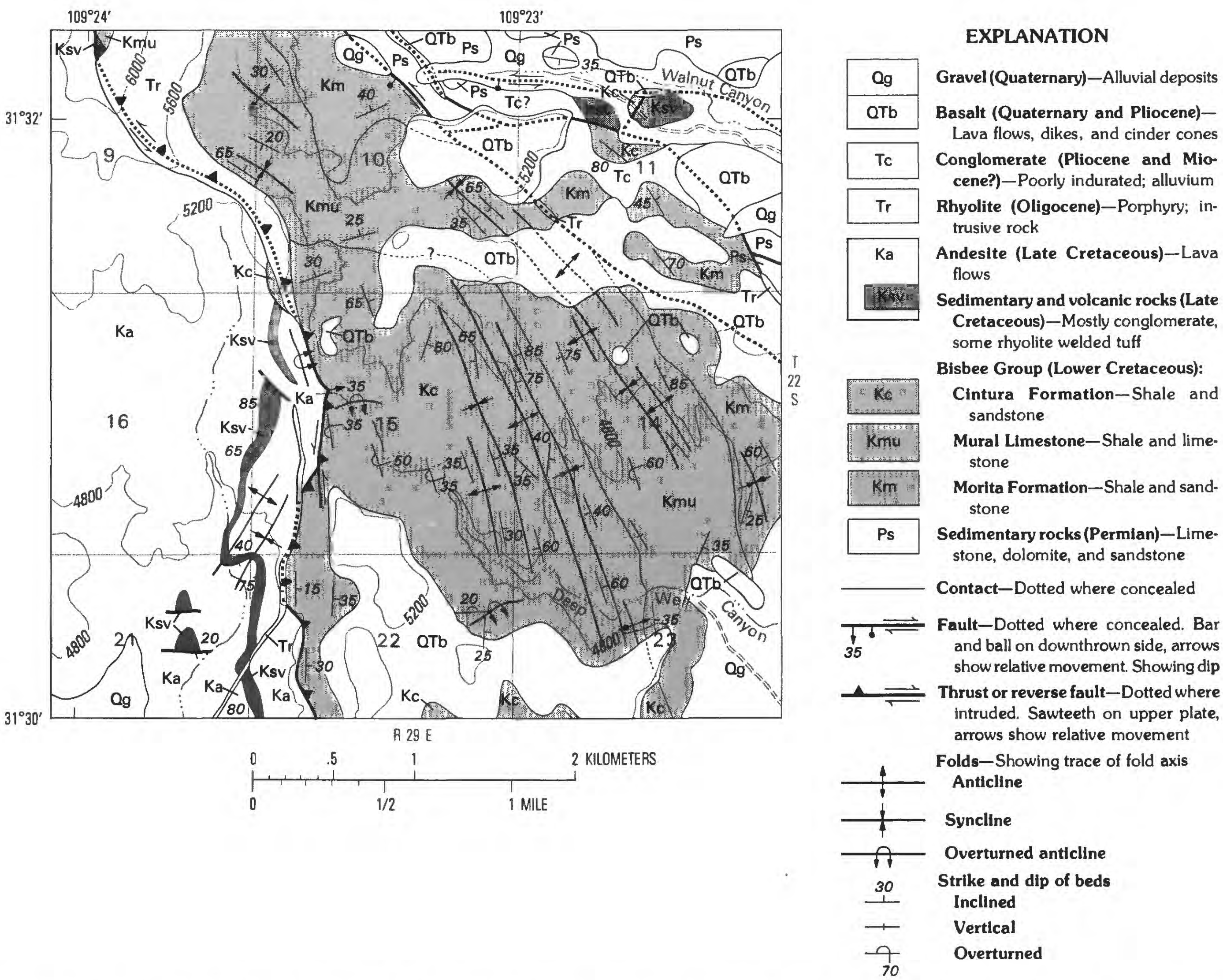


FIGURE 20.— Geologic map of the Deep Well Canyon area, southern Pedregosa Mountains, southeast Arizona, showing structural features from which several phases of movement are inferred, the youngest a Late Cretaceous(?) backthrust movement.

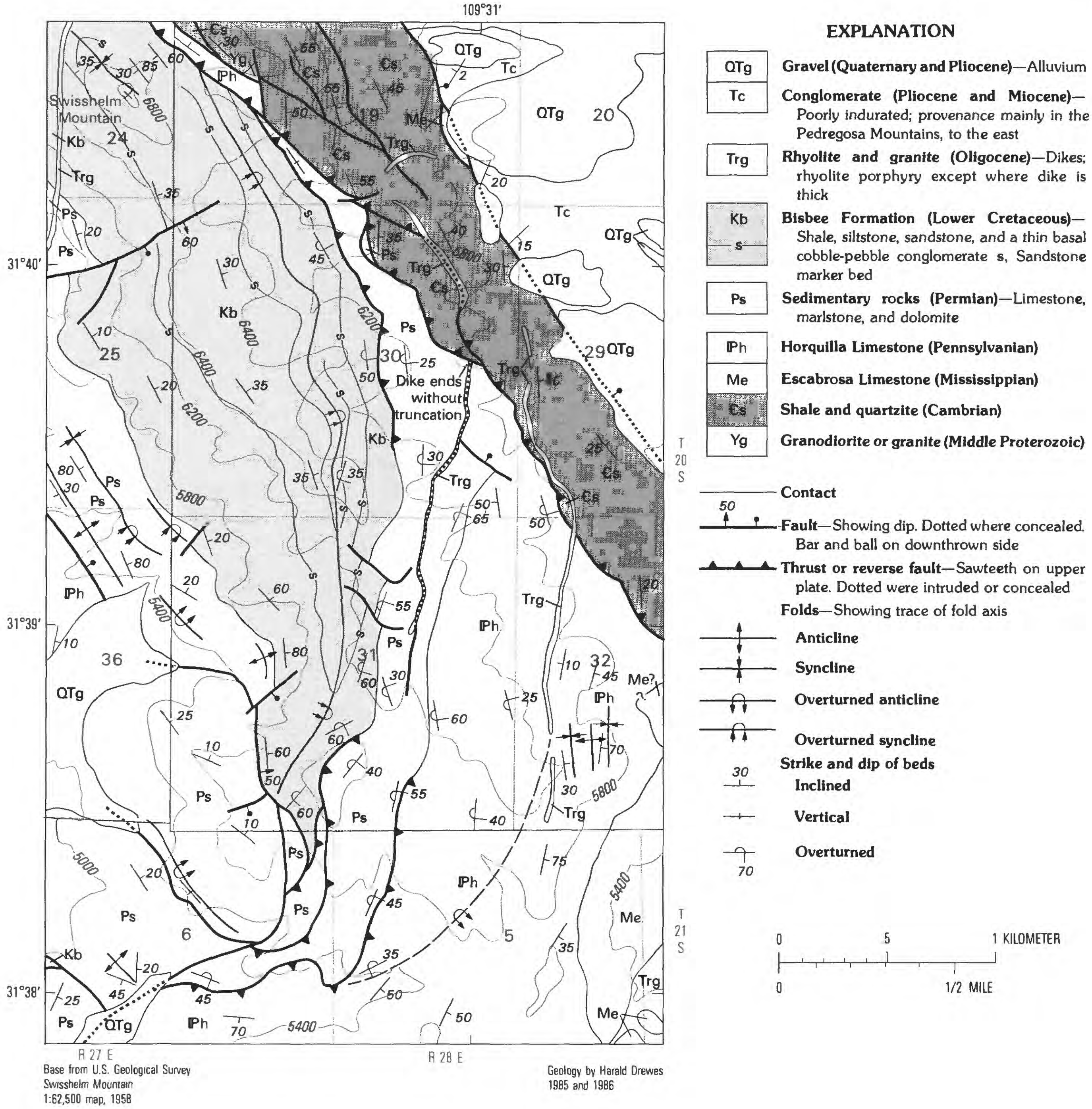


FIGURE 21.—Geologic map of a part of the Swisshelm Mountains, southeastern Arizona, showing westward-inclined folds beneath a major back thrust fault.

probably took place over a distance of a few kilometers, if not a few tens of kilometers. Although the metamorphic grade of the lower plates has not been established, all the carbonate formations are marmorized, not only near the early Oligocene stock, the only nearby source of heat known, but also far from that body. Rocks of UP are unaltered, even to the extent that delicate foraminifera are still preserved, except within about a kilometer of the Stronghold stock where contact metamorphism and some skarn mineralization occur (Drewes and Meyer, 1983). Again, such a juxtaposition of metamorphic facies with nearly unmetamorphosed rock implies a considerable but undetermined amount of tectonic transport—perhaps a minimum of the range width, or about 10 km, easily commensurate with the amount of tectonic telescoping proposed for the juxtaposition of diverse sedimentary facies.

The opposite directions of facing of the stratigraphic sequences in the lower and middle plates (LP and MP) implies the presence of a remnant of a major overturned fold. Small folds are also present in some of the thrust platelets of the lower plate. These resemble the small folds in other ranges of the intermediate tectonic zones, folds that are inclined at moderate angles to the nearby small thrust faults and are vergent to the northeast. A similar vergence is recorded in the structures of the central part of the Dragoon Mountains, south of figure 22, and strike-slip faults typically trend east-northeast with left slip.

These complex structures of the northwest end of the Dragoon Mountains are inferred to have developed in two stages. The relations between plates LP and MP formed during the first stage. The upended major thrust plates, LP and MP, flatten to the southwest downdip, bringing the eastern major plate (of younger rocks) beneath the western one (of older rocks). The sequences in the eastern platelets thus are upright; whereas, the sequences in the western plate are overturned. This requires a northwest-trending major recumbent fold with a gently southwest-dipping axial plane, and an upper, or western limb, all now eroded, as implied by Gilluly (1956) and more fully shown by Drewes (1987). During the first stage, the northern part of plates LP and MP was upended and metamorphosed, possibly reflecting the extra stress complications along the edge of the mobile belt or a relatively deep structural position during initial deformation. During the second stage of deformation, plate UP was transported from far to the southwest across the folded, thrust-faulted, and metamorphosed rocks of plates LP and MP. Clearly, the intensity of the deformation is greater in the western intermediate zone than it is in other zones to the east, and the present level of exposure lies deep in the allochthonous system.

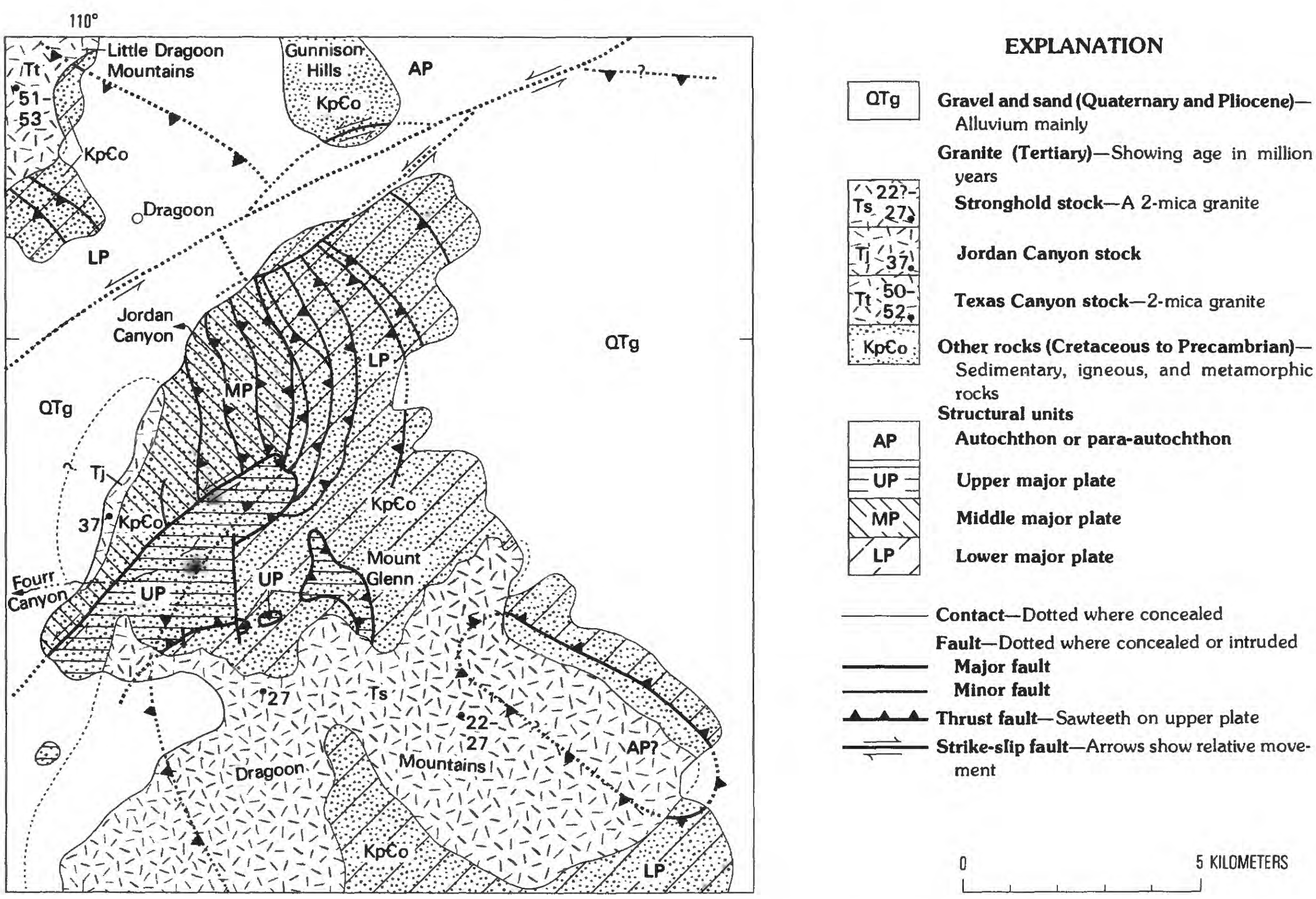


FIGURE 22.— Generalized geologic map of the northern part of the Dragoon Mountains, southeastern Arizona, showing the main tectonic terranes.

At least two of the major thrust plates, LP and UP, are mapped, almost without break, from the northern part of the mountains to the central part and, with some lateral offset, onto the eastern part of the southern Dragoon Mountains, where the Jurassic Gleeson stock overlies the Cretaceous and Paleozoic rocks on the middle thrust fault (Drewes, 1980). The structural relationship between the granite plate of the Gleeson stock and the underlying sedimentary plate, LP, is deduced from detailed mapping (Gilluly, 1956; Drewes, 1981a, pl. 4) and is verified through drilling (unpub. mining industry data) as much as 2 km behind the trace of the fault. These general relations have been projected from the central part of the Dragoon Mountains west-southwestward about 10 km (Drewes, 1980, sec. *D–D'*), where the southwestern side of the Gleeson stock was arbitrarily placed. However, a deep exploratory oil well subsequently drilled close to the line of this section and 2 km southeast of Tombstone penetrated Precambrian granite but reentered Lower Cretaceous Bisbee Group at about 10,000 ft (3 km), the approximate position of the projected Cochise thrust fault of Drewes (1980, sec. *D–D'*). Tempting as the acceptance of this major thrust fault interpretation is, an alternative solution, involving offset on a high-angle fault that trends southwest from Tombstone toward the well, is also possible; further testing is needed.

Rocks in and near the Dragoon Mountains have also received attention by other workers, most with study areas of local extent and none offering published maps in support of their alternative interpretations. One such local study of a part of the central Dragoons led Keith and Barrett (1976) to suggest that the nearly upended thrust faults of the east side of the range were normal or reverse faults. While some of their structures may be followed laterally into thrust faults, others may indeed be reverse faults, and, as I shall propose later, may be remnants of a foreland stage of development as that early deformation stage swept through this part of the orogenic belt.

Another such local study was made in part of the Little Dragoon Mountains (fig. 19) by Dickinson (1984, p. 5) without remapping the detailed work of Cooper and Silver (1964). The low-angle Lime Peak fault mapped by Cooper and Silver as a thrust fault, and so compiled by Drewes (1980), was simply reinterpreted by Dickinson as a low-angle normal fault, apparently because younger rocks lie upon older ones. From this interpretation Dickinson extrapolated that no regional thrust faults exist in southeast Arizona.

The coarse clastic facies of the Lower Cretaceous Glance Conglomerate of the Dragoon Mountains reflects unusual local topographic conditions at the time of deposition that has suggestive tectonic implications. Such local thick and coarse-grained deposits have been reported at several localities in southeastern Arizona, although more commonly it has been reported that the conglomerate is only 1–3 m thick and is a finer grained variety (Ransome, 1904; Sabins, 1957a, 1957b; Hayes, 1970; Finnell, 1970; Drewes, 1971a; Bilodeau, 1979, 1982, 1983; Drewes and Hayes, 1983). While high local relief has been recognized by all these workers to have been present, only Bilodeau (1979, 1982) indicated that relief was the result of tensional stress and rifting, albeit without providing support for his opinion and without blending into such an explanation the record of concurrent compressional deformation in western Arizona.

Local high relief, of course, may be generated in ways other than response to tensional stress; and a more plausible interpretation is to have the relief generated through compression, not only known to be active to the west (see sections on "Hinterland Tectonic Zone" and on "Age of Deformation"), but present throughout the region shortly after Glance (and Bisbee) time. As Keith and Barrett (1976) inferred, some of the faults along the northeast flank of the central part of the Dragoons may be reverse faults. Within a concept of gradual eastward spreading of orogenic conditions across all parts of the Cordilleran belt, it seems more likely that this part of the western intermediate zone once lay in the fold-and-thrust zone, and before that in a foreland zone. During that time when compressive stress was first propagated into the area, initial response, by analogy with the present record in New Mexico, would have included some reverse faults or upthrusts of the Bear Peak fold-and-fault zone type well in front of the main orogenic front. Voluminous coarse-grained deposits would have been shed off these local raised blocks, while the surrounding region would have received the finer grained conglomerate and its overlying shale and sandstone of the Morita Formation. Ultimately, the main orogenic conditions would propagate eastward to the present area of the Dragoons and would overwhelm this early orogenic style of deformation with the more intense compressional style. Such an explanation offers an advantage over that of tension-related relief in that it eliminates an accordian-like variation of stresses—first tension, then compression, and later more tension, and it accommodates the inference of reverse fault movement on some nearby Cretaceous faults. This local aspect of the tectonic development will be reviewed more fully in connection with other ties between the sedimentation and tectonic records, the focus of a later part of this report, "Key Sedimentary Features of the Orogenic Belt."

Of the four granitic stocks of the Dragoon and Little Dragoon Mountains (Gilluly, 1956; Drewes, 1980), the Gleeson stock (Jurassic) provides a possible measure of a large amount of tectonic transport; the Texas Canyon stock (Eocene) may be an eastern member of a 2-mica-garnet-bearing granite suite (hereafter just referred to as "2-mica granite"); and the others are post-orogenic (Oligocene and Miocene). In the southern part of the Dragoons, the Gleeson stock, of Jurassic age, and its host rocks may have a deeper level counterpart in the Santa Rita Mountains, 100 km to the south-southwest (Drewes, 1981a). The Gleeson stock, which is known to be underlain by Cretaceous sedimentary rocks in at least its northeastern part, may be the beheaded and transported upper part of the Squaw Gulch Granite of the Santa Rita Mountains to the west-southwest plus correlative rocks in adjacent parts of the Patagonia Mountains and hills near Nogales. While some arguments favor this correlation, strong support remains wanting.

The Texas Canyon stock northwest of Dragoon village (fig. 22) is dated as early Eocene age (53±3, 51±3, 49.5, and 48.5 Ma; Marvin and others, 1978) (late Paleocene in some older accounts, based on older editions of the geologic-time scale). It is a 2-mica body typical of stocks related to the development of the gneiss-cored domes of the hinterland tectonic zone. It lacks the foliation characteristic of the oldest of this group of stocks, and it is too old to be the youngest of the 2-mica stocks. Thus, the Texas Canyon stock may be comparable to an intermediate phase, such as the main part of the Youtcy or the Mica Mountain stocks of the Rincon Mountains, to be referred to in the next section of this report.

In the western intermediate zone west of the Dragoon Mountains, both severely deformed and little-deformed pre-orogenic rocks are found, with the abundance and variety of orogenic igneous rocks increasing. Many of these features have already been described in the literature and so only a few key features will be noted herein. In the Santa Rita Mountains (fig. 19), for example, two phases of deformation are dated, using presently accepted constants, at about 76 and 57 m.y., or, respectively, as Campanian and Paleocene (using constants of the last decade, about 73 and 55 m.y.; Drewes, 1971a, 1972b). Effusion of voluminous andesitic volcanic rocks predates the early deformation phase, and deposition of slightly less abundant rhyolitic welded tuff postdates this early deformation phase. Several large stocks of granite to diorite composition were emplaced in the early deformed rocks during the time interval 71–69 m.y., and other small stocks are of Paleocene age of which some predate and others postdate the second deformation phase. Mild contact metamorphism occurs around these stocks and extends sufficiently far beyond the usual range of purely contact effects to suggest a transition to regional metamorphism. Evidence from fold vergence and drag folds suggests that tectonic transport was chiefly northeast, with some late movement (or backthrust) to the southwest; other transport was to the northwest, in situations suggesting local gravity-impelled movement and responses to local stress fields near reactivated northwest-trending basement faults.

Finally, thick deposits of late pre-orogenic to early orogenic sedimentary rocks provide evidence of paleogeographic conditions prevailing during a time interval not well represented to the east. These deposits, the Fort Crittenden Formation, now lie unconformably on the Bisbee Group and probably were deposited in one of the last stages of a foreland basin. The Fort Crittenden Formation has a faunal assemblage of Santonian to Maastrichtian (Drewes, 1971a) and now is thought to represent the older part of this age range. Volcanic clasts in the upper part of the Fort Crittenden Formation may have been derived from the pre-Bisbee volcanic terrane west of the area or from local raised blocks in which were exposed thick rhyolite to andesite deposits of Late Triassic to Early Cretaceous age.

Many geologic characteristics similar to those of the Santa Rita Mountains are present in the Sierrita Mountains (fig. 19). Mesozoic igneous rocks, metamorphism, and mineralization associated with the igneous activity (Cooper, 1973; Drewes and Cooper, 1973) are also widespread in the Sierrita Mountains.

In the Baboquiviri Mountains, west of the Sierritas, a mildly metamorphosed sequence of sedimentary and andesitic volcanic rocks at least 5 km thick overlies a Late Jurassic batholith and is intruded by a Paleocene stock (Haxel and others, 1980). These volcanogenic deposits are both thick and widespread as far east as the Santa Rita Mountains. The greater abundance of correlatives of volcanogenic cover to the west relative to the size (extent) of the ranges, and lesser abundance to the east may reflect postorogenic erosion rather than nondeposition. Structural features in the Baboquiviri Mountains are recognized to have formed through compressive deformation, although deformation apparently was not severe (personal commun., presented in connection with Haxel, 1981).

Tectonic complications increase to the north in the chain of mountains that begin with the Baboquiviris and include the Waterman, Silver Bell, and Picacho Mountains (fig. 19). The Waterman and Silver Bell Mountains are underlain largely by Paleozoic and Mesozoic sedimentary rocks and by Cretaceous and Tertiary volcanic rocks. Paleozoic rocks of the Watermans are thrust faulted and unmetamorphosed; the Silver Bells are mainly of late orogenic and early post-orogenic Cretaceous and Paleocene volcanic and plutonic rock. A major northwest-trending strike-slip fault zone containing tectonic slices of metamorphosed Paleozoic rock separates the ranges and locally is intruded and mineralized. Still farther north, in the Picacho Mountains, foliated granite gneiss with hinterland zone affinity underlies mid-Tertiary volcanic rocks.

Finally, in the next tier of ranges in the Papago Indian Reservation west of the Baboquiviri–Silver Bell chain of mountains, rocks include some metamorphosed preorogenic sedimentary and volcanic rock (Haxel and others, 1978; Briskey and others, 1978). In most places, the original sedimentary formations are indeterminable, and so the recognition of structural features on stratigraphic succession is difficult to make. However, thrust faults are reported in several ranges, although commonly only in short segments far from other fault remnants. In the Slate Mountains (Blacet and others, 1978) and the Vekol Mountains (Docktor and Keith, 1978), for example, reconnaissance mapping shows the presence of Paleozoic and Mesozoic rocks in scattered sections cut by plutons and thrust faults (only small segments) or covered by Tertiary or Quaternary deposits. Typically, Late Cretaceous plutons cut thrust faults. In both ranges the Paleozoic and Mesozoic rocks appear to be mildly or moderately metamorphosed. In the Vekols some Precambrian basement rock is thrust faulted over Mesozoic conglomerate; in the Slate Mountains Devonian rocks are thrust over Proterozoic rocks. Structural analysis has not been made in either range and structure sections are unavailable. Again, terranes to the north of these ranges are part of the hinterland tectonic zone with a pervasive metamorphic fabric.

In summary, the western intermediate zone is marked by increasingly thick cover of Mesozoic rocks compared with zones to the east, more extensive involvement of plutonic and volcanic rocks, and associated metamorphism and mineralization. Where magmatism and metamorphism are most severe, the structural record is of a different character, seemingly fragmentary and obscure, so that the recognition of orogenic structures is difficult to make, and, I suspect, the structural relations have therefore been incompletely mapped. However, thrust faults of Cretaceous age are seen in many ranges but remain too fragmentary in the western half of the zone for any local synthesis.

HINTERLAND TECTONIC ZONE

A terrane of ductily deformed metasedimentary and metaigneous rocks lies generally west of, and locally beneath, the western intermediate zone of the Cordilleran orogenic belt. Typically, this terrane has a polymetamorphic history intertwined with several episodes of deformation that generated a subhorizontal planar fabric and, in most places, a northeast-trending lineation. Locally, these complexly deformed and metamorphosed rocks are bowed up into gneiss-cored domes or metamorphic core complexes. These metamorphic core complexes started to form late during the Cordilleran orogeny and then were involved with the late Tertiary extension. This part of the Cordilleran orogenic belt is believed to be a distal part of the hinterland zone whose core or proximal part lies at least 70 km west of the lower Colorado River, where a string of Jurassic and Early Cretaceous batholiths nearly obliterate the metasedimentary and metaigneous host rocks.

Hereafter, I shall avoid using the term "metamorphic core complex," despite its entrenchment in the geologic literature, because "complex" has been so widely used in so many geologic subdisciplines—sedimentary complex, ecological complex, metamorphic complex, and so on—it conveys little more than the notion of a group term for something that initially was baffling. "Complex," thus, is more a state of the mind than of the rocks. I shall offer a descriptive alternative term of "gneiss-cored dome." The use of "gneiss dome" is also avoided because of its application to features of metamorphic terranes, such as the Scandinavian shield area where both core and carapace are strongly metamorphosed and where a genetic connotation is made that may not be like that of the Cordilleran domes. The time to attribute more than a descriptive connotation to gneiss-cored domes should be held in abeyance until there is some agreement on their essential characteristics and until we have a better idea about their origin.

For this report, the gneiss-cored domes are inferred to be the product of a two-stage development. Often found in the older rocks between the domes are features of deep-seated compressive deformation, low- to intermediate-grade regional metamorphism, and voluminous plutonism. Such features are also found in the cores of the domes, formed mainly during an early stage. The features of the later stage of dome development are formed in a shallow tectonic environment, under tensional stress, and with a record of rapid young uplift, erosion and tectonic denudation, and thermal cooling. Some of these features also occur between the gneiss-cored domes, but more are present in them. While the intent is to focus on the features of the earlier stage of deformation, in some ranges a separation of features is difficult to make because of tectonic overprinting, a situation that has led to some diverse and even conflicting local interpretations.

Rocks deformed in the ductile style of the hinterland zone occur in southwestern Arizona and adjacent parts of northern Sonora and southeastern California, wherever Cretaceous and older rocks are exposed (fig. 23). The eastern boundary, between the hinterland and intermediate zones, trends north or northwest from near Lukeville, Arizona, swings east just south of Casa Grande, trends around a prong northwest of Benson, and

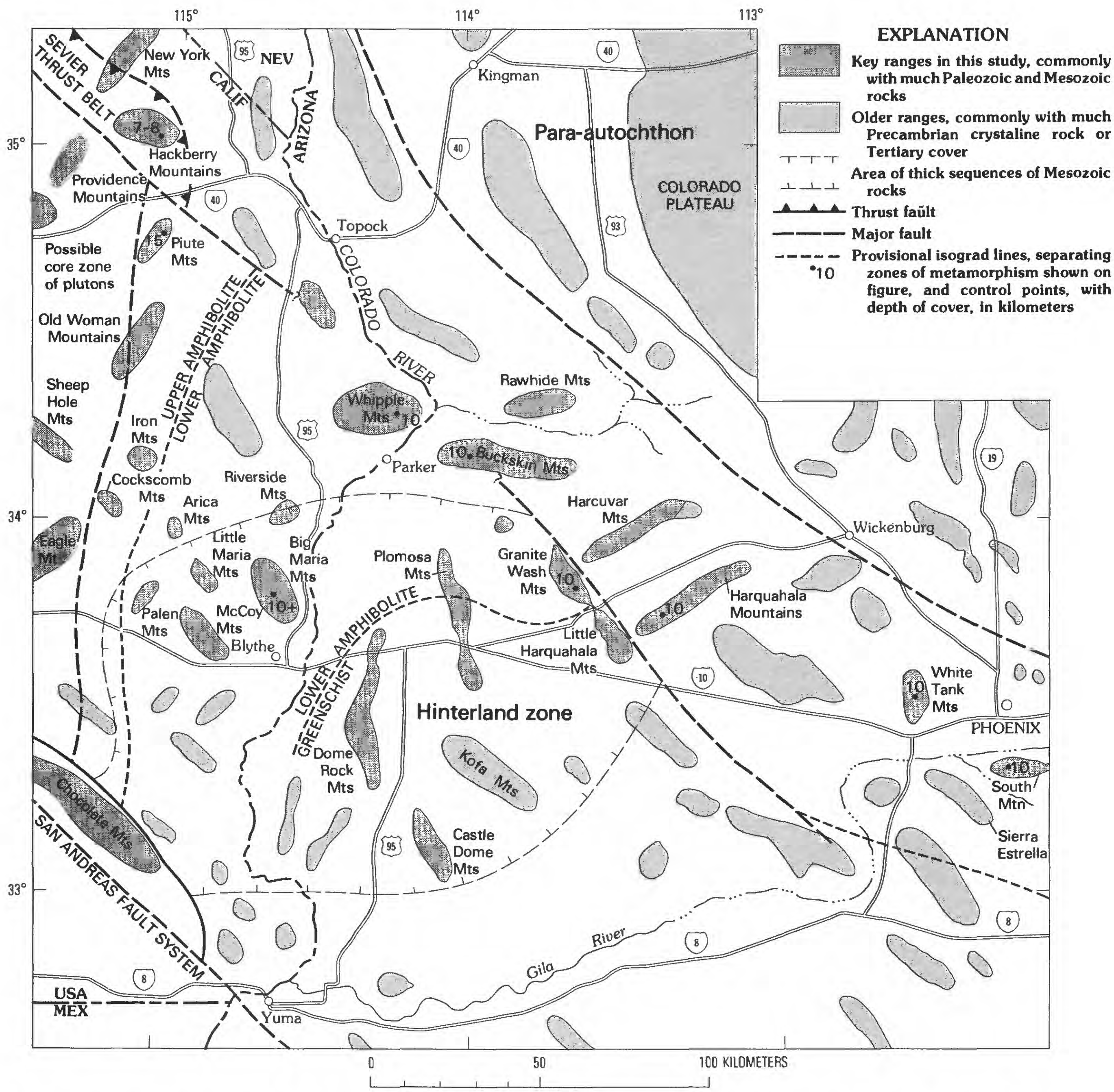


FIGURE 23. — Index map of much of the hinterland zone, showing key ranges providing data on Cordilleran tectonism and some other ranges, southwestern Arizona and southeastern California.

reaches the northeastern margin of the deformed belt at a point near Winkelman (fig. 19). The Pinaleno Mountains gneiss-cored dome near Safford may be an outlier of the ductily deformed hinterland terrane, or at least it has the late-stage tectonic overprint found in the gneiss-cored domes of the hinterland zone to the west.

This part of the hinterland zone southeast of Casa Grande is believed to abut the para-autochthonous or near-autochthonous foreland terrane of the Colorado Plateau. Features and enigmas of this margin of the Cordilleran orogenic belt are described in a separate following section.

In southeastern California, the San Andreas fault system appears to have truncated the entire orogenic belt, and bolson deposits conceal most of the region immediately southwest of that fault system.

To the northwest of Parker, Ariz., the contact between hinterland and intermediate zones seems to reappear, making a loop into southern Nevada. There, beyond the areal scope of this study, the contact separates the abundantly thrust-faulted but generally unmetamorphosed terrane of the Spring Mountains and Clark Mountains south and west of Las Vegas, from the intensely metamorphosed and deformed ranges to the south of them.

Core rocks in a gneiss-cored dome in the Rincon Mountains near Tucson (fig. 19) illustrate the hinterland style of deformation. This dome forms the southeastern of three such domes; the central one is in the Santa Catalina Mountains and the northwestern one in the Tortolita Mountains. Many interpretations have been made of features found in these mountains, but comprehensive detailed maps are not available for all of these ranges. The Rincons are probably the best known of the three and illustrate some key relations common to all. However, each dome has some unique features as well.

The core of the dome in the Rincon Mountains is underlain by crystalline rocks and its carapace is mainly sedimentary and metasedimentary rock with minor crystalline rock (Drewes, 1974, 1977; Thorman and others, 1981). The core and carapace rocks are separated by a major low-angle fault on which distinct Late Cretaceous (about 75 m.y.) and mid-Tertiary (about 25 m.y.) movements are recorded (fig. 24). A thick and varied suite of Oligocene sedimentary and volcanic deposits lapped onto the flanks of the dome, postdating the older compressional deformation but predating the younger extensional event. This suite of deposits is now only in low-angle fault contact with the core rocks and is either faulted or unconformable upon Paleozoic rocks of the carapace thrust plates. A few slivers of metasedimentary rock like that of the carapace are faulted into the core rocks on the southwest side of the dome. Unpublished deep seismic records across the saddle between the Rincon and Santa Catalina Mountains suggest that structural discontinuities occur within the crystalline terrane, as well as within the cover rocks. Lastly, a group of granitic stocks ranging in age from Paleocene(?) (previously considered to be Precambrian(?) by Drewes, 1974, 1977) to Oligocene intrude the core rocks and locally also the carapace. Evidence discussed in the section of this report on "Key Igneous Rocks of the Orogenic Belt" indicates that the emplacement of the older phase of the granitic stocks is associated with the last movement of the early deformation and perhaps also with the doming at depth.

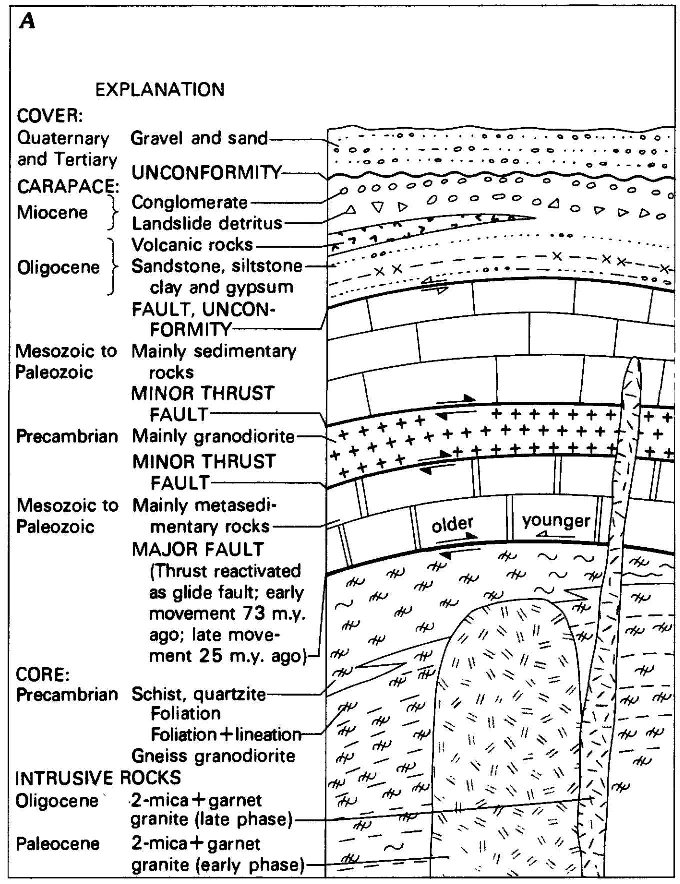
Rock types and tectonic features found in a gneiss-cored dome probably reflect conditions in two of the tectonic zones of the Cordilleran orogenic belt, as well as some post-orogenic features. The complex way that these features and conditions are intertwined both spatially in one gneiss-cored dome and temporally through tectonic overprinting is probably a major reason that the full significance of the older tectonic event has been difficult to discern. Tectonic features of the core rocks show most clearly the effects of an older, deep-seated, ductile style of compressional deformation characteristic of the hinterland zone. Tectonic features of the carapace rocks have stronger affinities with features of the western intermediate zone. Oligocene rocks are post-orogenic; yet they too have been involved in a late-stage, brittle-zone deformation of the dome. The emplacement of the group of 2-mica granitic stocks seems to span the period of development from late Cordilleran orogenic to the basin-and-range tectonic event, and the development of tectonic fabric in these stocks dwindles with the passage of time. There is also a record of a major reversal in local relief about 25 m.y. ago, and cooling of the core through various time-setting temperatures is protracted over a few tens of million years. The very complexity of these rock types and tectonic features suggests polytectonic origin.

Crystalline rocks of the core include Precambrian Pinal Schist and Continental Granodiorite, and a Late Cretaceous or early Tertiary (probably Paleocene) 2-mica granite, the Wrong Mountain Quartz Monzonite, shown diagrammatically (fig. 24) and in detail (Drewes, 1974, 1977; Thorman and others, 1981). (In an older rock classification system the rock is a quartz monzonite, but in a more current system it is a granite.) These core rocks are foliated, lineated, and typically cataclastically deformed.

Foliation is warped into a broad dome. On the southwest flank of the dome are gently southwest-plunging anticlines and synclines. The largest of these synclines forms the high saddle of Reddington Pass between the Rincon and Santa Catalina Mountains. Low on the southwest flank of the dome some of these folds are composite features of smaller amplitudes. Foliation intensity remains strong even at the deepest exposed levels of the core rocks.

Lineation appears in the plane of foliation mainly as elongate mineral pods enhanced by a stretching and rupturing of these pods through the cataclasis. Mostly, the lineation is oriented N. 60–70° E., but over an area of about 40 km^2 in the northeastern part of the Reddington Pass area the lineation swings gradually more northerly and ultimately to N. 10° W. (Thorman and others, 1981). Typically, the lineation carries across the crest of the dome undeflected in trend and undiminished in intensity, but it weakens on the northeast flank, and locally, south of the terrane with the N. 10° W. orientation, the lineation disappears. Lineation remains strongly developed at all levels of the core rocks.

FIGURE 24 (Right and following page).— Diagrammatic views of the gneiss-cored dome in the Rincon Mountains, Arizona. *A*, tectonic column, showing relationship among rock units; *B*, structure section, showing these relations in context of the dome. Distribution and density of foliation and lineation symbols reflect the intensity of their development in various units and across the dome.



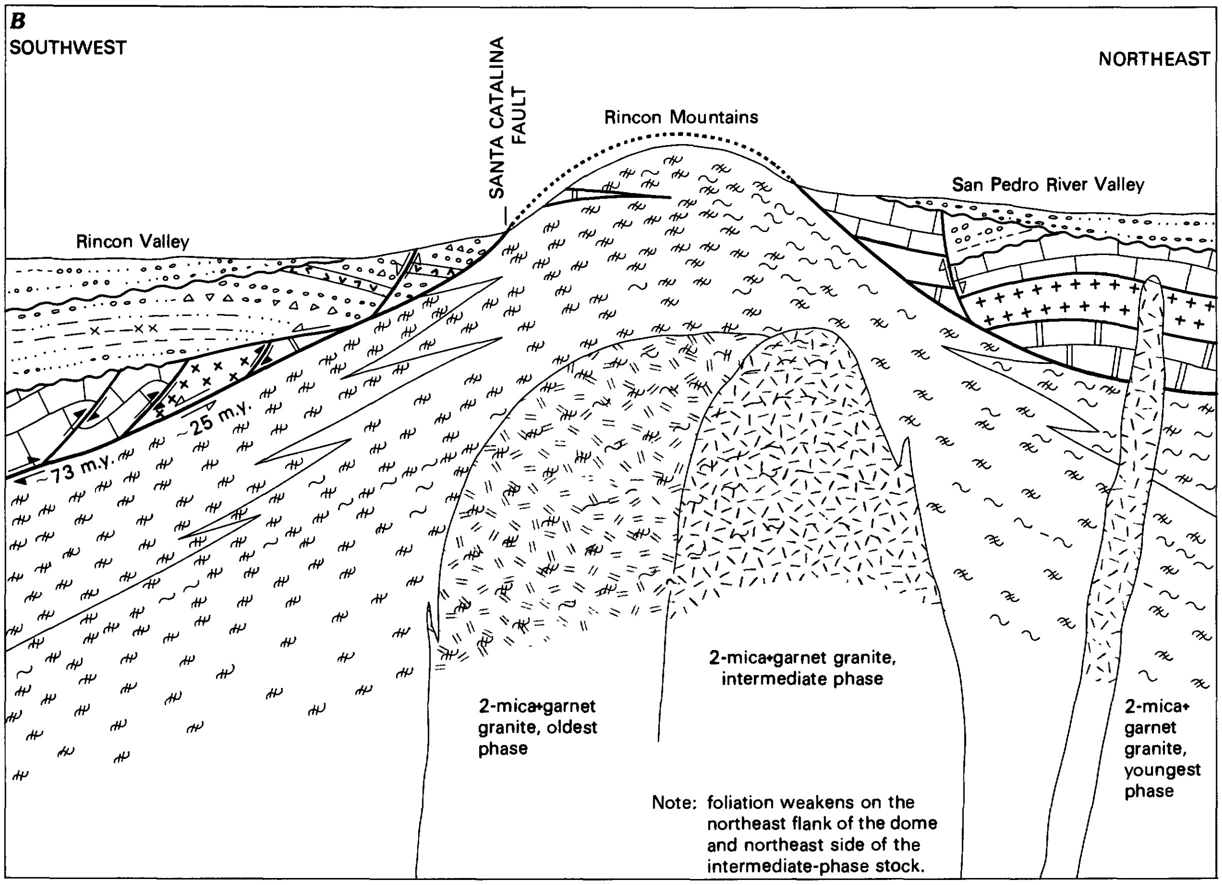
The cataclasis consisted of a brittle-style disruption or fragmentation of mineral pods and individual mineral grains. This feature is very strongly developed (generalized as the heavy dashed line in fig. 19) within a few tens of meters of the Santa Catalina fault and a distinctive meter-thick sheet of mylonite lies along much of that fault on the southwest flank of the dome. Cataclastic deformation diminishes a few hundred meters beneath this fault, but at several discrete horizons within the core it intensifies to mark lower level shear zones parallel to the foliation. Similar shear zones may bound tabular or lenticular bodies of schist and marble derived from the Pennsylvanian Horquilla(?) Limestone. Seismic reflectors at deep (unexposed) levels may mark similar intercalations and indicate that the core rocks are probably allochthonous, but movement involved ductile flowage rather than largely a lateral shift of semi-rigid thrust plates.

The 2-mica and garnet-bearing granite plutons are probably genetically related to some of the development of the gneiss-cored domes; yet details of this association remain enigmatic. The main stock of the Wrong Mountain Quartz Monzonite underlies much of the core of the dome; later phases, making smaller stocks lie progressively farther northeast. The older phases are cut by the last movement on the Santa Catalina fault separating core from carapace, but the younger phases penetrate the carapace. The main stock has a pervasive foliation and lineation, although weaker, like that of its host rocks. The next younger stocks have foliation developed on their west side only and have no lineation. The youngest of these stocks is neither foliated nor lineated.

The age of these 2-mica stocks probably ranges from Paleocene to Oligocene; complications of slow cooling or reheating affect the assessment of the history of the older stocks. The geologic relations of stocks to tectonic environments and to the thrust faulting of Cordilleran time is of greatest concern here. To this end the cross-range asymmetric distribution of fabric in the stocks is instructive.

The intensity of deformation varies across the Rincon Mountains gneiss-cored dome. Deformation along the Santa Catalina fault, in the overlying carapace, and also in the underlying core rocks is greater on the southwest flank than on the northeast flank of the dome. Such asymmetry is demonstrated by the development of subordinate anticlines and synclines only on the southwest flank, in the local lack of development of lineation to the northeast, and in the absence of foliation on the northeast sides of some younger of the 2-mica stocks. Along the Santa Catalina fault a mylonite zone is known only along the southwest flank. In the carapace the subordinate thrust plates are more continuous to the northeast and very lenticular and discontinuous to the southwest. Also, melange-like mixtures of Paleozoic and Mesozoic rocks within thrust plates occur only to the southwest.

This structural asymmetry across the Rincon Mountains is inferred to result from a stoss-and-lee side effect of thrust plates moving northeastward over an obstacle, such as a dome, much as a glacier may produce asymmetric features on knobs that they override. This inference is the basis for suggesting that the final movement on the thrust faulting along the Santa Catalina fault and genetically related faults in core and carapace was still in progress after the early or main mass of the Wrong Mountain magma reached the crustal level at which thrusting was taking place. Peak deformation is estimated to have been about 75 m.y. ago (Drewes, 1981a), so the final movement may have been



near the close of the Cretaceous or during the Paleocene, a time when 8–10 km of cover was present (see section on "Tectonic Environment" in this report). The main body of the 2-mica granite reached this level of 8–10 km probably during the Paleocene also and developed a dome in partly ductile material, at depth only. Interference effects of the final subhorizontal northeastward-moving mass over the rising partly cooled mass generated the stoss-and-lee side features. The regional tectonic stress dwindled, and the cover over the dome was gradually removed to bring the rocks to the level of brittle deformation. Although faulting is not recorded during Eocene time, regional stress that caused the faulting had not fully ceased in Eocene time, for some middle-stage Eocene(?) stocks have a protracted cooling history and weakly developed foliation on their stoss (southwest) sides (Thorman and others, 1981). Finally, with the emplacement of the latest stage magma during the Oligocene, the host rocks lay wholly within the brittle level, and so the dome was raised to the surface, there to form a topographic as well as structural feature. That late development is recorded in the sedimentary record of the youngest of the carapace rocks and initiates the late Tertiary gravity sliding and rapid deep erosion of the Rincon Mountains (Drewes, 1977, 1981a; Drewes, LeMone, and others, 1982, p. 94–96).

Key observations and inferences bearing on the hinterland style of deformation during the Cordilleran orogeny are (1) a terrane of more ductily deformed rocks underlies one of more brittlely deformed rocks; (2) ductile deformation took place under much greater cover than brittle deformation; (3) the ductile deformation is the older event; (4) while most of the ductily deformed rocks are of the crystalline basement or the early stage 2-mica granite, there has been some tectonic interleaving of the basement and the cover rocks; (5) both the events of tectonic deformation and of stock emplacement and cooling were of long duration; and (6) while the earlier ductile deformation was mostly unidirectional, top to the northeast, locally aberrant flowage direction developed for reasons as yet unknown. Many of these features and relations occur in other ranges of southwestern Arizona, but in few places do all occur together, and in other places some additional key observations have been made.

Several mountain ranges in the Phoenix area (fig. 23) offer data on the ductily deformed hinterland terrane. East and north of Phoenix is a belt of Tertiary rocks covering crystalline terrane. Where Mesozoic and Paleozoic sedimentary rocks crop out northeast of this cover, as they do 50 km east of Phoenix, they are only mildly folded and locally slightly thrust faulted. Closer to, and north of Phoenix, Tertiary volcanic and sedimentary rocks in a sequence 1–2 km thick form blocks listrically faulted upon ductily deformed crystalline basement rocks.

In the South Mountains, south of Phoenix, basement rocks are warped in a dome elongate N. 70° E. parallel to the lineation (Reynolds, 1985). A 2-mica garnet-bearing stock has Oligocene and Miocene (28–22 m.y.) ages, compatible with the late phase stocks of the Rincon Mountains. Both host and intrusive rock are foliated and lineated, and part of the intrusive rock is also cataclastically deformed with fabric indicative of northeast transport. While Reynolds placed the tectonic development of these rocks in the mid-Tertiary extension event, I suspect, through analogy with the gneiss-cored domes southeast and northwest of South Mountains, that an older chapter of development is obscured in this range; consequently, these obscured features could be attributed to the ductile tectonic elements of the older age, which were subsequently overprinted by the mid-Tertiary brittle-zoned features. The continuum of ductile to brittle deformation that Reynolds (1983) inferred, may reflect deformation first in a deep-seated and then a shallower tectonic environment, with a gradual downward decrease of the overprinted cataclasis.

The White Tank Mountains, west of Phoenix, are also underlain mainly by Precambrian gneiss and a Tertiary 2-mica garnet-bearing granite (Brittingham, 1985). These rocks contain a southwest-dipping foliation, a northeast-southwest-oriented lineation, and zones of cataclasis mostly tens of centimeters thick and tens of meters apart parallel to the foliation. The stock is dated as 34 m.y. (Rb-Sr isochron). Brittingham inferred this to be a minimum emplacement age of a granite, which may have had an emplacement age of Cretaceous or older. The cataclastic event, marked by southwest transport, is overprinted on an earlier northeast-directed deformational event. The age relation between the intrusive event and the cataclastic event is uncertain. By analogy with estimates of muscovite stability and adamellite solidus in another gneiss-cored dome (Brittingham, 1985; Davis and others, 1980), the older more ductile deformation features, such as *sc* (from the French words for cleavage (*s*cission) and schist (*s*ciste)) fabric, are believed to have formed beneath a cover about 9.6 km thick. Brittingham (1985) advised caution in interpreting ductile features like *sc* fabric, because, locally, he identified diverse directions of movement.

Assessed together, two ranges near Phoenix indicate that beneath the extensive cover of mid-Tertiary to Quaternary age strata is a crystalline terrane that has a complex structural fabric. This fabric was developed largely in a ductile style at a structural level 9–10 km beneath the surface during an earlier time (Cordilleran orogeny?). Some of the fabric elements, however, were formed under brittle (that is, shallower) conditions at a later time. Development of brittle features may be associated with listric normal faulting of the tilted blocks of Tertiary volcanic and sedimentary rocks and thus are extensional features. In some places both events seem to have had codirectional tectonic transport (as on the northeast flank of the Rincon Mountains); elsewhere, transport was in diametrically opposed directions (as on the southwest flank of the Rincons). Firm dating of the early (Cordilleran) event is lacking near Phoenix.

Many of the ranges of western Arizona and adjacent California are known only from reconnaissance studies, and those which have had detailed work were studied more for their mid-Tertiary extensional deformation than for their older history. However, in such key ranges as the Harquahala, Little Harquahala, Granite Wash, Whipple, Piute, Old Woman, Plomosa, Dome Rock, McCoy, and Big Maria Mountains (fig. 23), new data having a direct bearing regarding the hinterland zone has been derived in the past few years. In general, these ranges are underlain by a crystalline basement and intercalated plates of metasedimentary and metavolcanic rock of Paleozoic and Mesozoic age. These rocks are intruded by granitic stocks mainly of Cretaceous age, but some are intruded by stocks of Jurassic or early Tertiary age. Metamorphic grade is at least in the greenschist facies; some localities are in the lower, or even upper, amphibolite facies. Deformation is almost wholly of a ductile style, except of course for the ubiquitous brittle deformation of mid-Tertiary or late Tertiary age overprinted on the features of the Cordilleran event.

The Harquahala Mountains are one of several anomalously northeast-trending ranges 100–150 km west-northwest of Phoenix (fig. 23). They are underlain by an extensive core of igneous and metamorphic rocks and a cover of Tertiary volcanic rocks. This core and cover are separated by a major low-angle fault known as the Bullard detachment fault (Varga, 1977; Reynolds, 1980, 1982; Keith and others, 1982; Reynolds and Spencer, 1985; and others). These rocks form a northeast-trending antiform with a gently plunging northeastern nose. In many ways the Bullard fault resembles the Santa

Catalina fault of the Rincon Mountains, having been reactivated and recording diverse movement, although some early studies along both faults led to inferences of one-stage movement.

Core rocks of the elongate Harquahala dome provide information as to the style of deformation during the Cordilleran orogeny. Thick plates of Precambrian gneissic granite are intercalated with thin plates of attenuated Paleozoic metasedimentary rocks. Paleozoic rock within the thin plates are folded into a few large northeast-trending and southeastward-inclined recumbent structures and into many smaller northwest-trending structures (Keith, in Drewes, Keith, and others, 1982, p. 80–82). These rocks are intruded by a 2-mica granite of Late Cretaceous or early Tertiary age. Most rocks are foliated and lineated parallel to the antiform axis, N. 70° E., and in places are also cataclastically deformed. Post-orogenic dikes are of Miocene (25 m.y.) age. Of special note here is the evidence, in two sets of folds, of two phases of Cordilleran deformation involving an older southeast-directed and a younger northeast-directed event.

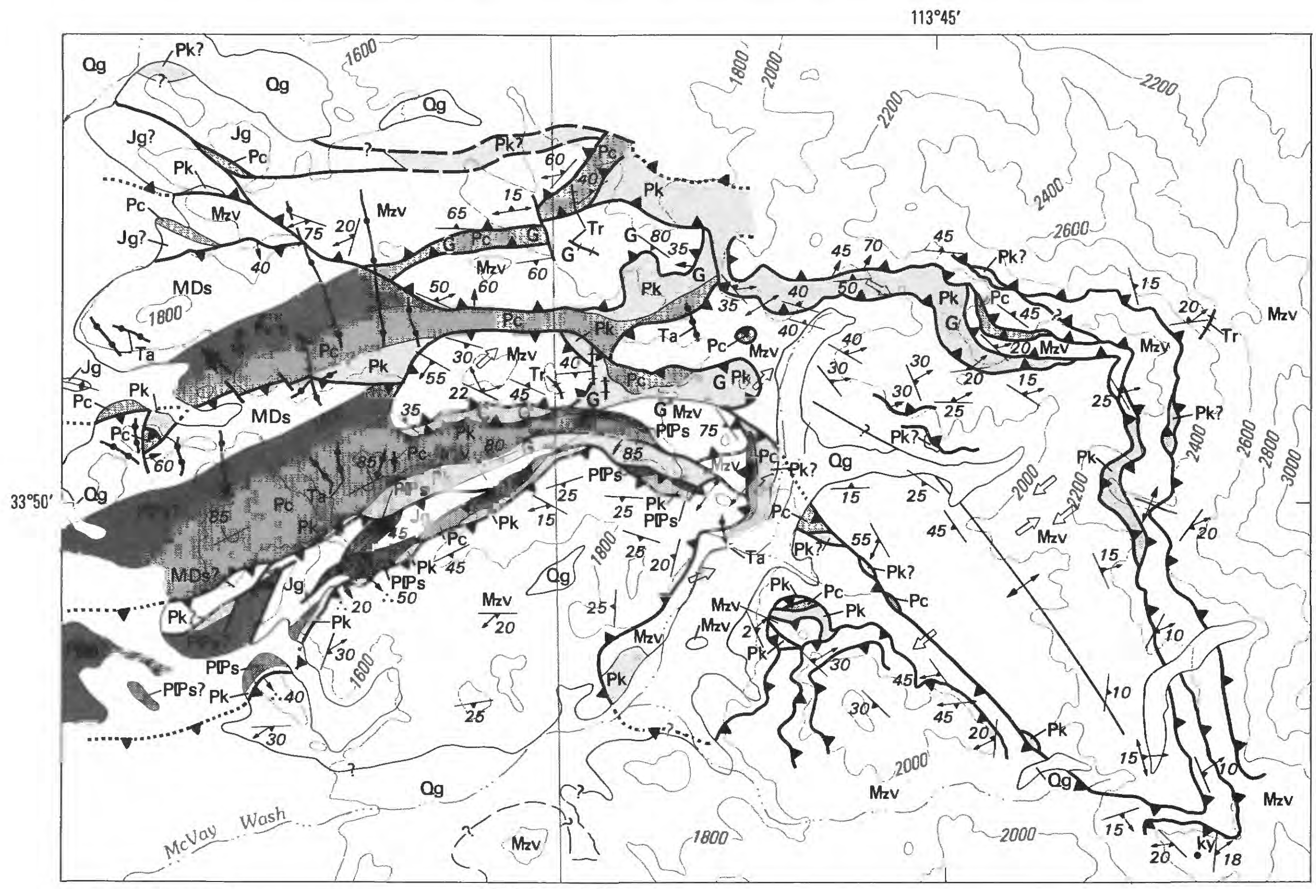
The Little Harquahala Mountains (fig. 23) are underlain by rocks similar to those of the Big Harquahalas, but Paleozoic units are less attenuated in the former mountains (Richard, 1982). This is the easternmost range in which the upper Paleozoic rocks are correlated with the Grand Canyon sequence rather than with the southeast Arizona sequence. They are overlain by a 1,650-m-thick sequence of Jurassic and Cretaceous rhyolite to andesite volcanic rocks and sedimentary rocks that are correlated with the McCoy Mountains Formation. The rocks of the Little Harquahalas are thrust faulted and warped in an older set of southwest-inclined, large-amplitude recumbent folds and a younger set of north- or northeast-inclined small folds. Cleavage is common in some of these sequences and formations, foliation is widespread, but lineation is absent. All together, these features of the Little Harquahala Mountains seem to represent a slightly higher structural level than those of the Harquahala Mountains, and all predate the emplacement of an undeformed Late Cretaceous (69–65 m.y.) granite stock in the pass separating the Little Harquahala Mountains from the Granite Wash Mountains to the northwest.

The Granite Wash Mountains (fig. 23) are underlain by strongly metamorphosed metaigneous and metasedimentary rocks of Mesozoic age and intercalated thrust slivers of Permian rocks of the Grand Canyon sequence. These rocks and structures are illustrated by a map of the McVay Canyon part of the mountains (fig. 25). These thrust-faulted, attenuated, foliated rocks form an asymmetric northwest-trending and northeast-inclined open anticline. A mineral lineation is well developed in the gneissic metaigneous rocks and typically trends northeast-southwest. From study of the *sc* fabric, transport direction is about equally common to the northeast and southwest; either polyphase deformation is recorded or, as indicated by Brittingham (1985) for the White Tank Mountains, the use of *sc* fabric indicators is not always reliable. Kyanite-bearing rocks in a structurally high plate at the south end of the area demonstrate amphibolite metamorphic grade, indicative of 8–10 km of cover.

Some notable evidence for diverse directions of tectonic transport occurs at the north end of the Granite Wash Mountains, about 4 km north of the area of figure 25. There, feldspathic gneiss that hosts a post-orogenic (Paleocene) granodiorite stock is well exposed on a ridge some 200 m high (secs. 1 and 2, T. 6 N., R. 15 W., and sec. 7, T. 6 N., R. 14 W., Utting 15-minute quadrangle). The gneiss has well-developed foliation, lineation, *sc* fabric, and, near the crest of the ridge, small intercalated slivers of metasedimentary rocks like those in the McVay Wash area. Using the *sc* fabric as a transport direction indicator, Steven Reynolds (personal commun., 1984) inferred that tectonic transport at the base of the ridge was upper mass to the southeast and at mid-slope upper mass to the northeast. Near the crest and a bit to the north of the place of these observations, the *sc* fabric indicates a northwest tectonic transport. At all sites the small folds—amplitudes mostly of a few centimeters—have northwest-trending axes implying transport and compression or shear oriented northeast-southwest.

The initial observations provided a basis for inferring two directions of transport. Similar inferences had been reached from other kinds of evidence in nearby ranges (Reynolds, 1980, 1982; Keith and others, 1982; Richard, 1982). However, with evidence of a third direction of transport other interpretations may be needed. Perhaps transport direction cannot be reliably measured from *sc* fabric, or perhaps transport in a tectonic environment of great ductility involves laminar flow in diverse directions in response to a single regional stress but various local pressure gradients, or perhaps the tectonic history is, indeed, three-phased in this region. For the present, this quandary remains unresolved; in any case, different kinds of evidence at many sites demonstrate at least two phases of movement during the Cordilleran orogeny of this region.

Thin slivers of Paleozoic rocks that alternate with either Mesozoic gneiss (McVay area) or Precambrian gneiss (Harquahala area) resemble the intercalations in various tectonic zones in several ranges far to the east. For instance, the slices of limestone or marble at mid-flank on the southwest side of the Rincon Mountains resemble those in the McVay area. Likewise, the slices of Precambrian granite among the Paleozoic rocks of the



Base from U.S. Geological Survey, Utting and Salome, 1:62,500, 1962

R 15 W R 14 W

0 .5 1 2 KILOMETERS

0 1/2 1 MILE

Geology by Harald Drewes, 1985, with contributions by S.J. Reynolds, written commun., 1987

EXPLANATION

- Qg — **Gravel (Quaternary)**—Alluvium
- Tr — **Rhyolite (Tertiary)**—Dike
- Ta — **Andesite (Tertiary)**—Dike
- Mzv — **Metavolcanic rock (Mesozoic)**—Includes sedimentary units
- Jg — **Granite (Jurassic?)**—Coarse-grained, metamorphosed
- Pk — **Kaibab Limestone (Lower Permian)**—Metamorphosed, bluish-gray, cherty. Where queried may include older Paleozoic rocks
- Pc — **Coconino Sandstone (Lower Permian)**—Sandstone and quartzite
- **Supai Formation (Lower Permian and Pennsylvanian)**—Red beds, phyllite
- MDs — **Metasedimentary rocks (Mississippian and Devonian)**—Limestone
- **Contact**

- 85 ∴ 50 **Fault**—Showing direction and amount of dip. Dip value following 3 dots obtained by the 3-point method. Dashed where projected beyond mapped area
- 40 **Thrust fault**—Showing direction and amount of dip. Sawteeth on upper plate
- **Anticline**—Showing trace of axial plane
- 80 **Strike and dip of beds**
- 85 **Strike and dip of overturned beds**
- 15 **Strike and dip of foliation**
- **Strike of lineation**
- 15 **Foliation and lineation**—Showing dip of foliation and direction of plunge of lineation
- ⇨ **Direction of tectonic transport**—Using sc fabric or vergence of small folds
- G **Site of gypsum**—Occurring as intrusive pod
- •ky **Site of kyanite**—Indicative of amphibolite grade metamorphism

FIGURE 25.—Geologic map of part of the McVay Wash area in the Granite Wash Mountains, western Arizona, showing thrust faulted and metamorphosed Paleozoic and Mesozoic rocks.

Big Hatchet Mountains or among Mesozoic rocks of the Chiricahua Mountains are similar (see details in last two sections). These intercalations may have developed during the stage of development of an intermediate zone. The stronger metamorphism and expressions of great ductility of the western sites, were then overprinted features. This composite aspect of interior tectonic zones is developed more fully in the last section of this report.

In the Buckskin Mountains (fig. 23), core rocks of another gneiss-cored dome display amphibolite-grade metamorphism (Rehrig and Reynolds, 1980). Two lineations are in the lower plate but only one occurs on the detachment fault beneath the extensionally faulted upper plate. Reactivation of movement on this fault, likely associated with a major change in tectonic environment is inferred; a hiatus intervened between amphibolite-grade metamorphism and extensional deformation of unmetamorphosed Tertiary rocks of the carapace of the dome.

In the Whipple Mountains of California (fig. 23) another gneiss-cored dome has been thoroughly studied by Dickey and others (1980), Davis and others (1982), and Thurn (1982). A mid-Cretaceous older deformation is recorded by a 94.6-m.y.-old deformed dacitic dike swarm and a late kinematic 80–91-m.y-old granite stock. The dike swarm may be feeders to a late orogenic cover, now eroded. A younger deformation is dated at 23–13 m.y. old; this younger deformation resulted in reactivated movement on a detachment fault separating carapace from core rocks. This late movement followed the development of some mylonite and formed a brittle cataclasis. Only one direction of tectonic transport, east-northeast, is recognized; either both episodes of movement were directed east-northeast, or the later movement obscured evidence of an older direction of movement.

In the Big Maria and Little Maria Mountains (fig. 23) Paleozoic and Mesozoic formations are extremely attenuated. These rocks are cut by low-angle faults variously identified as thrust faults, or as detachment faults (used by some workers in a narrow sense as meaning tension-generated, rather than in the broader sense of equivalent to a décollement fault of either compressional or extensional origin) (Baltz, 1982; Emerson, 1982; Hamilton, 1982; and Ellis, 1982). Tectonic transport to the northeast, southwest, or south-southwest are proposed by various workers either for various areas or, in some cases for the same areas, with some conflicts between northeast movement versus southwest movement remaining unresolved. Likewise, workers differ on whether there was only a single, extensional, mid or late Tertiary event or two, compressional and then extensional, mid or late Mesozoic and then mid or late Tertiary events. In any case, amphibolite-grade metamorphism is recorded on a lower plate, which suggests that some of this deformation took place beneath about 10 km of cover.

In the Dome Rock and McCoy Mountains (fig. 23) and other nearby ranges, are two thick sequences of sedimentary and volcanic rock whose past distribution may have been widespread (which would account for at least some of the thick cover that has been surmised to have existed during the Cordilleran orogeny). The McCoy Mountains Formation is commonly 6 km thick and in one place is even reported as 7.3 km thick (Harding, 1982; Emerson, 1982). The McCoy is variously assigned a Late Cretaceous or Paleocene age, or a Late Jurassic age. The Palen Formation was originally assigned a Triassic to Jurassic age by Pelka (1973) and was so viewed by LaVeque (1982). Subsequently, Harding and Coney (1985) correlated the Palen Formation with the lower part of the McCoy Mountains Formation, which they viewed as Jurassic, rather than as Cretaceous and Paleocene, as proposed by Pelka (1973) and used by others. The Paleocene age of part of the McCoy is based on fossil wood whose initial appearance in Paleocene time is still under debate. From this background I view the Palen as being Jurassic to Early Cretaceous and the McCoy as being Jurassic to at least Late Cretaceous. The Palen Formation is about as thick as the McCoy Mountains Formation (LaVeque, 1982) and it includes rhyodacite, and it was deformed about 90 to 100 m.y. ago, near the end of Early Cretaceous time. Some or all of the Palen Formation may, therefore, be correlative with the lower part of the McCoy Mountains Formation. These formations may be tectonogenetic analogs of volcanic and clastic sequences of the intermediate and fold-and-thrust zones, such as the Fort Crittenden and Salero Formations, the Ringbone Formation and Hidalgo Volcanics, and the Love Ranch Formation.

Studies in these ranges of southwestern Arizona and southeastern California have generated the greatest diversity of interpretation of the number and age of tectonic events and of transport direction and environment of deformation. Such a situation is to be expected in that part of the hinterland zone of an orogenic belt close to the core zone. Complications may reflect a combination of a westward deepening of the tectonic environment and a greater duration of application of compressional stress. Such an increase in complications is not as easily accounted for by a simple one-event extension of mid or late Tertiary age.

A string of ranges, in position the third or fourth west of the Colorado River, provides indications that a core zone is nearby. In the Piute Mountains (fig. 23) high

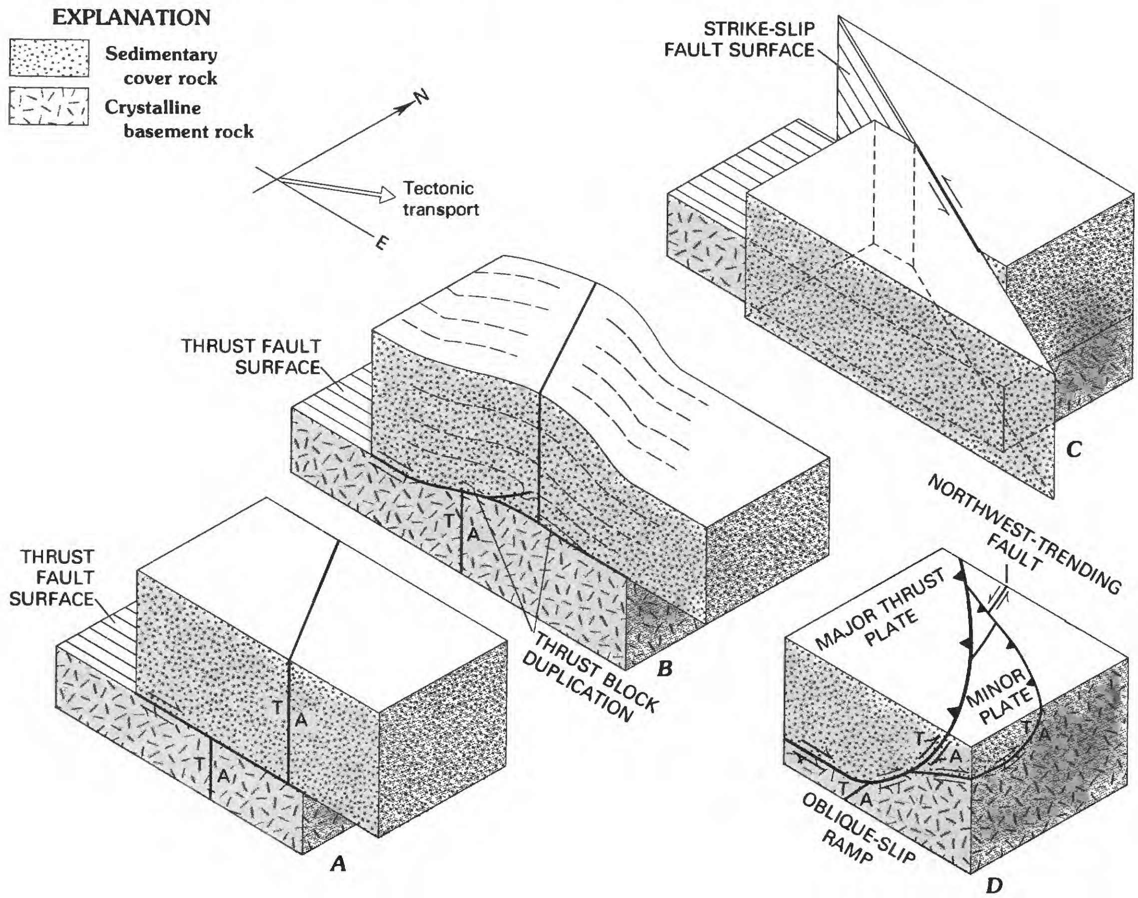


FIGURE 26.—Block diagrams showing three idealized basic relationships between older strike-slip faults in basement and cover rocks and a décollement thrust fault between them; *A*, simple offset; *B*, ramp; and *C*, reactivation of segment of the strike-slip fault. *D*, offers a more realistic *B* and *C* combination. Along faults: T, movement towards viewer; A, movement away from viewer.

amphibolite-grade metamorphism implies an original cover of about 15 km (Miller and others, 1982). Farther south, in the Iron Mountains and Sheep Hole Mountains, an Early Cretaceous batholith may be part of the core zone. Foliation, northeast-southwest-oriented lineation, and cataclastic deformation are older than a 60- to 61-m.y.-old aplite (Howard and others, 1982). Still farther south in the Coxcomb and Eagle Mountains, is another batholith, of a Late or Middle Jurassic age (152.8–167 m.y.). Mid or late Tertiary extensional faulting is not recorded in this string of ranges and in one instance a pull-apart edge of such a terrane is proposed, remarkably close to the site of the 15 km cover. Clearly, these events reflect vastly different tectonic environments from one another and therefore different times.

Finally, the ranges nearest the northeast side of the Coachella–Imperial Valley and the San Andreas fault system introduce a new tectonic enigma not necessarily related to the systematic change across the Cordilleran orogenic belt described thus far. In the Oricopia and Chocolate Mountains (fig. 23) an assemblage of Mesozoic or older oceanic rocks, including graywacke pelite, basalt flows, diorite, chert, and limestone, were thrust northeast over Precambrian granite and gneiss in Late Cretaceous and Paleocene time (Haxel and Dillon, 1978). Assessment of this terrane will not be made herein.

KEY FEATURES AND CONDITIONS

In the foregoing descriptive sections covering sequentially the several structural zones and areas adjacent to the Texas-California transect of the Cordilleran orogenic belt, some key features and conditions were recurrently mentioned. These included orogenic sedimentary and igneous rocks, orientation of stress or direction of tectonic transport, the age and amount of that movement, and the deformational environment or cover beneath which deformation occurred. Also mentioned at several localities was the enigma of the truncation or ending of the tectonic zones to the north. A separate review of the enigma and then of the key features and conditions will indicate the systematically changing characteristics of this belt of deformed rocks and will lend considerable support to my inference of its unity. Finally, these topical reviews set the stage for an attempt to portray the development of this segment of the orogenic belt.

STRUCTURAL FEATURES OF THE NORTHEASTERN MARGIN

In the foregoing description of tectonic zones of the Cordilleran belt along the Texas-California transect, the zones were shown to end northward against the margin of the belt. Oblique convergence of the mobile belt with the southwest side of the semi-cratonic Colorado Plateau block (fig. 3) generated a band of complex strike-slip and oblique-slip structures and may also have caused that side of the block to be raised and thus deeply eroded. This section describes structural features on the northeast margin and interprets their origin.

The irregularities and peculiarities of the northeastern margin of the strongly deformed terranes of southern Arizona and New Mexico have created as much uncertainty about the validity of the hypothesis of regional thrust faulting in this region as any other feature. Much of the margin is covered by Tertiary volcanic rocks or by younger basin deposits. Also, much of the margin is eroded to deep levels where deformation styles are highly ductile; and the origin of such deformation has been variously interpreted. Finally, only a few seismic records of this terrane are available for systematic study. Consequently, the following interpretation remains provisional.

The characteristics of the northeastern margin of this part of the Cordilleran orogenic belt are believed to reflect various responses or adjustments to oblique structural barriers in the subsurface, to various levels of exposure and variable thickness of the mobile terrane, and to a change from more brittle style to more ductile style of deformation.

Three simplistic types of responses of a décollement along the interface between cover and basement rocks are shown in figure 26 (parts *A* to *C*). A more realistic diagram, *D*, illustrates the combined effects of *B* and *C*, on or along a barrier whose bounding strike-slip fault trends oblique to the tectonic transport direction and dips moderately southwest.

The characteristics of the margin are illustrated in a series of block diagrams (fig. 27) that utilize the principles of figure 26 from El Paso to a point northeast of Benson, Ariz. (fig. 27, block 7), where ductily deformed terrane first crops out. The separate blocks of this figure do not represent an exploded block diagram; there are gaps in continuity between adjacent blocks. Also, the position on their respective axes, shown on the index map, is one of cartographic convenience. Thus, contacts and map units do not necessarily match from block to block. Furthermore, late movements have been added to some peak-deformation features, and so some diverse offsets and movement symbols are shown. For each block, those data are shown that best illustrate the local relations along that part of the margin of the mobile belt.

The northwest-trending faults are reactivated basement flaws of Precambrian heritage that occur in a broad belt from El Paso at least to Las Vegas. Various of these faults were reactivated before, during, or after the Cordilleran orogeny. Segments of some of these faults vertically offset the interface between basement and cover rocks. Where such offset raised the northeast

EXPLANATION

(Broken lines are projected. Queries indicate a more uncertain part of a line or feature. Side surfaces of block diagrams follow convention changes from map (top) view, as used in structure sections; A, movement away from viewer; T, movement toward viewer)

Mz	**Mesozoic rocks**
Pz	**Paleozoic rocks**
p€	**Precambrian rocks**
p€+	**Precambrian plus other rocks**

Contact

Line showing possible limit of thrust plate before erosion

Major fault
Minor fault } **The following symbols are used only in map (top) views**

Thrust fault—Sawteeth mark upper plate

Strike-slip fault—Arrow couple shows relative movement

Folds:

Anticline

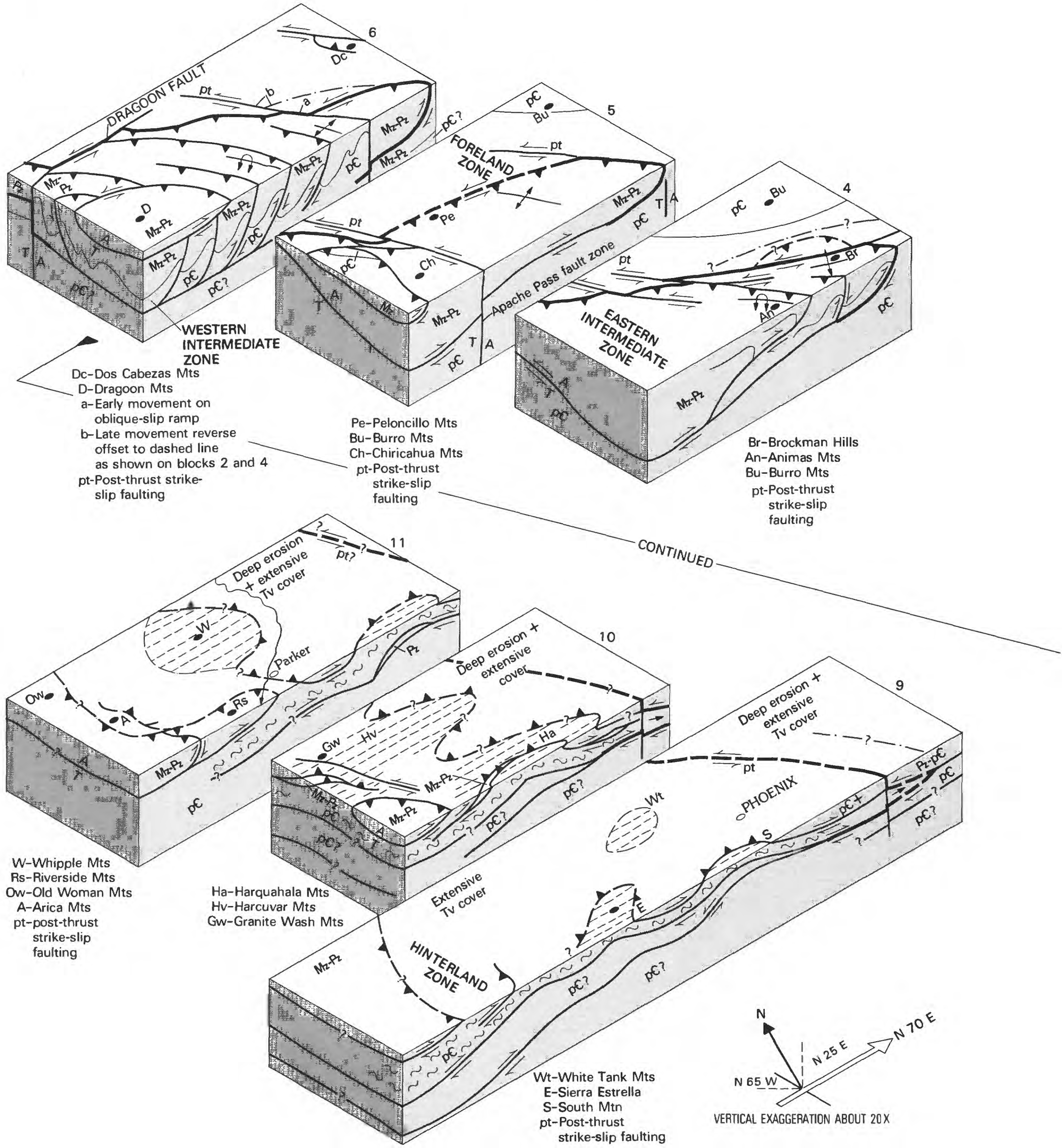
Overturned anticline

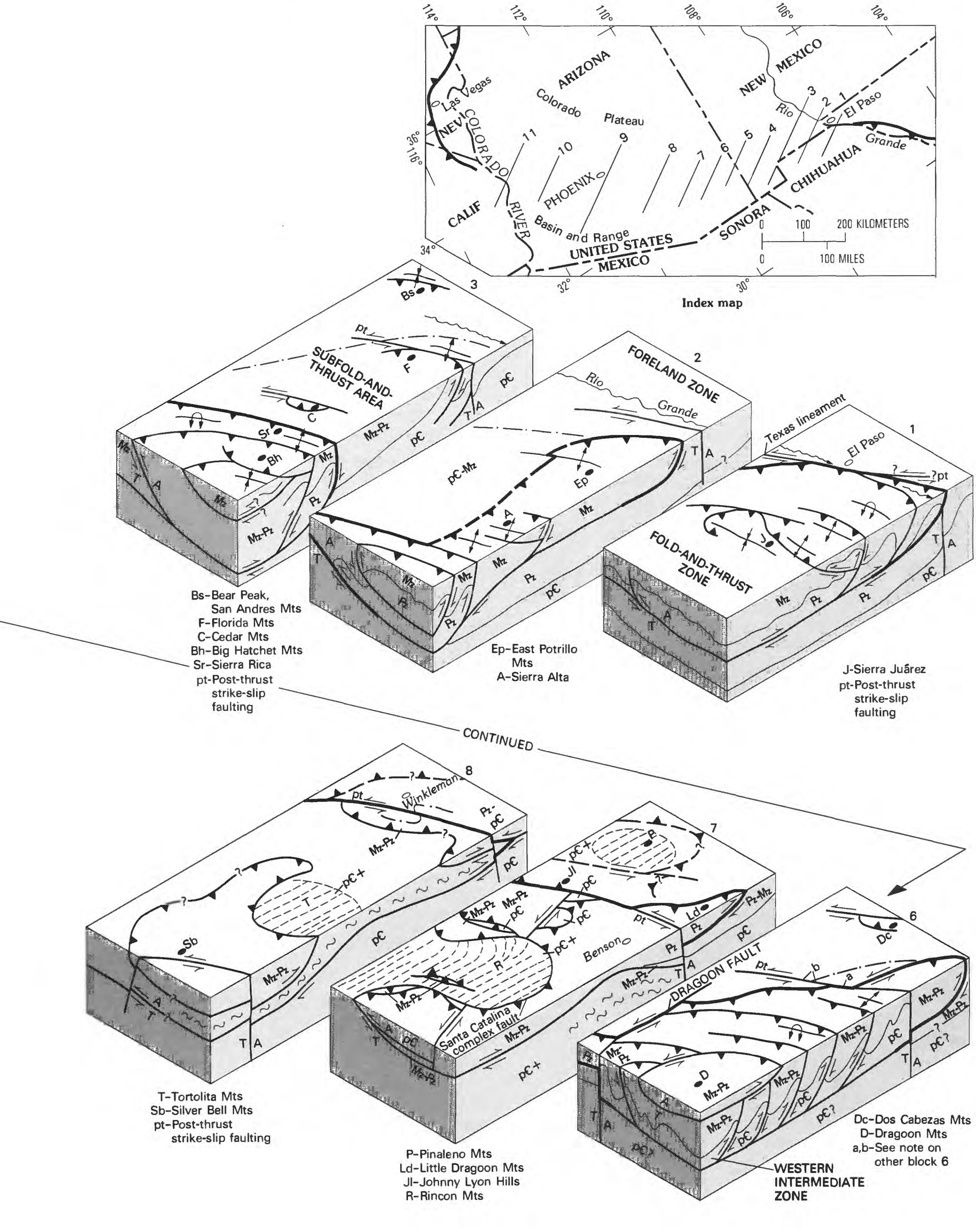
Syncline

Foliation, shown in side views (sections)

Lineation, shown only in top views (maps)

FIGURE 27 (Explanation above with figure on following two pages).—Block diagram series (not an exploded block diagram) showing key tectonic features along the northeast border of the mobile belt (in heavy lines) Cordilleran orogen segment between El Paso, Texas and Las Vegas, Nevada. The several movement phases of the Cordilleran orogeny are shown in places by separate diverse movement symbols. Mid-Tertiary and younger block and detachment faulting omitted; erosion level approximates present conditions. In profile, foliation is projected only a short distance from this terrane.


6
5
4
DRAGOON FAULT
FORELAND ZONE
Apache Pass fault zone
WESTERN INTERMEDIATE ZONE
EASTERN INTERMEDIATE ZONE
Dc-Dos Cabezas Mts
D-Dragoon Mts
a-Early movement on oblique-slip ramp
b-Late movement reverse offset to dashed line as shown on blocks 2 and 4
pt-Post-thrust strike-slip faulting
Pe-Peloncillo Mts
Bu-Burro Mts
Ch-Chiricahua Mts
pt-Post-thrust strike-slip faulting
Br-Brockman Hills
An-Animas Mts
Bu-Burro Mts
pt-Post-thrust strike-slip faulting
CONTINUED
11
10
9
Deep erosion + extensive Tv cover
Deep erosion + extensive cover
Deep erosion + extensive Tv cover
Parker
PHOENIX
Extensive Tv cover
HINTERLAND ZONE
W-Whipple Mts
Rs-Riverside Mts
Ow-Old Woman Mts
A-Arica Mts
pt-post-thrust strike-slip faulting
Ha-Harquahala Mts
Hv-Harcuvar Mts
Gw-Granite Wash Mts
Wt-White Tank Mts
E-Sierra Estrella
S-South Mtn
pt-Post-thrust strike-slip faulting
N
N 25 E
N 70 E
N 65 W
VERTICAL EXAGGERATION ABOUT 20X


Index map
ARIZONA
NEW MEXICO
NEV
CALIF
COLORADO RIVER
Colorado Plateau
Basin and Range
PHOENIX
Las Vegas
El Paso
Rio Grande
CHIHUAHUA
SONORA
UNITED STATES
MEXICO
0 100 200 KILOMETERS
0 100 MILES
SUBFOLD-AND-THRUST AREA
FORELAND ZONE
FOLD-AND-THRUST ZONE
Texas lineament
Bs-Bear Peak, San Andres Mts
F-Florida Mts
C-Cedar Mts
Bh-Big Hatchet Mts
Sr-Sierra Rica
pt-Post-thrust strike-slip faulting
Ep-East Potrillo Mts
A-Sierra Alta
J-Sierra Juârez
pt-Post-thrust strike-slip faulting
CONTINUED
Winkleman
Benson
DRAGOON FAULT
Santa Catalina complex fault
T-Tortolita Mts
Sb-Silver Bell Mts
pt-Post-thrust strike-slip faulting
P-Pinaleno Mts
Ld-Little Dragoon Mts
Jl-Johnny Lyon Hills
R-Rincon Mts
Dc-Dos Cabezas Mts
D-Dragoon Mts
a,b-See note on other block 6
WESTERN INTERMEDIATE ZONE

blocks, barriers to décollement movement along the cover-basement interface were created that generated some complexities. Near El Paso (fig. 27, block 1), the margin of the fold-and-thrust zone against the foreland zone (that is, the frontal line) may be a combination of strike-slip fault and thrust fault, perhaps in part ramping against or upon the northeast block of the strike-slip structure known as the Texas lineament (or the Clint fault), much as Uphoff (1978) and Lovejoy (1980) envisioned it.

Northwest of El Paso (fig. 27, block 2) the margin swings west and perhaps even southwest along a trace first eroded to a sub-thrust plate level and then concealed by younger deposits. The original extent of continuous plate system is shown on blocks 2–4 as a dot-and-dash line. The foreland zone lies east and north of the orogenic belt and merges with the subfold-and-thrust area, where the frontal plate is eroded away.

In southwestern New Mexico, where plate thickness probably increases to an extent that it was not entirely eroded away, the margin is again a thrust fault or strike-slip fault zone (fig. 27, blocks 3 and 4). The thrust fault or faults in the Sierra Rica probably are parts of this structure, with those wholly within the Cretaceous sequence, as viewed by Zeller (1960), representing only minor branch faults.

Still farther west, in the Brockman Hills (fig. 27, block 4), a trace of the margin is again exposed as a fault between Paleozoic rocks and underlying Late Cretaceous or early Tertiary andesite (Thorman, 1977a). With the combination of inferred strike-slip east-trending fault and east-trending folds in and near the hills (Corbitt and Woodward, 1970) this fault probably is an oblique-slip (left-slip and reverse) fault.

At scattered sites across southwestern New Mexico thrust faults interact in various ways with the northwest-trending fault system. In some places the nature of this interaction can only be inferred. In the Peloncillo and Chiricahua Mountains, however, this interaction is shown (fig. 27, block 5), much simplified from several studies (Drewes and Thorman, 1980a, 1980b; and Drewes, 1982a), from exposed structural relations. In the Chiricahua Mountains particularly, the interaction probably is that of an early phase oblique-slip(?) ramp structure and of a late-phase left-slip (Drewes, 1981b; Drewes and others, 1988).

In the Dragoon Mountains (fig. 27, block 6) a northeast-trending left-lateral strike-slip fault, the Dragoon fault, (that is, one parallel to tectonic transport direction of the thrust plates) offsets the frontal line or northeast margin. Only splay faults off this strike-slip fault are seen (Drewes, 1987), and the main thrust fault beneath the plates, the Hidalgo thrust fault of Drewes (1980), is present, with some subordinate lower strands, in the Little Dragoon Mountains and the southern part of the Johnny Lyon Hills (Cooper and Silver, 1964; Drewes, 1980), near Benson (fig. 27, block 7).

The northeast margin of the orogenic belt has some different characteristics northwest of Benson than east of Benson. Most obviously, the trend of the margin is mainly northwest, similar to the inferred jog in trend in the Sierra Rica–Sierra Alta area. Two sets of thrust plates exist; the one set is part of a northwestern orogenic lobe, and the other is part of a southeastern orogenic lobe. Some movement between these lobes was independent of each other in rate or time. A break between a northwestern and southeastern set occurs on both the northeast and southwest flanks of the Rincon Mountains, affecting only the carapace rocks (Drewes, 1980, 1981a). Accordingly, I have proposed that the break is a left-slip fault, which, like the fault north of the Dragoon Mountains indicates the direction of tectonic transport during the main phase (in this area) of the Cordilleran orogeny (fig. 27, blocks 6 and 7).

Later movement then occurred on a northwest-trending strike-slip fault, as recorded in the structural complications of several ranges. In the first complication, along the western foothills of the northern Johnny Lyon Hills, the little-deformed nearly autochthonous or para-autochthonous northeastern terrane that has a Precambrian crystalline basement is underlain by a plate (or thrust wedge) of Paleozoic and Mesozoic sedimentary rocks, apparently jammed into or beneath some of the little-deformed rocks. This relation is shown in block 7 (figure 27) by the segment of older northwest-trending fault with sawteeth on the northeast side and, to the south with a younger reactivated strike-slip movement symbol (but the actual wedge of rock is not shown at the scale of this block).

In another situation, farther northwest, near Winkelman (fig. 27, block 8), an oroclinal flexure in upended Late Proterozoic, Paleozoic, and Mesozoic rocks (Krieger, 1974a, 1974b) suggests major left-slip drag on a concealed northwest-trending fault.

A third complication involves the significance and distribution of metamorphic fabric among the little-studied Precambrian crystalline terrane of the areas of blocks 7 to 11 (fig. 27), particularly that of blocks 8 and 9. Some fabric elements are probably formed early (Cordilleran, compression); others are late features (Miocene, extension). Still others probably first formed under compressive stress and were then modified in an extensional system; with the movement reversal occurring after a hiatus, the thickness of cover was much reduced. Such a two-step movement history is favored in the Rincon Mountains (fig. 27, block 7) and other gneiss-cored domes. Evidence for inferring that a metamorphic basement moved out from beneath the western part of the Colorado Plateau during regional extension is discussed by Lucchita (1982). Possibly, the Pinaleno Mountains of southeastern Arizona, with its deformed basement rock (Thorman, 1981) and its position among ranges having para-autochthonous or foreland zone cover rocks, represents a window through the kind of terrane Lucchita has in mind for western Arizona.

Other complications involving strike-slip faults probably also mark the frontal line of the orogenic belt of the northwestern lobe. Left-slip faulting (fig. 27, blocks 7 and 8) is inferred to have produced the large oroclinal flexure of Paleozoic and Mesozoic rocks mapped southwest of Winkelman by Krieger (1974a, 1974b). The flexure in lineation on the north side of the Rincon Mountains also suggests a left drag along a concealed northwest-trending strike-slip fault. In western Arizona, anomalously northeast-trending ranges like the Harquahala and Harcuvar Mountains (fig. 27, block 10) seem to be abruptly terminated against a concealed northwest-trending high-angle fault that may be a major strike-slip structure. If this is the case, however, the juxtaposed rocks have similar physical properties, for geophysical evidence offers no support of a major fault at this location.

In summary then, I view the northeast margin of orogenic belt between El Paso and Las Vegas to be chiefly a left-slip adjustment around a salient mass of the southwestern part of the Colorado Plateau block. Complications developed for at least four reasons: (1) The eastern half of the margin is an oblique collision in more brittle rocks; whereas, the western half involves more ductile rocks. (2) Deep erosion along the margin stripped off some of the overriding plates, resulting in an irregular trace, and later deposition buried the frontal line. (3) At some localities offsets along the older northwest-trending faults produced barriers to plate movement that were variously overcome as ramps (fig. 11), right-oblique-slip ramps, or oblique-slip reverse faults. Truncated barriers may even have been the source of some masses of Precambrian crystalline rocks in plates dominantly of Paleozoic or Mesozoic rocks of the intermediate tectonic zones. Finally (4), late left-slip movements took place along rejuvenated segments of the northwest-trending faults.

KEY SEDIMENTARY FEATURES OF THE OROGENIC BELT

The orogenic development of the Texas-California transect is closely tied to the sedimentation history, albeit that history is still poorly known. A review of two aspects of the sedimentation record will be undertaken here: that of the thickness of the cover rocks present during orogenic time and that of provenance of some key deposits. Cover thickness is a major factor controlling the style of deformation. Provenance of deposits in major basins of deposition reflect the orogenic or anorogenic conditions of the region; as an orogeny develops systematically, so do the related depositional features.

The rocks of interest in this context, summarized in table 1, range in age from about Late(?) Jurassic to early Eocene and in lithology from shale to volcaniclastic conglomerate.

The ages given these units are not all well established, and further work may show some of the western units to be correlatives of one another. Some of the given ages are only estimates obtained through bracketing dates on associated igneous rocks, a few are based on but a single-dated horizon, then extrapolated for a thick sequence of rocks. One or two dates are my own estimates, based on a best fit of a mixture of vague or conflicting estimates in the literature. Despite some uncertainties, these sequences are systematically younger to the east, a pattern also reflected in the age of deformation, reviewed in the second following section. Probably, separate depocenters were involved.

The configuration and kind of basins of deposition in which these sediments accumulated (fig. 28) is better known in the fold-and-thrust zone than the hinterland zone because the geologic record is better preserved in the distal parts of orogens. The Love Ranch Formation is viewed (Seager, 1981) as having formed in several intermontane basins wholly subaerially. These basins trended askew to the present basin-and-range basins of south-central New Mexico and probably were not much different from the largest basins of mid-Tertiary age of the same area. At least one of these ancient basins, of the southern San Andres Mountains, lay close to the foot of a concurrently rising range off which much of the detritus of the basin was shed (Love Ranch Formation).

The Ringbone and Fort Crittenden Formations and their correlatives were deposited in a basin or pair of basins of larger size than the Love Ranch troughs. These deposits typically coarsen upward and include some tuffaceous material and volcaniclastic conglomerates. Whereas the upward coarsening of the volcanigenic deposits, from shale and siltstone to conglomerate, could reflect an increase in relief of their provenance, it could just as well indicate the eastward spread of andesitic volcanism, thereby reflecting a normal progression of a tectonic-magmatic process. When definitive sediment transport studies of these rocks are made, I predict that many of them will be found to have come from the west or northwest. With at least the Fort Crittenden Formation and the Cabullona Formation of northeastern Sonora (Taliaferro, 1933), the basins contained large lakes which left a record of a rich faunal assemblage of a late Late Cretaceous age.

TABLE 1.—*Age and location of major sedimentary units associated with the Cordilleran orogenic events along the Texas-California transect*

Name, lithology	Age	Location	Thickness (km)
Love Ranch Formation: clastic rocks ...	Paleocene–Eocene	South-central N. Mex.	0.6+
Unnamed marine beds	Late Cretaceous, Paleocene, and Eocene	South-central N. Mex.	1.0
Assorted units of dacite and andesite, mostly unnamed	Paleocene–Eocene	South-central N. Mex.	---
Ringbone Formation: conglomerate......	latest Cretaceous–Paleocene	SW. N. Mex.	5.8+
Hidalgo Volcanics: mostly andesite	latest Cretaceous–Paleocene	SW. N. Mex.	1.6+
Fort Crittenden Formation and correlative informal units: clastic rocks..	Late Cretaceous	SE. Ariz., Sonora, Mexico	6.9+
Salero Formation and correlative mostly informal units: mainly andesite and rhyodacite	Late Cretaceous	SE. Ariz., Sonora, Mexico	1.5
Bisbee Group: shale, siltstone, and sandstone	late Early Cretaceous	N. Mex., Ariz., Chihuahua and Sonora, Mexico	3.1+
Assorted informal and unnamed units: mainly graywacke, pelite, conglomerate...	mainly Early(?) Cretaceous; in part may be correlative with Bisbee Group	South-central Ariz.	4.0+
Livingston Hills Formation: sandstone and conglomerate..........................	Cretaceous or Tertiary	W. Ariz.	3.6
Unnamed andesite	Cretaceous	W. Ariz.	1
McCoy Mountains Formation: sandstone and conglomerate..........................	Late Jurassic–Early Cretaceous	W. Ariz., SE. Calif.	7.3
Palen Formation: sandstone and conglomerate	Jurassic–Early Cretaceous[1](mainly Late Jurassic)	SE. Calif.	7

[1]See discussion in text on this age assignment, see page 49. In the original work on the Palen Formation, Pelka (1973) assigned the unit a Triassic or Jurassic age. Subsequently Harding and Coney (1985) correlated the Palen Formation with the lower part of the McCoy Formation, which is Jurassic and Cretaceous. With the Cretaceous evidence being weaker than the Jurrassic evidence, accordingly, I view the Palen as Jurassic to Early Cretaceous.

The Bisbee Group was deposited in a still larger basin shown by Hayes (1970) to have had a northwest-trending axis that plunged gently southeast. The northwestern part of the basin was subaerial—estuarine and fluvial—and the southeastern part was marine, as indicated by the distribution of reefal limestones. Most of the Bisbee Group is gray shale and siltstone not particularly rich in organic material; feldspathic sandstone is interbedded with these gray or, less commonly, greenish-gray or reddish-gray beds, and a thin limestone pebble and cobble conglomerate underlies the group. Locally, however, the conglomerate is very coarse, thick, and rich in clasts of Precambrian rocks and of Paleozoic limestone. These varied and thick deposits reflect different local conditions within the overall basin. For instance, the feldspathic sand of the New Mexico part of the basin is inferred to have come from the Burro uplift to the north, a positive area extending from the site of the present Florida Mountains across the site of the Burro Mountains to possibly the Pinaleno Mountains (fig. 5). It is one of the few areas along the southern part of the Colorado Plateau from which the Paleozoic limestone cover had been eroded. Local high areas are inferred to have supplied the detritus for several thick conglomerate units in southeastern Arizona (Ransome, 1904; Sabins, 1957b; Finnell, 1970, 1971; Hayes, 1970; and others). At a few sites in southeastern Arizona the basal conglomerate of the Bisbee Group is strongly time transgressive, reaching the stratigraphic level of the medial, or limestone unit; at one site, the northern Dragoon Mountains, these coarse deposits include many lens of landslide detritus.

A third source is likely for the voluminous gray mud deposited in the Bisbee basin. A southeastern source is unlikely because the sea lay there. A northeastern and northern source is equally unlikely because the limestones capping the Colorado Plateau do not provide such detritus. However, the western terranes, toward the core of the orogenic belt and thus capped by andesitic volcanic rocks, would be an ideal source of gray mud. Therefore, I suggest the bulk of the Bisbee Group was derived from the west: sediment was likely transported east into the broad fluvial basin along the orogenic front and then southeast along the axis of the basin to the Bisbee marine embayment.

The western and older basin or basins are so thoroughly eroded that little evidence of their configuration is left. They were clearly very deep basins, for preserved sequences are 4–7 km thick; preserved deposits are typically unconformably covered by post-orogenic rocks, so that these thicknesses are minima. The McCoy Mountains Formation is inferred to have been deposited in a Jurassic transtensional rift basin (Harding and Coney, 1985), but evidence of faulting of the basin margins concurrent with sedimentation is difficult to demonstrate. Also, age constraints on these thick deposits are inadequate to rule out an Early Cretaceous (or even Late Jurassic) component of this unit; arguments for a still earlier age, however, seem to be invalid, as they pointed out. Inferred directions of local transport can be attributed to other kinds of basins and to other considerations of erosional history of adjacent areas than those suggested by Harding and Coney. Better identification and dating of metasedimentary deposits in the hinterland zone is likely to show that these very thick deposits were more widespread than a relatively local feature such as an aulocogen or rift basin. This formation, and the others of the western terranes, also contain much volcanogenic detritus whose source is not readily attributed to erosion of underlying rocks. Instead, a source from an andesitic terrane, such as an arc or orogenic belt, to the west where batholithic roots of Late Jurassic to Early Cretaceous age are now exposed, seems most plausible, although by analogy with the other basins that were tilted south or southeast, southerly components of transport along a basin axis may be expected.

Several aspects of the late pre-orogenic and orogenic sedimentation, previously described, are noteworthy, despite the fact that systematic studies of these rocks has barely begun: (1) Depositional basins were large—both areally and in thickness. (2) They were elongate features with northwest-trending axes. (3) At least two or three of them were tilted south or southeast, reflecting a

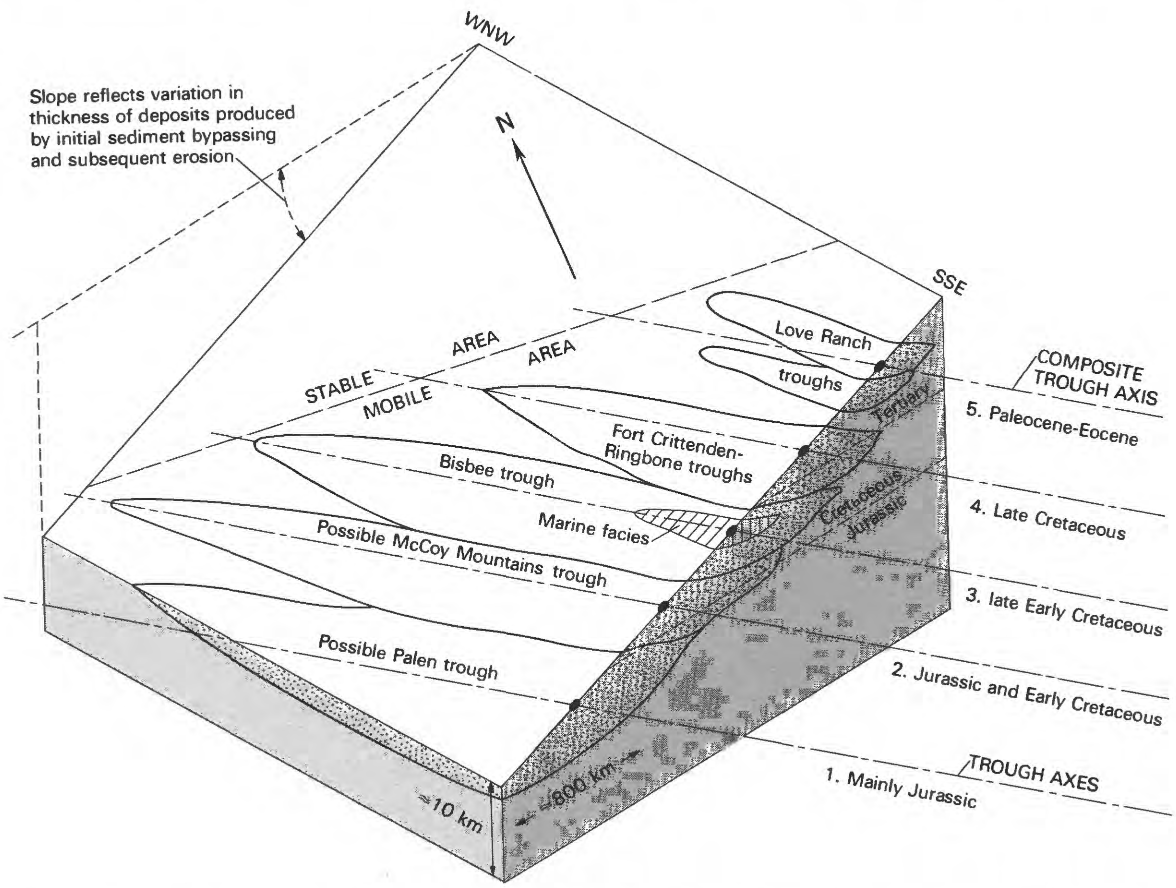


FIGURE 28.—Block diagram showing inferred shift of traces of depocenter axes across the southern segment of the Cordilleran orogen, Jurassic, 1, through Paleocene and Eocene time, 5. Identity of individual basins with specific formations may represent an artificial "freezing" of a continuum of depositional process. The southeasterly tilt of depocenter axes is supportable only in stages 3 and 4. Since the time this diagram was prepared, a deep exploratory well about 20 miles northeast of the East Potrillo Mountains is reported (Sam Thompson, personal commun., 1988) to have penetrated 3,000 feet of marine Maastrichtian, Paleocene, and Eocene beds, thereby providing another link in this series of depocenters.

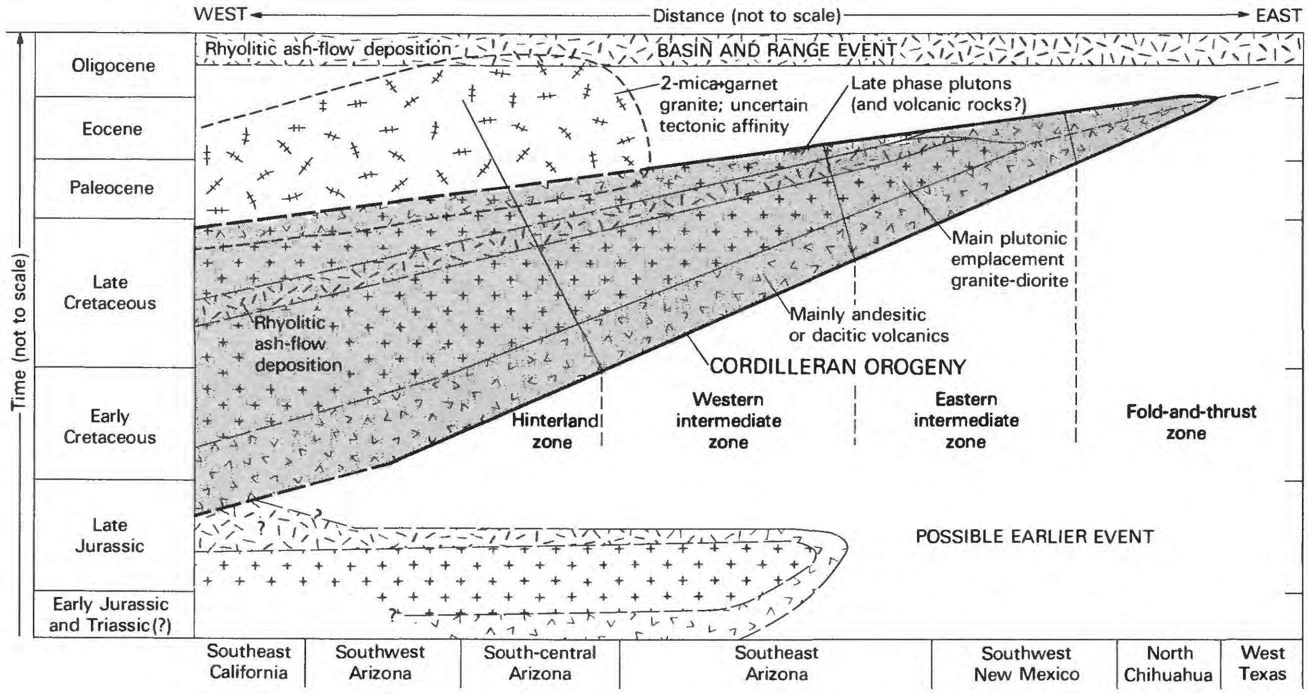


FIGURE 29.—Generalized time-space diagram along the Texas-California transect, showing the distribution of igneous rocks. Heavy line around Cordilleran orogeny. Long dashed line where age controls sparce. Heavy dotted line shows field of 2-mica-garnet granite, having affinities to both Cordilleran and basin-and-range events.

persisting (or eastward propagating?) upland north of the mobile belt. (4) All but the smallest, easternmost, and youngest basins contain detritus derived from several sources. (5) The major component of these deposits is volcanogenic. (6) Provenance of this volcanogenic component is probably mainly to the west, although transport may have had a local southerly vector along basin axes. (7) The several basins seem to form an eastward-younging series of foreland basins, nourished primarily from older parts of the orogenic belt, but secondarily from other uplifts. (8) Ultimately, all basins were also caught up in the deformation.

KEY IGNEOUS ROCKS OF THE OROGENIC BELT

Igneous rocks of the Cordilleran orogenic belt are abundant along much of the Texas-California transect of the belt. This Texas-California transect, which trends west-northwest oblique to the El Paso–Las Vegas segment, provides some important information about the development of the belt. Unfortunately, the igneous rocks have not yet been studied in detail systematically across the belt. Indeed, most sites where the igneous rocks have been most carefully studied are near mining districts where complexities and alteration of the rocks are strongest; thus, the igneous rocks are difficult to synthesize. Nevertheless, one suite of igneous rocks is unequivocally tied to the orogenic development; two other suites appear to be older or younger than the first suite and have only nebulous ties to the orogeny.

The unequivocally tied orogenic suite of igneous rocks runs the gamut from plutonic to hypabyssal to volcanic, and ranges in composition from silicic to intermediate—granite and rhyolite to diorite and andesite. The suite typically postdates the deposition of the late pre-orogenic foreland basin deposits and predates the early Tertiary unconformity beneath post-orogenic deposits. To the west the age range of this orogenic igneous suite is Late Jurassic to about early Paleocene; to the east it is entirely early(?) Eocene, as shown inside the westward tilted wedge with the heavy border of figure 29. The distribution in time and space along the Texas-California transect of the two equivocal suites above and below this wedge, respectively, delineate the Late Cretaceous to Oligocene 2-mica granite and the older Jurassic granite.

The orogenic volcanic rocks seem to be a bimodal suite, commonly of thick, widespread, and strongly propylitized andesite and dacite, overlain by more localized, but locally also thick, rhyolite or rhyodacite. In some places peak deformation occurred between the times of effusion of these volcanic end members; in other places such refinement in geologic dating is not yet available. Andesitic rocks include mainly lava flows and flow breccias; the rhyolitic rocks are mainly ash-flow deposits but include some air-fall and lava-flow units. Andesite and rhyolite are not now coextensive, but they may have been more nearly so initially because the rhyolite tuffs are usually the youngest orogenic deposit and thus were the first to be eroded in late orogenic and post-orogenic time.

Andesitic volcanic rocks probably belonging to this orogenic suite occur at least as far east as the Victorio Mountains (Thorman and Drewes, 1980) and are voluminous in the Hidalgo Volcanics of southwest New Mexico, in the eastern intermediate zone. Farther west, in Arizona, the same lithotectonic unit occurs, mostly in small erosional remnants but in places as an extensive sheet, such as the andesite of Hunt Canyon of Epis (1956), the lower member of the Salero Formation, and the Silver Bell Formation. In south-central Arizona equivalents of these formations are mapped as metavolcanic rocks, meta-andesite, greenschist, and greenstone. In western Arizona and southeastern California equivalent rocks are schistose or gneissic, with only overall color, phenocryst type, and some remnants of texture providing evidence of their intermediate composition.

Rhyolitic volcanic rocks of orogenic affinity are known as far east as the Chiricahua Mountains (Drewes, 1981b) and Pedregosa Mountains (Drewes and Brooks, 1988), where thin lenses or sheets of welded tuff occur in the top of, or above the andesite flows (shown in fig. 29 by a unit of mixed symbols). The upper member of the Salero Formation and the Cat Mountain Rhyolite are examples of such rhyolite. Thick andesite successions commonly contain some rhyolite, suggesting a possible common set of volcanic source centers; a genetic tie is reinforced in places by the presence of coeval intrusive rocks of various compositions. Several sites of rhyolite-welded tuff or of the andesite and tuff are also sites of coarse block breccias, variously referred to as "chaos" or exotic blocks (Brown, 1939; Simons and others, 1966; Drewes, 1971a). Some of these breccias may be formed around cauldron structures (Drewes, 1980, 1985b; Drewes and others, 1988; Lipman and Sawyer, 1985) of Late Cretaceous age. Distinctly younger rhyolitic volcanic rocks are not known in the western terranes, although units designated as "metavolcanics" and rhyodacite may include such rocks. Had they been present as extensively as they were farther east, they would subsequently have been strongly altered (as the andesites were) and then eroded.

Intrusive rocks of orogenic or early post-orogenic age are distributed even more widely than the andesite and rhyolite. In the fold-and-thrust zone and nearby part of the foreland zone small hypabyssal plugs and dikes of andesite, such as the early post-orogenic Campus Andesite at El Paso and the dike swarm in the Sierra Juárez (Lovejoy, 1980; Drewes and Dyer, in press) are dated as early Eocene (47–48 m.y.). In the eastern intermediate zone intrusives include small plutons of quartz monzonite and diorite of Paleocene age (where directly dated) or Late Cretaceous and Paleocene age (where only indirectly dated). All are associated with base metals and silver enrichment, chiefly in veins, pipes, and stockwork deposits.

In the western intermediate zone granitoid intrusive bodies are more numerous, larger, and more commonly made up of several composition phases, ranging from granite to diorite. Associated mineral deposits are larger, and deposit types are more varied; they include most of the porphyry-type deposits. Most stocks are dated as latest Cretaceous; a few late-phase bodies are early Paleocene. Hypabyssal rocks persist in both of the intermediate zones but are less voluminous than the stocks. Loci of some large stocks and associated volcanic fields are along northwest-trending faults (Drewes, 1981a, fig. 2). Contact metamorphic zones, which were of little consequence farther east, are much wider here and, in the western part of the region, tend to merge into a semiregional low-grade metamorphism.

Intrusive rocks of orogenic affinity of the hinterland zone retain some characteristics of those of the intermediate zones. Those stocks have a normal granitoid texture and are of (early?) Late Cretaceous age in Arizona, but some in southeastern California are of Early Cretaceous age and a few are of Late Jurassic age. Other rocks of late orogenic or post-orogenic age associated with gneiss-cored domes are also present.

Peraluminous 2-mica garnet-bearing granite stocks are typically associated with the gneiss-cored domes. The early phases of these stocks are among the latest magmas of orogenic age, seeming to extend from a Late Cretaceous body in the Chemuhevi Mountains of California (John, 1982) to Paleocene bodies in southeastern Arizona. Some petrographic and petrologic accounts of these stocks are given by Rehrig and Reynolds (1980) and by Davis and others (1982).

The age of the 2-mica granite stocks and their relationship to tectonic events has attracted much study and diverse inferences. Results of the first radiogenic ages obtained ranged from Precambrian(?) (my error on the Wrong Mountain Quartz Monzonite) to Miocene (probably only the youngest phase or reset ages). Many of these stocks seem now to be Paleocene, with late phases in the Eocene or Oligocene, and with some to the far west of Late Cretaceous age. Many of these stocks and their host rocks have had a protracted cooling history, in some cases accentuated by a record of final cooling through the apatite and zircon (fission-track dating method) isotherms, 20 m.y. or more after emplacement. Many 2-mica stocks are foliated, lineated, and cataclastically deformed, suggesting that the times of emplacement and of latest Cordilleran tectonic movement overlapped and ductile conditions prevailed. Other late-phase 2-mica stocks, of the Rincon Mountains and the Little Dragoon Mountains, for example, escaped the tectonic overprint, apparently because movement had ceased in those areas by the time of emplacement. The gradual decrease in extent and intensity of development of tectonic fabric in the 2-mica stocks, described in the Rincon Mountains (Drewes, 1974, 1977; Thorman and others, 1981), may be related to the gradual loss of ductility, which ended with the relatively abrupt uplift of the dome, to the surface, during the Oligocene.

While the early part of the 2-mica granite development is believed to be of Cordilleran orogenic affinity, the late part extends into the following period of extensional deformation. This situation is one bit of evidence that there may be a genetic tie between these events.

One other enigma remains, concerning the Jurassic stocks and volcanic rocks of southern Arizona. Across southern Arizona and into southeastern California they mostly are of one age, between 145 and 165 m.y. In southeastern Arizona these rocks predate the foreland basin deposits, but to the west the foreland basin deposits get older and in southeastern California the magmatic and depositional events may overlap in time. Consequently, the Late Jurassic and Early Cretaceous stocks of southeastern California may be genetically tied either to the Jurassic magmatic event or to the Cordilleran orogeny. Or, is it possible that the Jurassic magmatic event may be genetically tied to the Cordilleran orogeny?

In summary, the magmatic record of the Cordilleran orogenic belt is systematically linked to the orogenetic development, as well as to the sedimentary development. Stocks and volcanic rocks invaded and covered the deformed terranes and, as deformation propagated across the orogenic belt, magmatism followed. Toward the frontal line, where tectonic stresses dwindled, so did the volume and variety of igneous rocks. The andesitic volcanic cover over the deforming (or just deformed) terrane not only added the final load beneath which thrust faulting took place but also provided the detritus for the foreland basins that developed concurrently to the east. In turn, as these basins were overwhelmed by compressional stress, deformed, and covered by andesitic volcanic rocks, they were recycled to basins being formed a bit later and still farther east.

AGE OF DEFORMATION

Along the Texas–California transect of the Cordilleran orogenic belt, as along others, the concept of "age of deformation" has increased markedly in complexity as geologists have acquired data from more sites, topics of study, and generations of workers. At the outset, I shall point out some past geologic practices in dating and their limitations, thereby leading to a rationale for some improvement.

In general, an age of deformation that is fairly precisely determined in some accessible locality (through the standard means of dating a late pre-orogenic and an early post-orogenic rock) is applied over great distances along and across a deformed zone. Such has been the case, for example, with the "Laramide orogeny," a term derived from early studies of an accessible site along the Front Range of northern Colorado having suitable rocks, and then applied over a broad region where independent dating was not done until much later. Even after more dating was done, specific field relations or laboratory procedures for dating varied, so that obtaining slightly different ages of deformation was of little concern. Soon an abundance of literature on the "Laramide" was established, and "its" age became set, now less commonly at the age of the original specific site at the Cretaceous-Tertiary boundary, but more commonly as including the Late Cretaceous, Paleocene, and Eocene. So broad an application of an orogenic time does not help in a regional tectonic synthesis in which it is essential to follow with some precision the development of deformational responses to a spreading stress field. In this synthesis I will therefore emphasize the numerical ages and their geologic sources more than the term "Laramide."

Analogous to the "Laramide" case, in central Utah another terrane was found to be deformed in Early Cretaceous time in a different style than the Front Range rocks. The separate name of "Sevier orogeny" was applied in Utah for expediency apparently, and subsequently that term, too, was applied over a broad region, even though the age of the Sevier elsewhere in the region did not correlate precisely with that of the type Sevier of western Utah. As a result, we now contend with two "orogenies" in one major orogenic belt, without any evidence that the events are so different as to justify their full separation. After all, tectonic style differences may reflect variations in tectonic environment within a single orogen, and the age difference of these orogenies has steadily been shrinking as the term "Laramide" is applied ever more loosely to terranes between those of the type Laramide and type Sevier. My point here is that, although, indeed, there are variations of deformation style and age within an orogen, many of them are normal variations that reflect the largely systematic changes in tectonic environment, rate and duration of stress application, and rock strength variations. Where changes are not quite systematic and consequently style or age of deformation changes more abruptly, distinctive terranes do develop but, I submit, these are better labeled as phases of an orogeny than as separate orogenies. Thus, in this synthesis I shall view the deformed belt of the Cordilleran region as having been formed during a Cordilleran orogeny, and the "Laramide" and "Sevier" are phases of this orogeny, characterized, as is currently accepted, by differences in age and style of deformation. The following review of the age of deformation along the Texas-California transect will illustrate the merits of this view of an orogeny involving the phase ranking of tectonic events otherwise viewed as separate orogenies.

The relations of tectonic stress to deformation may be simplistically illustrated on a time-space diagram (fig. 30) by means of which I shall introduce a more realistic diagram. Assuming that stress is applied and then released gradually, it may be shown as a sine curve (fig. 30*B*). The stippled wedge (fig. 30*A*) represents the deformation, which lasts from time X to X + 40 m.y., that results from the stress. Peak deformation occurs about during the time of maximum stress application, and in a major orogen this may be the time of development of thrust faults. Such structures may take a relatively short time to develop compared to the duration of waxing and waning orogenic conditions; but they are the spectacular structural features; thereby, efforts of dating the orogen focus on obtaining ages, *a* and *b*, of some youngest deformed rocks and ages, *a′* and *b′*, of some oldest undeformed rocks, commonly intrusive igneous rocks. Empirically, we find the available age range at *a–a′* to be greater than at *b–b′*, and usually the ± bar, or reproducibility factors, of ages are greater at *a–a′* than *b–b′*. While figure 30 is meant to be a simple case, depicting the distribution of deformation in time and space as a wedge already accommodates the somewhat ductile property of rocks, particularly under thick cover. Because of this partly ductile condition, much stress is absorbed by the hinterland rocks and less stress remains to act on rocks of the foreland; thus, the wedge shape to the time-space field representing an orogeny.

With these introductory remarks, we can turn to a more realistic representation of the timing of the Cordilleran orogeny along the Texas-California transect. The more elaborate and more realistic space-time diagram of figure 31 summarizes the age controls of deformation available from the sedimentary and igneous record. In essence, the diagram shows that: (1) the orogeny endured from at least Early Cretaceous (possibly even Late Jurassic) time in the west to Eocene time in the east; (2) the orogeny lasted longer in the west than in the east; (3) the orogeny was mostly older in the west than the east; and (4) the orogeny probably comprised more phases in the west and so was more complex in the west than in the east.

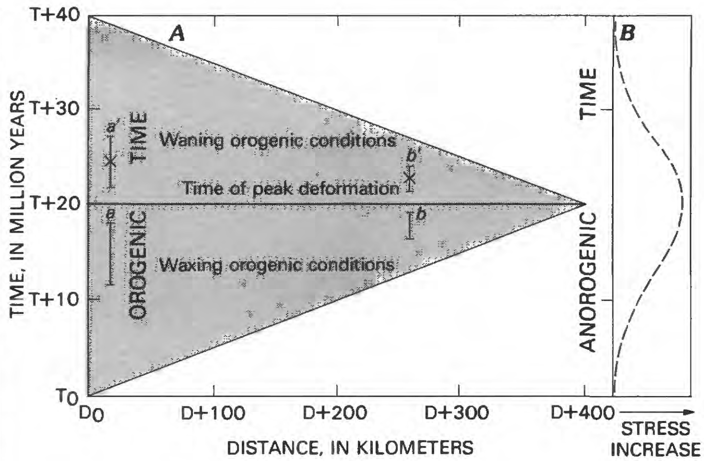
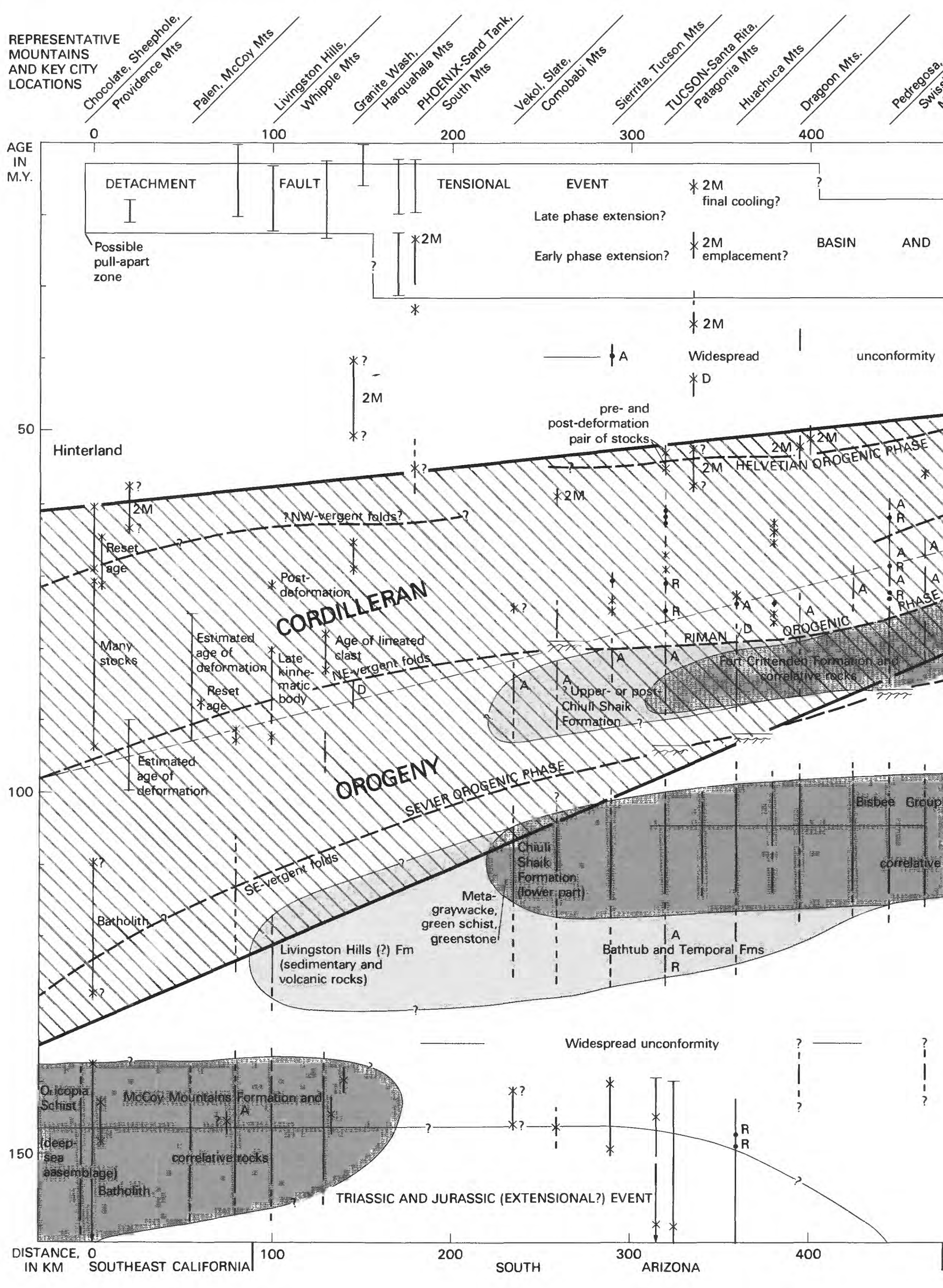
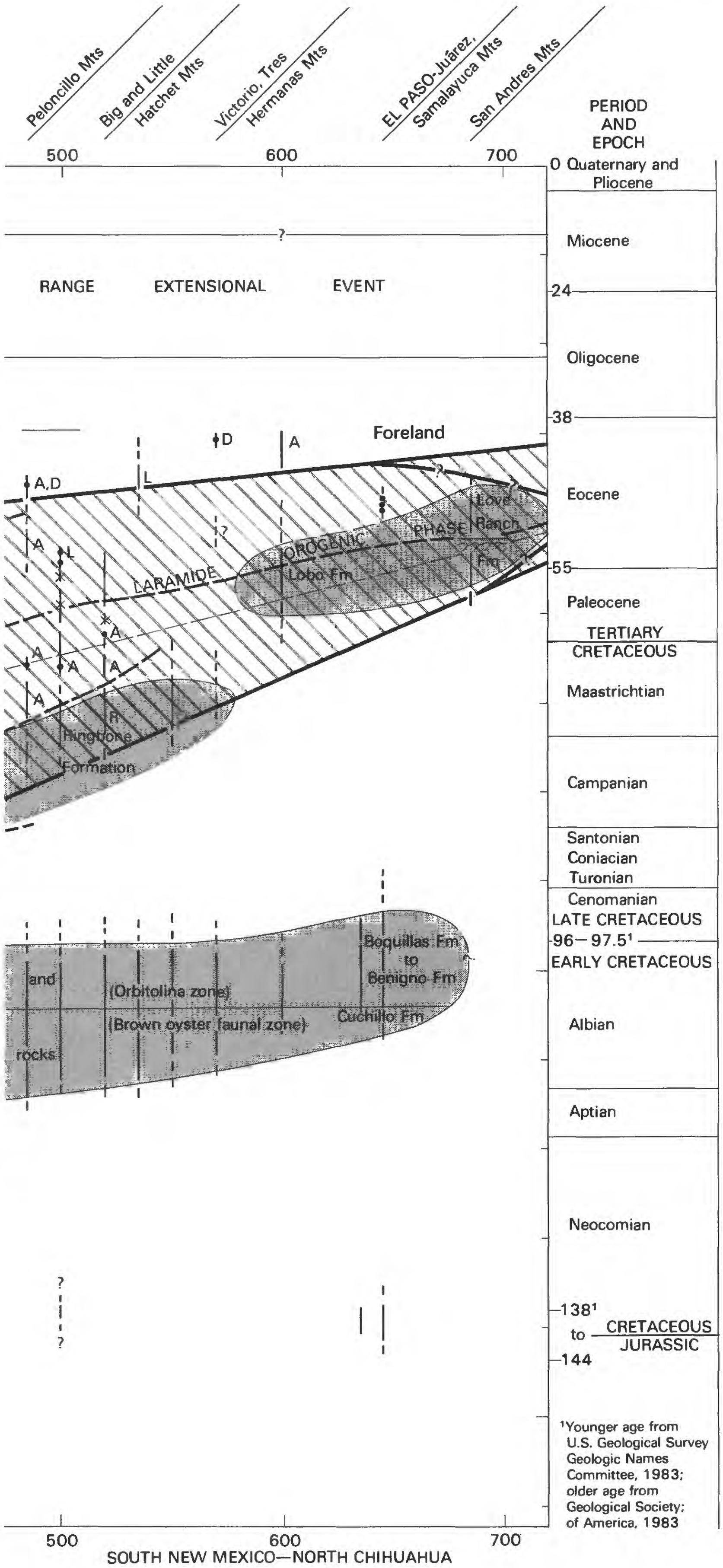


FIGURE 30.— Simplified time-space diagram of an orogen, showing the likely distribution of deformation. *A*, in response to stress. *B*, Peak deformation is typically dated, sites *a* and *b*, through the ages of youngest deformed rock, here taken to be sedimentary rocks having ages shown by lower bars, and the oldest undeformed rocks, here taken as plutons. See text for further discussion.

In the construction of figure 31, I adopted the following procedures. All data were projected normally into a single transect line. Near the foreland zone this line trends N. 70° E., parallel to the direction of tectonic transport, but in the hinterland zone the line was changed to east-west, about normal to a possible core zone and about midway between two common transport directions. Data from a row of mountain ranges normal to a point on this line are composited, and only one or two representative ranges are listed above each column. Sources of these data were obtained from nearly 100 references, too many to be listed separately on the illustration; however, many of them are already incorporated in the text sections on structural style, and others are given in such recent guidebooks, symposia, and syntheses as Callender and others (1978), Lovejoy (1980), Drewes (1981a, 1982b), and Frost and Martin (1982). Local interpretations offering supporting evidence were preferred to regional interpretations lacking such evidence.

In this construction, supporting data are both more abundant and better along the eastern half of the transect than along the western half, largely because in the east deformation and metamorphism are milder; whereas, many western studies are still in progress, and the separation of Cordilleran features from older and younger ones is more difficult to make in the west. With so many western studies still incomplete and with the geology so complex there, a wide variety of interpretations on particular issues is available; I favor those accounts reporting more than the presence only of


REPRESENTATIVE MOUNTAINS AND KEY CITY LOCATIONS
Chocolate, Sheephole, Providence Mts
Palen, McCoy Mts
Livingston Hills, Whipple Mts
Granite Wash, Harquahala Mts
PHOENIX-Sand Tank, South Mts
Vekol, Slate, Comobabi Mts
Sierrita, Tucson Mts
TUCSON-Santa Rita, Patagonia Mts
Huachuca Mts
Dragoon Mts.
Pedregosa, Swisshelm Mts
AGE IN M.Y.
50
100
150
DETACHMENT FAULT TENSIONAL EVENT
final cooling?
Late phase extension?
Possible pull-apart zone
Early phase extension?
emplacement?
BASIN AND
Widespread unconformity
pre- and post-deformation pair of stocks
Hinterland
HELVETIAN OROGENIC PHASE
NW-vergent folds?
Reset age
Post-deformation
CORDILLERAN
Many stocks
Estimated age of deformation
Late kinne-matic body
Age of lineated clast
NE-vergent folds
RIMAN OROGENIC PHASE
Fort Crittenden Formation and correlative rocks
Upper- or post- Chiuli Shaik Formation
Estimated age of deformation
OROGENY
SEVIER OROGENIC PHASE
Bisbee Group correlative
SE-vergent folds
Chiuli Shaik Formation (lower part)
Meta-graywacke, green schist, greenstone
Batholith
Livingston Hills (?) Fm (sedimentary and volcanic rocks)
Bathtub and Temporal Fms
Widespread unconformity
Orocopia Schist (deep-sea assemblage)
McCoy Mountains Formation and correlative rocks
Batholith
TRIASSIC AND JURASSIC (EXTENSIONAL?) EVENT
DISTANCE, IN KM
SOUTHEAST CALIFORNIA
SOUTH ARIZONA
0
100
200
300
400



EXPLANATION

- Deposition of sedimentary rocks; basins shown in gray
- Age range of dated plutons
- Age range of volcanic rocks and dikes
- R Rhyolite
- A Andesite
- D Dacite
- L Latite
- 2M 2-mica+ garnet granite
- Angular unconformity

FIGURE 31.—Detailed time-space diagram along the Texas-California transect of the Cordilleran orogen, showing kind and distribution of support for dating the orogeny and its phases.

a mid-Tertiary extensional event. In figure 31 only enough is shown of earlier and of older tectonic events to compare with the Cordilleran data; these events surely may be better documented.

Figure 31 shows some well-defined, mostly major features, as well as some poorly defined, generally minor features. Clearly, the distribution of deformation in time and space forms a wedge tapering eastward, as was anticipated in figure 30. Deformation lasted about 120 m.y. in the west, where a zone of batholiths may mark the tectonic core zone, but it lasted only about 12 m.y. along the frontal line in the east. Probably, the time-space wedge is blunt tipped, because the closer that the stress field is restricted to the brittle surface rock the more likely will that stress be "used up" or absorbed by an upward shearing of a reverse-fault-like frontal line. The deformation wedge is also tilted, being older to the west than the east. In other words, it took time for the stress field to propagate toward the foreland zone because of the partly ductile property of the rocks. Likewise, the reduction of compressive stress also took time to become felt in the east but, because deep erosion was already under way by late orogenic time and the rocks were becoming more brittle, less time lag was probably involved and, consequently, the top of the wedge slopes less than the bottom.

Consider for the moment the following ramification of having a westward slope of the top of a deformation wedge: at a time when compressional deformation was still in progress to the east the orogenic compressive stress had ended to the west. It is thus conceivable, although not implied to have occurred in this region, that a new stress regime—say, an extensional one—could be started to the west while compressional stress response was still in progress to the east. In practical effect, a basin-and-range event could be initiated in the hinterland of an orogen concurrently with the last compressive responses along the frontal line of an orogen. Again, I do not mean to imply that this situation occurred along this transect, but somewhere such seemingly conflicting evidence may turn up.

The time-space diagram of the Texas–California transect seems to distinguish clearly between the Cordilleran orogeny and earlier and later tectonic (or magmatic?) events. The older and comparatively little-known Jurassic event seems to be time-uniform across southern Arizona and southeastern California; whereas, the Cordilleran orogeny is time-transgressive, which suggests that the two events underwent different genetic processes. Similarly, the younger and well-known basin-and-range tectonic event (or the basin-and-range plus detachment-extensional event) results from a process that acted concurrently across the region. If the younger tectonic event is time-concurrent because the rocks were more brittle, being near the surface, then the ductile deformation attributed to extension appears irrational. Yet, if the rocks were ductile during this event one would need a mechanism of deformation that is applied everywhere at once rather than one that is propagated gradually. Either way, I conclude that the Cordilleran orogeny and coextensive basin-and-range event are separate tectonic developments and tectonic overprinting is rampant here.

The time-space diagram further serves to show the relations between tectonic, igneous, and sedimentary activity. Pre-orogenic and early orogenic sedimentary basins seem to form discrete entities, although few deposits are well dated throughout their sequence. Provisionally, then, these staggered basins or depocenters are seen as being separated from one another by unconformities and, in the case of the Bisbee Group–Fort Crittenden Formation contact, by an angular unconformity that may be a distal record of the Sevier phase of the Cordilleran orogeny, centered about western Utah. As a group, the depocenters are time transgressive, as shown in figure 28, and shift sporadically, perhaps reflecting minor irregularities in a generally systematically changing environment. Only the erosional remnant of the Bisbee basin, taken by itself, appears to be slightly time transgressive, younging to the east, for it has a widespread *Orbitolina* zone that may be considered to mark a time line running askew to the transverse profile of the basin.

Igneous rocks provide the basis for most of the available age determinations; of these rocks only a few representative ones are shown in figure 31. In some places andesitic volcanic rocks predate tectonism; elsewhere, they postdate it; and, in still others, such as the Pyramid Mountains (fig. 16) (Thorman and Drewes, 1978), both situations occur. This relationship has prompted the placement of an andesite "shell" around the pluton zone in figure 29. Plutonic rocks commonly provide the upper limit of the age range of deformation. Among the plutons, the oldest emplacements of 2-mica granites overlap in age the last phase of deformation. Emplacement of the youngest, however, extend into the basin-and-range event, a situation that hints at a genetic link between the two tectonic events that I cannot explain at present.

The time-space diagram also demonstrates that the Cordilleran orogeny is a composite of smaller, more localized phases. Only the fold-and-thrust and foreland zones have but a single phase of peak deformation; two deformation phases are recorded in the middle reaches of the orogen, and three deformation phases are present in several of the western ranges. The dating of these separate phases is, of course, not as good as that of the whole orogeny; particularly, the dating of the oldest of the three western events is nebulous. Nevertheless, the patterns of eastward younging and, locally, the clear-cut separation of adjacent phases fits well with the overall concept of the anatomy of the orogenic belt. Of these phases, the single phase in the foreland matches in style and age the classical Laramide phase. Likewise, the oldest of the western phases may be assigned to the Sevier phase, whose southeasternmost signs probably are the angular unconformities between Upper and Lower Cretaceous rocks (Drewes, 1971b; Hayes and Raup, 1968). Other phases are named for the local record near Tucson (Drewes, 1972b) and, of these, the Piman phase appears to be most widespread along the Texas-California transect.

This review of the data bearing on the age of deformation provides abundant support that the Cordilleran orogeny extended across southern Arizona and New Mexico. Furthermore, it clearly separates in time and kind this tectonic development from the basin-and-range and (or) detachment faulting event. The orogeny is tied to the development of a series of foreland basins that shift, somewhat sporadically, eastward as the site of tectonism propagated eastward, always remaining on the eastern flank of the orogen, and each basin (except the last) later being incorporated in the deformation. Finally, the orogenic event is shown to be a composite of several discrete phases, reflecting, in part, events centered well outside the region of the transect, as well as one or two of more local derivation.

REGIONAL STRESS AND ITS ORIENTATION

The evidence of the kind of regional stress and orientation of that stress which formed the orogenic belt is derived from a study of local structural features. Along the Texas-California transect many of these structural features were formed as a result of compressional stress oriented with a maximum stress direction subhorizontal and east to northeast. Shear features indicate that generally upper masses moved east to northeast relative to lower ones. From consideration of global plate movements, both masses were actively converging; the Atlantic and Pacific spreading centers, from which the plates were propelled, were both active during late Mesozoic time.

A compilation of inferred directions obtained from assorted kinds of observations by many workers along the transect shows both a systematic pattern and some diversity (fig. 32). Taken indiscriminately, diversity of transport directions is most apparent, although statistically a northeast-southwest orientation of transport, and therefore, causative stress, is apparent.

The conventional observations of fold orientation, strike-slip faults associated with thrust faults, and certain kinds of lineation are used to infer orientation of transport direction. Likewise, fold vergence, drag features and *sc* fabric help to determine direction of transport in the northeast-southwest orientation. Local studies by themselves sometimes provide too limited a data base for interpreting certain observations and thus have generated some needlessly diverse conclusions; these I shall simply reinterpret from the original observations along the lines indicated in the description of mountain ranges zone by zone. Source sites are listed in the caption of figure 32 and references to the sites appear mainly in the section on "Style of Deformation" and in the articles collected in the key symposia, guidebooks, and syntheses listed in the section on "Age of Deformation." The term "vergence," is used herein for the vector normal to the line of intersection between two planes in the direction of divergence of these planes, which typically are the axial plane of a fold and of a genetically related thrust fault. Vergence is herein not used as a synonym with "inclination," as it appears in some recent literature. Vergence (sensu strictu) indicates tectonic transport direction; "inclination" and its equivalent *may, but need not*, indicate the same.

Folds, reverse faults, and thrust faults emplacing older rocks over younger ones occur throughout the region. These structures have had the net effect of thickening the rock column and consequently are inferred to result from horizontally applied compressive stress. That this stress was a regional one, and not a series of locally derived stresses, depends on the uniformity (or systematically changing condition) and extent of the evidence of one stress condition. The solid arrows of figure 32, pointing predominantly east to northeast, are believed to reflect the main deformation in the region of the transect, including the foreland zone and the Colorado Plateau. In detail, these vectors shift slightly, being more easterly near Las Vegas and more northerly near El Paso. A few exceptions occur, most of them for known cause. Among the exceptions are some sites near the northeastern margin of the mobile belt where the transport was northerly. Near some northwest-trending reactivated basement flaws, transport directions have a northwest-southeast orientation. Also, west of Phoenix some ranges have evidence of two or three major transport directions of Cordilleran age, from which several phases of diverse movement are inferred.

Evidence of southwest-directed transport is present in a number of ranges. Commonly, this evidence occurs close to sites that also provide evidence of the dominant northeast direction of movement. Such a record has led to the proposal of diametrically diverging tectonic transport off a northwest elongate major uplift of Late Cretaceous age (Davis, 1979). However, there is neither

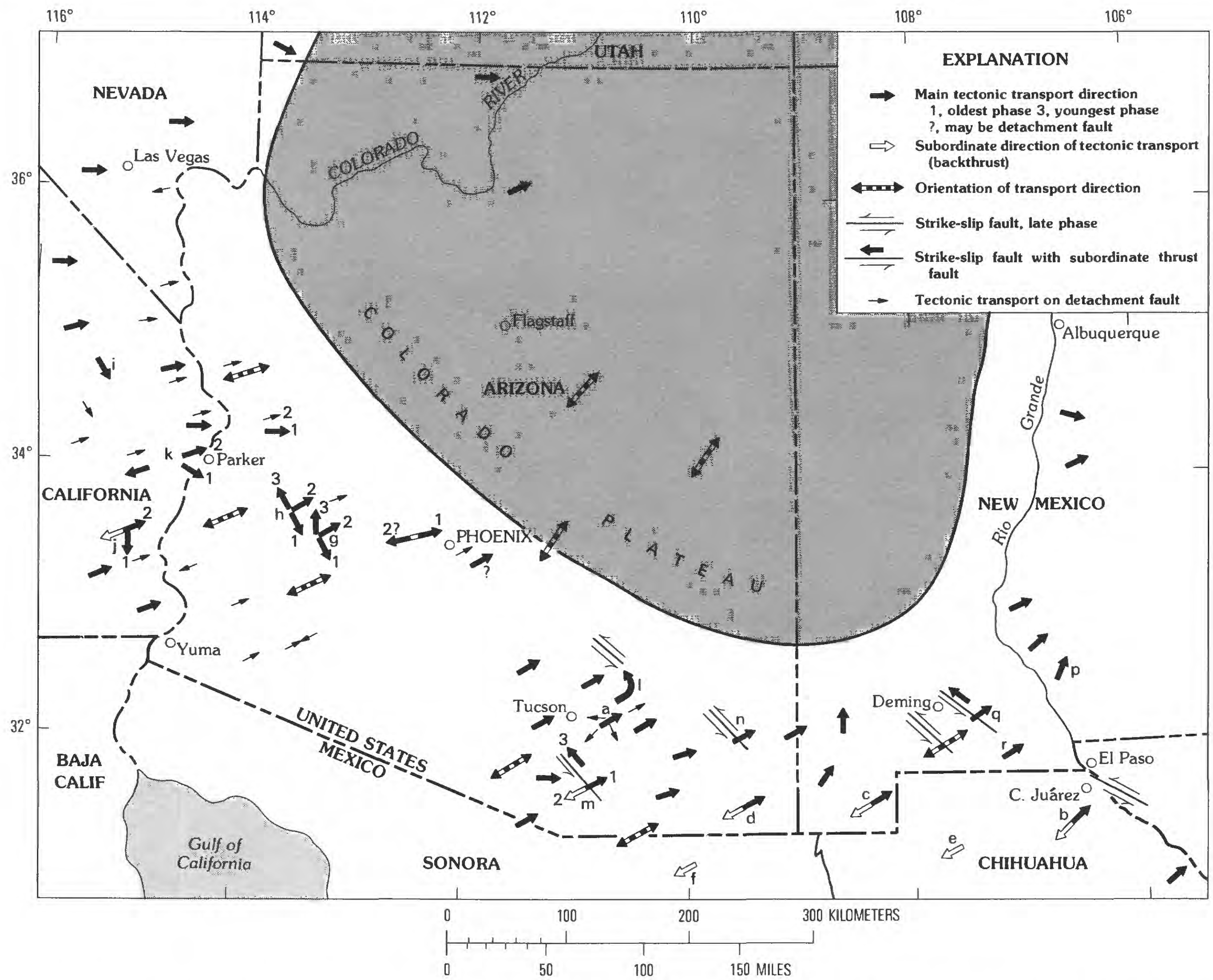


FIGURE 32.—Map showing distribution of data bearing on direction of tectonic transport during the Cordilleran orogeny and the detachment fault event, between El Paso, Texas, and Las Vegas, Nevada. Key to sites referred to in text, with exception of site f: a, Rincon Mts.; b, Sierra Juárez; c, Big Hatchet Mts.; d, Pedregosa Mts.; e, Sierra de los Chinos; f, Sierra Anibacachi (Rangin, 1978); g, Harquahala Mts.; h, Granite Wash Mts.; i, Piute Mts.; j, Little Maria Mts. and adjacent hills; k, Riverside Mts.; l, northeast flank, Rincon Mts.; m, Santa Rita Mts. and adjacent hills; n, northern Chiricahua Mts.; p, San Andres Mts.; q, Florida Mts.; and r, East Potrillo Mts.

any indication of such a systematic diverse orientation of tectonic transport, nor supporting evidence in the stratigraphic and structural record of such uplift. Furthermore, the best preserved local record of Upper and Lower Cretaceous rocks occur on the proposed uplift, where mainly old rocks should be exposed. Finally, scattered evidence of southwest-directed movement occurs in sites well away from the proposed uplift.

As an alternative interpretation to the concept of diametrically diverging gravity-propelled movement off an ancient uplift, I suggest that the southwest-directed transport is related to backthrusts, features that are subordinate to the major thrust faults and that occur in all orogenic belts. A recent illustration of a typical backthrust, in the Pakistan Himalayas, is shown in the eastern third of a structure section of the figure 2 of Conrad and Butler (1985). Small backthrusts have been described from the Sierra Juárez in the section on the fold-and-thrust zone (figs. 10 and 11), as well as from the Pedregosa Mountains of the western intermediate zone (fig. 20).

Finally, there is the evidence in the more ductily deformed rocks of diverse directions of tectonic transport provided by the orientation of lineation, *sc* fabric, and very small folds or wrinkles. These features are either entirely of late Tertiary extensional origin, according to some workers (Davis, G.A., 1983; Davis, G.H., 1983), or as of mixed early compressional and later extensional affinities, according to others (Reynolds, 1982; Thurn, 1982). In figure 32 this diverse evidence is shown separately, using slender arrows where inferences only of a young origin are made. Also included with this body of evidence are the inferences made from folds in the carapace around the Rincon Mountains (fig. 32, site a) that are accepted by all workers as having had at least a late history of gravity movement off the gneiss-cored dome. Remarkably, however, the lineation in the core rocks of this dome is unidirectional, providing evidence that this fabric was not formed at the time of, or under the conditions of, the late gravity radial movement.

In some ranges of western Arizona and southeastern California the evidence of lineation, *sc* fabric, and minor wrinkles suggests several directions of tectonic transport. Wherever age relations between these directions are known, the direction can be shown to have shifted counterclockwise through Cordilleran time. Such a systematic shift in stress orientation is still compatible with the hypothesis that the region was deformed by major compressive stress during Cordilleran time; global plate movements do shift gradually. The earliest movement phase may have been generated from the Sevier belt of western Utah, as proposed for the observations in the Riverside Mountains (fig. 32, site k) by Lyle (1982). The next, or main movement may be part of the Piman phase (fig. 31), of southeast Arizona studies, but the "peak deformation" lines only fit imperfectly. The third phase is still too little known to deserve much discussion; if indeed, it is a genuine separate event. Evidence of diverse movement in the western part of the transect may, alternatively, be viewed as the products of a single stage of deformation in which laminar flow of highly ductile rocks differed from layer to layer, reflecting unknown local variations in rock composition or temperature, and so on. The crux here is that, regardless of the explanation favored, these structural features are best explained as having been formed in a mobile environment compatible with an older (Cordilleran) event, beneath a thick cover, rather than with a younger (Miocene extension) event beneath a shallow cover.

In the eastern part of the transect, where the northwest-trending basement faults have segments reactivated as strike-slip faults during late orogenic time, some thrust faults have a northwest direction of transport. These features are believed to have formed in response to local stress fields near the tear faults. Similar directions of transport inferred from evidence in the Harquahala and Granite Wash Mountains (fig. 32, sites g and h) may, however, have encouraged development of a regional explanation.

On the northeast flank of the Rincon Mountains (fig. 32, site l) the evidence of lineation as a guide to tectonic transport offers another complexity. Over an area of at least 10 km^2, the lineation, which elsewhere trends N. 70° E., gradually trends north, and, ultimately N. 10° W., along the northeast flank of the range. Perhaps the still ductile rock, having received its fabric imprint during the final movement of the Piman phase—about during early Paleocene time—was broadly warped or drag-folded near a northwest-trending strike-slip fault, such as the Mogul fault at the north end of the Santa Catalina Mountains (Wilson and others, 1969), whose southeastern extension is concealed by young gravel deposits near the lineation flexure. Late Paleocene strike-slip movement is documented on such faults in the Santa Rita Mountains (Drewes, 1971a, 1971c, 1972b) and in the Chiricahua Mountains (Sabins, 1957a; Drewes, 1982a).

This review of observations and inferences regarding the kind of stress and orientation of stress confirms the complexity of the Cordilleran orogeny. Signs of major compressional deformation occur across all tectonic zones, as well as to the northeast beyond the mobile belt. Much of the evidence shows a consistent, or a systematically changing, orientation of the stress, more easterly near Las Vegas and more northerly near El Paso. Indeed, northeast of Las Vegas this transport direction is directed southeast. Perhaps the southwest salient of the Colorado Plateau block acted as a tectonic buttress against which, and around which, the mobile terrane converged, much as the trajectory of waves converge around a headland on a shoreline.

Other tectonic transport directions of Cordilleran age are also recognized. They reflect backthrusts, local late thrust faulting near active strike-slip faults and, to the west, a terrane probably of polyphase deformation or possibly of some erratic or less constrained sort of ductile flow.

The evidence of Cordilleran direction of transport can generally be distinguished from that of the Miocene extensional event of the hinterland zone. In many places there, the overprinted compressional fabric has the same orientation as the younger extensional fabric, a situation once again hinting at some genetic tie between the Cordilleran orogeny and the basin-and-range detachment faulting event.

AMOUNT AND RATE OF TECTONIC TRANSPORT

Estimates of the amount of tectonic transport that has occurred across an orogenic belt are always difficult to make because of the scarcity of hard data, and the Texas-California transect of the Cordilleran orogenic belt is no exception. Along various transects of this belt three methods have been used to estimate tectonic transport amounts; and in each place and with each method that amount is 150–200 km. In each case, I suspect that even these values err on the conservative side, for there likely are deformational features that have escaped the various talleys.

Along a transect in southern Canada, Price and Montjoy (1970) arrived at their estimate through a summation of the increments of tectonic transport on all known faults. Because subsequent extensional faulting has offset or occurred along most of the individual thrust faults and because the extensive cover of intermontane gravel and of post-orogenic volcanic rocks, such a summation method is impractical along the Texas-California transect.

Along a transect in southwestern Wyoming–northern Utah–northeastern Nevada, Crittenden (1961) estimated the amount of tectonic transport through a measure of offset between thrust terranes of particular stratigraphic isopleths. This method is also difficult to apply along the Texas-California transect because suitable stratigraphic data are not yet established. Furthermore, this method probably will be difficult to apply here because of the inadequacy of the distribution of stratigraphic sections near the mobile zone. However, hints that diverse stratigraphic facies have been tectonically juxtaposed occur in a few localities in southeastern Arizona (Drewes, 1981b, 1987, 1989, 1990c).

Along the Texas-California transect, the "sheared rivet" method has been proposed to arrive at an amount of tectonic transport of about 200 km in southeastern Arizona (Drewes, 1981a). A pre-orogenic stock, the Squaw Gulch stock of the Santa Rita and adjacent mountains, is tentatively thought to be cut and offset by the upper of two major thrust faults, the Cochise thrust fault, with the top part offset about 100 km in a N. 70° E. direction relative to the lower part, and known as the Gleeson stock of the Dragoon Mountains. A similar amount of offset must be assumed along the lower thrust fault. The test of this method depends on the strength of the correlation between the two stocks and of their host rocks; in this case, correlation is suggestive but is inadequate to prove conclusive.

The various estimates offered (Drewes, 1976, 1978, 1981a) for the amount of tectonic transport in individual ranges along the Texas-California transect, plus some general observations of the rock structures and fabric suggest that, regardless of the acceptability of the 200 km estimate in southeastern Arizona, the amount of transport was less to the east and more to the west. In several areas of the foreland zone (or possible subfold-and-thrust area), such as the San Andres Mountains (fig. 32, site p) and Florida Mountains (site q), a few kilometers of shortening through folding and faulting accounts for the structures observed. In the Sierra Juárez (site b), Nodeland (1977) and Lovejoy (1980) suggested that deformation within the ranges is a few tens of kilometers; apparently they did not consider the movement on a major décollement fault that remains concealed beneath the range and the nearby part of the Rio Grande Valley but crops out along the Rio Grande in West Texas and adjacent Chihuahua. In the eastern intermediate zone of southwestern New Mexico, the overthrusting, shingled thrusts, and various upright, inclined, and recumbent folds, together, suggest a tectonic transport of many tens of kilometers to perhaps 100 km.

The marked structural complications of the Dragoon Mountains (fig. 32, site n) and flanks of the Rincon Mountains (fig. 32, sites a and l) require far more than merely mid-Tertiary gravity or extensional faults (Thorman, 1977b; Thorman and Drewes, 1978). In the Dragoons, Proterozoic crystalline rocks are massively interleaved with Phanerozoic sedimentary and metamorphic rocks. A large overturned fold and fields of small tight folds occur, and sedimentary and metamorphic facies are juxtaposed. While no firm estimate of transport amount is possible from such relations, an amount of many tens to a few hundred kilometers seems reasonable.

In western Arizona, the tectonic complexities like those of the Dragoon and Rincon Mountains continue and other complexities are also present. Ductile flowage is common in the Phanerozoic rocks of thrust-plate remnants, with a thickness reduction in some instances to a thousandth of the probable initial thickness of a sequence. Crystalline rocks are penetratively foliated (that is, foliation is as common far from some low-angle faults as near them, barring perhaps the complications of the first few meters beneath some faults). The widespread *sc* fabric in this foliated terrane and the ubiquitous lineation bespeaks of distributive shear, largely related to Cordilleran tectonism. However, some post-Cordilleran development of these features, such as cataclastic deformation, also is recognized. Altogether, these features and the several estimates of movement of a minimum of several tens of kilometers (Davis and others, 1980), suggest that a total amount of tectonic transport during the Cordilleran orogeny could easily be 200 km, and even 300 km would not be out of line.

The rate of tectonic movement commonly is calculated along the spreading center, or generating edges, of global plates because along their collision zones, or degenerating edges, evidence of such movement is more difficult to get. For instance, the northeastern Pacific plates were calculated by Atwater (1970, p. 3515) to have moved northeasterly from the East Pacific Rise about 5 cm per year through at least late Mesozoic time. A somewhat smaller movement rate is proposed for the westerly transport of the North American Plate from the Mid-Atlantic Rise. Plate convergence, then, may have been about 8 cm per year for a substantial span of time.

The consequence of this convergence was the development of the North American Cordilleran orogen, a part of which is herein under scrutiny, and also the development of a subduction plate beneath this orogen. Some observations derived from both the near-surface and the subducted terranes provide bases for estimating a 1–2 cm per year rate of differential movement between the plates (or within the orogen) along this part of the degenerating edge of the plates.

The case of plate movement rate reflected in the subducted plate is derived from the belief that this plate melts to generate magmas that then rise through the underridden part of the North American Plate. This case was presented by Drewes (1982c) without benefit of placing those observations and deductions in the context of a tectonic synthesis. In brief, that contribution was based on the premises that the magma generation rate is likely to be spatially systematic, given uniform or uniformly changing conditions, and the egress of that magma to hypabyssal or surface levels also will probably be spatially systematic, given uniform conditions in the underridden edge of the North American Plate. In a general way, the rate of gradual eastward younging of the orogenic igneous rocks is an expression of the gradual convergence of plates, minus the rate of melting of the subducted plate.

The systematic southeastward shift of a particular set of granitic stocks was recognized by Sillitoe (1972) and Livingston (1973a, 1973b) in connection with porphyry copper deposits and a hot-spot idea. Instead of utilizing a hypothetical hot-spot that shifts along a vector parallel to sites *a–a′* (fig. 33), an alternative explanation is favored. This alternative involves shifting sites of magma leakage along a pre-existing, partly concealed northwest-trending basement fault (nonreactivated segment) during the east-northeast shift of orogenic features, including a zone of disintegrating subducted plate. These aligned stocks are the first that were emplaced in their areas after peak deformation (locally the Piman phase). Simple geometric calculation, based on the relationships shown in figure 33 yield a rate of shift of magma emplacement, or movement of comparable zone of subducted and melting plate, across part of the orogenic belt of 1 cm per year (Drewes, 1982c). (Note also that mining camps are alined on the northwest-trending faults.)

The rate of plate convergence near the collision zone is also reflected near the surface by the rate of tectonic transport across the width of the orogen. This tectonic transport rate should be less than the plate convergence rate because much energy was dissipated in deforming the terrane west of the Texas-California transect and was dissipated near the concealed collision zone itself. In figure 31, the slope of the peak deformation line, approximately bisecting the tilted wedge depicting the distribution of the orogeny in time and space, is a measure of the average rate of stress propagation. This line shows the propagation of the stress field across a belt about 700 km wide in about 70 m.y., or about 1 cm per year. Accepting the 200 km amount of tectonic transport around Tucson, described earlier, movement, therefore, consisted of 2.2 cm per year over 45 m.y. Both stress propagation and tectonic transport values are average rates; they were probably higher during peak deformation times of local orogenic phases, and less during the more quiescent intervals.

Tentative as these individual measures of tectonic transport or stress propagation rates—8, 2.2, 1, and 1 cm per year—may be, they are in good agreement with one another. Perhaps most impressive is the fact that these rates were calculated on the basis of different kinds of observations.

TECTONIC ENVIRONMENT

The load or thickness of rock cover beneath which deformation occurs is a major factor in determining the kind of deformational response rocks make to a given stress. In general, the thicker the cover, the more readily is compressional stress transmitted. Along the Texas-California transect, the question of load adequacy for substantial stress transmission has occasionally been raised, apparently because the Paleozoic sequence is viewed as having a thin cratonic affinity rather than a thick miogeosynclinal affinity. In lithologic facies, the Paleozoic rocks of southern Arizona are more like those of the miogeosynclinal rocks of eastern Nevada than those of cratonic northeastern Utah or western Colorado. Thickness studies of southern Arizona and New Mexico rocks have been made only at scattered localities and commonly only for some units, but they have not been undertaken systematically. Also, we have no knowledge of the lower threshold of load beneath which compressive stress may be transmitted. Apparently, other contributing factors, such as rock strength, effective pore fluids, duration, and rate of stress application are also sufficiently unknown, so that there has been little incentive to scrutinize only the load factor. However, because the question of load adequacy is raised, the topic of cover thickness is reviewed in this section.

A construction of the cumulative thickness of cover in the Tucson area is presented in figure 34 to illustrate the

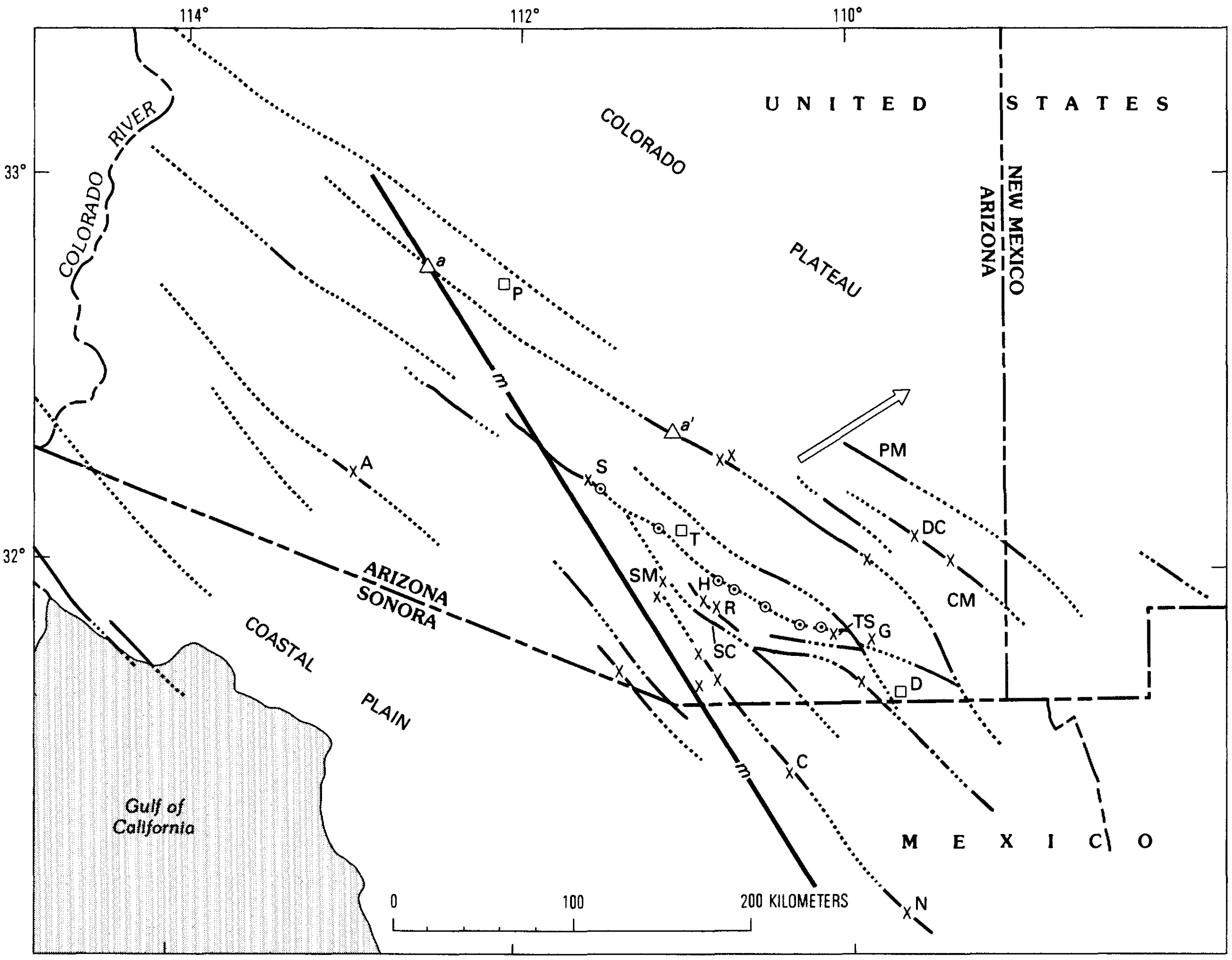


FIGURE 33.—Map of southern Arizona showing location of known (solid line) and inferred (dotted line) faults of the northwest-trending system of basement flaws and some major ore deposits. Open arrow shows inferred direction of tectonic transport of thrust faults near the surface and of the proposed subduction plate at depth. Line, m, shows orientation of melting zone at leading edge of subducted plate, much generalized. Circled dots are sites of a suite of the earliest stocks of Late Cretaceous age near Tucson. Open triangles are reference sites discussed in the text. Location of mining districts, x and/or towns, : A, Ajo; C, Cananea; CM, Chiricahua Mountains; D, Douglas, DC, Dos Cabezas Mountains and chief mining districts; G, Gleeson; PM, Pinaleno Mountains; H, Helvetia; N, Nacozari; P, Phoenix; R, Rosemont; S, Silver Bell; SC, Sawmill Canyon fault zone; SM, Sierrita Mountains; T, Tucson; and TS, Tombstone.

fragmentary nature of the depositional record. Of particular concern is the deposition under partly or wholly subaerial conditions of late pre-orogenic time, for their full thickness and extent are markedly affected through late-orogenic and post-orogenic erosion. Available control points are shown as dots along a pair of curves that represent the cumulate of thickness maxima and minima. Some key formations are identified above these curves. By the end of Paleozoic time only about 2 km of sedimentary rocks had accumulated, and the range between the two curves is negligible. But, beginning in the early Mesozoic, conditions changed; thick sequences of volcanic and continental sedimentary rocks were deposited, leading to a sufficient spread between the maxima and minima curves that increased enough so that an intermediate "likely" curve is added (heavy line), reflecting my evaluation of the conditions near Tucson. The final sharp peak on the "likely" curve is given an assumed 15–20 percent of thickening, as proposed by Price and Mountjoy (1970) for conditions around peak orogenic time in a compressionally deformed belt. The construction, then, suggests that near Tucson the likely thickness of cover just before peak orogenic time (Piman phase) was about 8 km, and that this likely cover may have been tectonically thickened during peak deformation time to 9–10 km.

Most localities along the Texas-California transect provide fewer controls on the thickness than does the Tucson area, and so more projections must be made in preparing a regional summary of thickness of cover

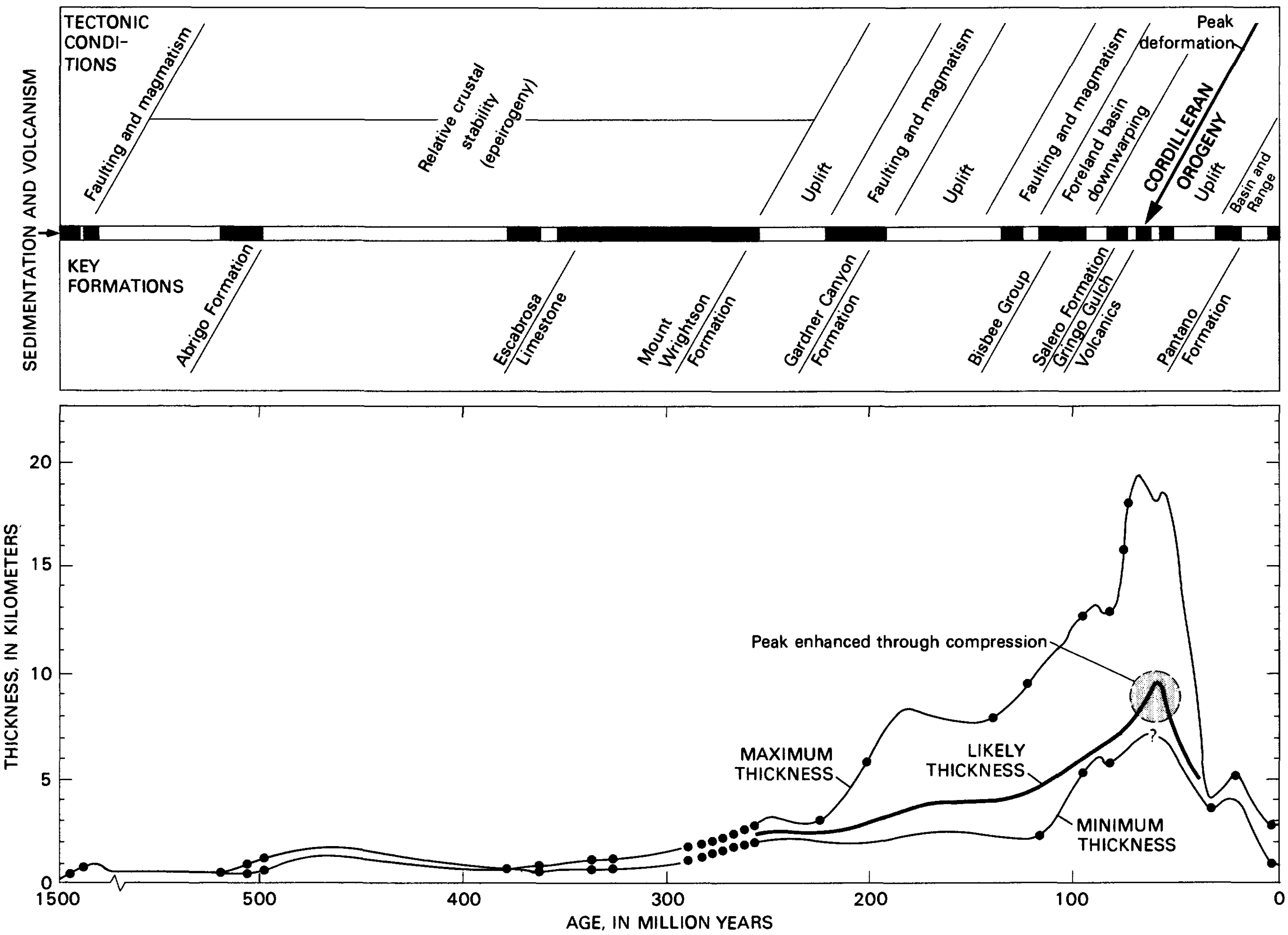


FIGURE 34.—Profile of thickness of deposits over crystalline basement rocks in the Tucson, Arizona, area through geologic time.

(fig. 35). To the east, for instance, the stratigraphic horizon of the basal décollement is not firmly established. On the other hand, to the west, remnants of Paleozoic rock have commonly been ductily thinned and metamorphosed, and Mesozoic rocks are found in a few areas where they are difficult to date and to correlate with those of other areas. A thickness of 2 km of Paleozoic rocks is assumed because most correlations of these rocks are made to the fairly thin sequences of southeast Arizona or of the Grand Canyon area and only a few correlations are made to the thick sequence of the region west of Las Vegas. To this Paleozoic thickness is added a much thicker pile of Mesozoic rocks, which have all the characteristics of widespread foreland basin deposits and little evidence in support of their deposition in local structural basins, such as the one proposed for the McCoy Mountains Formation (Harding and Coney, 1985). The usual evidence of a nearby provenance as indicative of a local basin of deposition has little merit in a mobile situation of shifting depocenters and "tectonic centers." The logic of the analysis may also be turned around, with the thought that the regional metamorphic conditions (to be discussed next) require a large load and so deposits like the McCoy Mountains Formation must have been widespread.

Other kinds of data reinforce the estimates of thickness of cover based on reconstruction of the sedimentary columns. Much of the metamorphism of Phanerozoic rocks of southern Arizona and southeastern California is attributed to events during the Cordilleran orogeny, or late Mesozoic time in general. A few estimates of cover thickness are available in the literature with the deductions of metamorphic grade based on stated mineral assemblage observations; in other instances grade estimates are offered without mention of specific support. Both situations are utilized in figure 35.

In general, the rocks of the hinterland tectonic zones are regionally metamorphosed, probably to an amphibolite grade, and retrograded to the greenschist facies. Age of metamorphism is typically Cretaceous; locally, dating

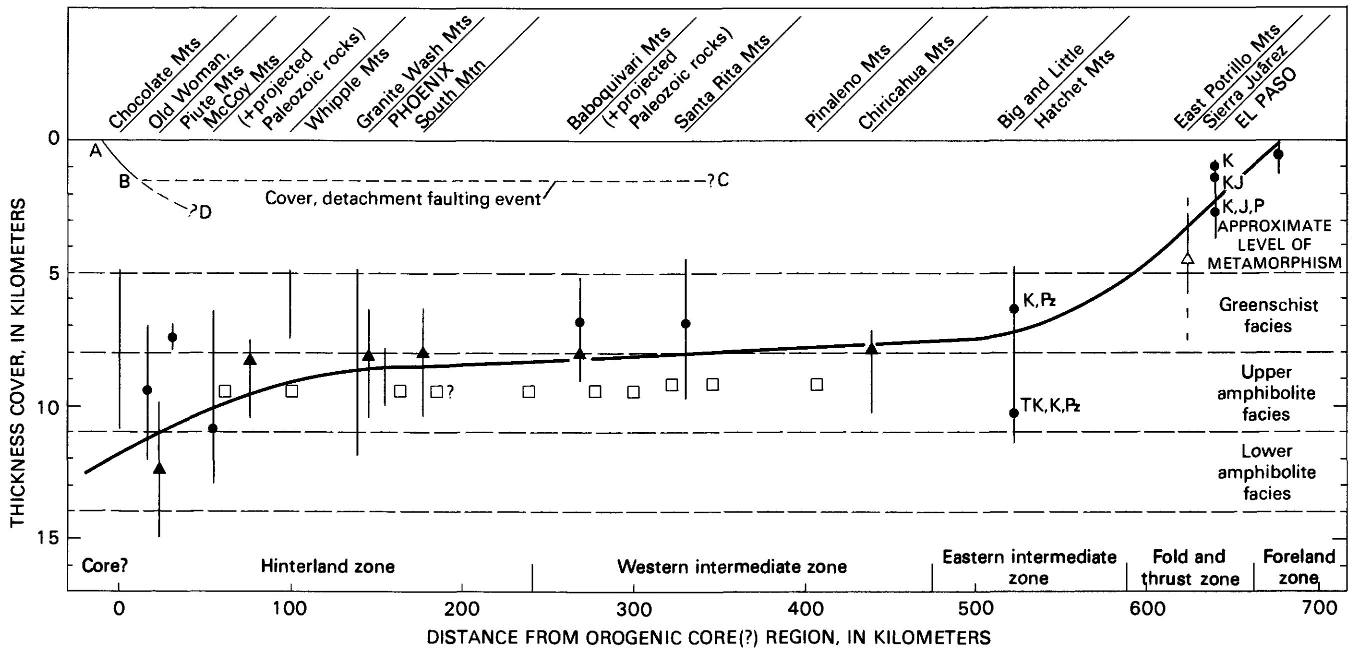


FIGURE 35.—Profile of thickness of cover along the Texas-California transect during peak deformation (that is, polychronous). Vertical bars show range of thickness based on the sedimentary record, with circles giving best estimate (or alternate estimates based on rocks of the indicated ages); solid triangles show depth of development of metamorphic minerals; open triangle estimated depth of development of cleavage; squares show depth of emplacement of 2-mica granites (shallower to east where stock tops occur), according to Davis and others (1980). From the large spread between emplacement and cooling ages of some of these stocks, an old emplacement age is assumed for all. This also avoids the problem of emplacing them at the shallow level of detachment faulting. A–D, Approximate thickness of cover over the detachment faults, which is further described in the text.

is uncertain or by circumstantial argument, as was the case in the Dragoon Mountains. A kyanite-staurolite mineral assemblage is present in some ranges, such as the Granite Wash Mountains (fig. 25), and possibly also in the Phoenix area. A few of the westernmost ranges of the hinterland zone, such as the Old Woman and Piute Mountains, even have an upper amphibolite-grade sillimanite assemblage (Miller and others, 1982). More scattered occurrences of an amphibolite-grade assemblage even occur in isolated localities of the western intermediate zone, as in the Chiricahua Mountains, where retrograde effects are seen through mineral replacement of biotite after staurolite and andalusite after kyanite. More commonly (and excluding the case just cited), this far east, the metamorphic effects are of contact origin, although across the western part of the western intermediate zone contact metamorphic halos around stocks tend to merge into a low-grade semiregional metamorphism. Greenschist metamorphism is commonly reported west of Tucson. Distinctions between prograde and retrograde effects have not normally been made across southwestern Arizona.

Observations bearing on pressure-related alterations to rocks of the foreland and intermediate-foreland zones are very scarce and provide little basis for inferring thickness of cover. Cleavage unrelated to axial planes of local folds and associated lineation are known from several ranges, such as the East Potrillo Mountains and Sierra Alta, usually over areas of a few square kilometers and nowhere for the full extent of a range. Cleavage unrelated to axial planes of local folds also occurs in the Chiricahua Mountains in the metamorphosed Paleozoic rocks previously described and in the hinterland zone. This unrelated cleavage probably reflects rock slippage under a substantial confining pressure, a pressure distinctly higher than that of adjacent terranes. Such a situation may develop on the stoss side of suitably offset segments of northwest-trending faults northeast of which basement rocks were raised.

A third measure of the thickness of cover is based on an inference of the confining pressure at which muscovite is in equilibrium with other mineral components of the 2-mica garnet-bearing granite stocks that are associated with the gneiss-cored domes. This pressure is given as equivalent to a load of 9.6 km by Davis and others (1980) for the Whipple Mountains situation, and is further used in the White Tank Mountains by Brittingham (1985). Such granite, with the tectonic fabric of the gneiss-cored domes, occurs as far east as the Rincon Mountains of eastern Arizona, and without that fabric such granite occurs as far east as the Texas Canyon stock of the Little Dragoon Mountains (Cooper and Silver, 1964). As explained in the section "Structural Features of the Northeast Margin," this eastern prong of gneiss-cored domes is believed to be the result of a gently inclined contact between western intermediate and hinterland zones plus post-orogenic southward-tilting and differential erosion that exposed these deeper rocks and the domes. The thickness of cover along this part of the transect may, therefore, have been as much as 9.6 km.

Along much of the transect, then, the cover above the basal décollement or base of strongly deformed rock was 8–10 km thick. To the east in the Sierra Juárez that thickness tapers off to only about 2–3 km and then presumably shallows more rapidly to the frontal line. To the west in the Piute Mountains area of California the calculated thickness of cover increases to 12–15 km, and, presumably, the thickness of cover was even greater and increased even more rapidly near the tectonic core.

For comparison, figure 35 also shows the approximate thickness of cover over the detachment faults. That cover thickness is taken by most workers to be the thickness of the suite of mid-Tertiary rocks, mostly strongly tilted sedimentary and volcanic rocks above the detachment fault, commonly 1–2 km (fig. 35, detachment fault line, segment BC) at least as far east as Phoenix. To the far west the thickness of cover overlying the detachment fault rapidly approaches zero where the fault comes to the surface and a pull-apart zone may be found (segment AB and, according to Hamilton (1982), ABD and possibly extending down to the level of 10 km). In sum, then, the environment of deformation of the detachment faulting is generally viewed as distinct and separate from those of the Cordilleran orogeny; clearly, two times and two stress systems were involved.

The reconstructed thickness of cover, however, does not imply that the entire region of the Texas-California transect was buried to this depth of 8–10 km throughout the orogenic period; rather, these were the likely conditions as the stage of peak deformation migrated through a specific site. Consider the following conditions in, for instance, the Tucson area about peak deformation time. To the west, the cover was already being thinned through the late-stage uplift and erosion of the peak(±) orogenic volcanic rocks and of the underlying older foreland basin sedimentary rocks. This eroded material may in large part be transferred from the uplifted area to a new foreland basin which at this stage of development lay to the east of Tucson, where the cover thickness was still growing. Through such a process of cannibalization of cover rocks the total volumes represented by an 8- to 10-km-thick cover would have been considerably less than that implied by a static, time-uniform, orogenic model. That such sedimentary transfer, or foreland basin cannibalization occurred, has already been inferred in the section on foreland basin deposits.

CONCLUSIONS

Observations and inferences from this study indicate that the Cordilleran orogenic belt is continuous across the Texas-California transect. The original bases for the hypothesis remain valid, but some modifications are required because of the large area from which new features and conditions are seen or inferred. Objections that had been presented to this hypothesis have been overcome. In the process of overcoming them, further new observations and inferences have been assembled which strengthen the support for the regional hypothesis. These deserve further scrutiny, of course, and at the least they may guide the way to a still better hypothesis covering the deformation of rocks in so critical an area through a span of about 100 m.y. The final conclusion to be offered comprises an interpretation of the development of the region in which the main emphasis will be on illustrating the mobility of the rocks under stress and the mode of stress propagation.

ORIGINAL HYPOTHESIS, MODIFICATION, AND NEW SUPPORT

The description of the styles of deformation of Cretaceous and early Tertiary age and the inferences derived from key tectonic features and conditions provided the basis for a synthesis of the tectonic development of the region. This tectonic development lends broad support for the hypothesis that the Cordilleran orogenic belt does extend between El Paso and Las Vegas and is the product of a large northeast-southwest-oriented compressive stress similar to the stress applied to other parts of the belt. Some unusual local features also generated complications. The presence of unanswered problems indicates that my understanding of the tectonic evolution of this segment of the orogenic belt will be further modified through additional studies.

The original presentation of the hypothesis of a through-going Cordilleran belt was based on a study of only a part of southeastern Arizona and consequently developed a relatively local perspective (Drewes, 1976, 1978, 1981a). Parts of this hypothesis also appear in Drewes (1988; 1989). Original support for regional tectonism was obtained from (1) observations that tectonic transport was large—a minimum of 16–32 km, with stronger field support for the shorter minimum—(2) the recognition that the style, age, and direction of tectonic transport are uniform, and (3) the regional consideration that the kind and intensity of deformation occurring north of Las Vegas and south of El Paso through the convergence of the northeast Pacific global plates with the North American plate is not likely to have a gap between El Paso and Las Vegas.

Item (2) of the original hypothesis, concerning uniformity of factors, is modified and includes the new supporting evidence. Seen across the entire belt, I now infer that either uniformity or **systematically changing** factors support the hypothesis of a through-going orogenic belt. Deformation, for example, did not occur across the entire belt at one time; rather, it progressed systematically, from older in the proximal part to younger in the distal part of the belt. Likewise, a measure of uniformity of direction of tectonic transport should be modified in view of obliquely converging masses, development of backthrusts, and responses to several orogenic phases generated in various regions or to local stress fields that modify a regional field. Similarly, the concept of a uniform amount of tectonic transport must be modified to a systematically changing amount, from larger toward an orogenic core to smaller near the frontal line. Thus, the site at which the amount of tectonic transport in the orogen is measured is as vital as the actual amount measured.

Several items have been raised that needed to be addressed to properly support or modify my hypothesis. For example, the first point raised was that the overriding plates were too thin to transmit compressional stress, though no thickness values for either the Arizona case or for an adequately strong plate system were given. This response prompted the study reported on in this report on the available thickness of cover across the orogenic belt (fig. 35). During the orogeny the thickness of cover available across southern New Mexico and Arizona was probably substantial—about 8 km—and in the range of thickness known or implied along other transects of the orogenic belt.

The second point raised was about the possible inadequate strength of the rocks to transmit stress. This point must be answered with due consideration of, among other factors, the duration of application of stress. Under a given set of conditions, such as thickness of cover, the greater the stress duration, the more favorable is the situation for transmission of stress over a large distance. Figure 31 illustrates the duration of two stress systems, the extensional one for the mid- and late-Tertiary event and the compressional Cordilleran orogenic even. The compressional event was clearly long lived. Therefore, the duration factor of compressional stress application is probably significant in aiding stress transmission over a long distance. Indeed, the mechanism for generating a broad field of extensional stress in mid- and late-Tertiary time deserves careful consideration. Why doesn't such extensional stress first develop a line of weakness, which thereafter continues to "neck" (following mechanical principles) and thereby develop a rift? One may well conclude that the development of the Cordilleran tectonics is less enigmatitic than the development of the Tertiary extensional event.

A third point raised was that a foreland basin, so commonly associated with orogenic belts, was absent in southern New Mexico and Arizona, and this absence denied the hypotheses of a through-going orogenic belt. The late pre-orogenic and early orogenic sedimentary record of the region provides evidence for a series of eastward-shifting and younging depocenters. These depocenters, possibly part of one long-lived foreland basin analog, contain much detritus that apparently was derived from older and westward-lying parts of the orogenic belt. Thus, instead of becoming an argument against the through-going belt hypothesis, the geologic record provides another kind of support for it.

A fourth point to consider consists of an alternative hypothesis that was offered to cover the Late Cretaceous and early Tertiary development of part of the original study area. In this hypothesis Davis (1979) suggested that a major block, trending about from Bisbee to Tucson, Arizona, and about three ranges (of the present topography) wide, was raised between a pair of normal or reverse faults. Some of the low-angle faults flanking the postulated raised block are related to flanking reverse faults and others are proposed to be related to the tectonic denudation of the raised block. While this alternative hypothesis explains some local field relations, it fails to explain many others, and it is contradicted by still other facts. This alternative hypothesis does not indicate which specific faults are remnants of the border faults of the raised block; thus, this key feature cannot be tested. Low-angle faults occur not only along the flanks of the proposed structurally controlled highland, but also within it and far away from it, and thus cannot be only related to it. Some of the thickest and most complete sections of Cretaceous rocks, youngest of the pre-orogenic deposits, occur on the so-called raised block, precisely where post-orogenic erosion would be expected to have removed them, as well as much of the underlying Paleozoic rocks. The thickest sections of the Bisbee Group, for instance, occur in the Mule, Whetstone, and Huachuca Mountains (Hayes, 1970). Furthermore, so large an upland would be expected to have been flanked by aprons of coarse alluvium laden with locally derived Cretaceous and Paleozoic detritus, but remnants of such deposits are not known. In effect then, there is little support for this alternative tectonic hypothesis, and there is considerable evidence against it.

Returning to the hypothesis of a through-going Cordilleran orogenic belt, in addition to the modifications to the older supporting evidence and rationale, new support exists from the records of sedimentation and magmatism. The inference of the eastward shifting and simultaneously younging series of depocenters shows that the foreland basin characterizing orogenic belts is present along this transect, too. At least some of these depocenters axes were tilted southeastward during and after deposition. They are subaerial to the northwest and marine or lacustrine to the southeast. Therefore, much of the detritus was bypassing the northwestern parts of the foreland basins, making them seem smaller to the northwest than the southeast. Earlier formed western parts of the orogenic belt were probably cannibalized to provide much of the detritus for the next younger depocenter to the east. This part of the development not only explains the origin of the voluminous black or gray shale and siltstone of the basin deposits, but also it explains the fragmentary condition of the andesitic cover in the older parts of the orogenic belt. Erosion that follows so promptly and cuts so deeply at the sites of the older depocenters also explains the earlier development and subsequent earlier exposure of metamorphic terranes in the proximal part of the orogenic belt. Ultimately, too, such sedimentary recycling greatly decreases the mass of deposits required to help propagate the orogenic stress. In other words, a 10-km-thick cover need not have been present across the width of the mobile belt at one time, but it was present in specific parts of the belt at, or just before the time of peak deformation. As the "bulge" of maximum thickness migrated across the belt, so did the condition of maximum strength. In the distal parts of the orogenic belt, where the effects of orogenic development dwindled, the foreland depocenters were smaller and the cannibalization process declined. These principles of mobile conditions of development of this southern part of the Cordilleran orogenic belt are illustrated in figure 36.

Finally, the partial explanation of the complexities along the northeast margin of the deformed terrane provide some permissive support for the regional hypothesis, or at least it anticipates a potential objection to this hypothesis. Clearly, the enigma of the relationship between the ductily deformed terrane and the autochthonous Colorado Plateau block remains incompletely resolved.

DEVELOPMENT OF THE OROGENIC BELT

The development of an orogenic belt of the Cordilleran kind, as shown in figure 36, is devised from the Texas-California transect, but it is generalized and not site specific or time specific. The vertical scale is exaggerated an arbitrary amount. Therefore, spatial and temporal control lines are simply designated, respectively, as vertical lines 0 through 6 and as profiles of times T+0 through T+6, with T+0 being the beginning. Bedding lines are added to show fold styles and are not rigidly controlled datum planes. Changes in vent or intrusive size and shape are arbitrary and are meant to show their possible growth; to avoid clutter on succeeding profiles those intrusives no longer being discussed are removed. Two reference points, X and Y, arbitrarily placed 150 km apart at the same tectonic level, will help to follow some aspects of the tectonic development that probably takes place along any transect of the Cordilleran orogen.

At an initial time (**T+0**) the highly simplified conditions are shown to involve a cratonward (northeastward) thinning sedimentary rock sequence over a crystalline basement (fig. 36). Compressional stress (shown by arrows) is applied to the rocks as global plates converge, and surface rocks deform while subduction moves most of the left-hand plate, relatively, beneath the right-hand plate.

At time **T+1** (about 10 m.y. after initial time (T+0)), cover rocks to the southwest have been folded disharmonically with respect to the basement rocks, a, and the cover rocks have thereby been thickened. In part, this thickening has generated a highland; in part, the base of the highland has pushed the crust down (arrows above site a) and consequently point X has gone down relative to point Y. Deposits shed off the highland accumulate at its foot in a basin, (c) whose base also sinks (open arrow), at least in part, under the additional load. This is a common association of processes at an initial stage of orogenic development.

At time **T+2**, the tectonism and sedimentation have continued. Axial planes of some folds are cut by reverse faults, b, or thrust faults, so resembling structures of a foreland tectonic zone. The basal thrust fault and disharmonic folds, at a, have propagated farther northeast, and the older foreland basin deposits are partly destroyed through uplift and erosion. A new depocenter c develops still farther northeast. Point X is closer to Y but probably still below it, with both reference points probably below their initial level.

At time **T+3**, the same incremental changes occur. At the position of the 100 km control line, conditions resemble the eastern intermediate zone near the surface and the western intermediate zone at depth (but only to be exposed later through erosion). Basement rocks are thrust faulted in the internal part of the belt, a; thrust faults propagate northeastward, b; the foreland basin depocenter shifts northeastward again, c; the older foreland basin deposits are deformed, d; and late orogenic andesitic-dacitic volcanism covers the older deformed terrane to the southwest, e. The reference points X and

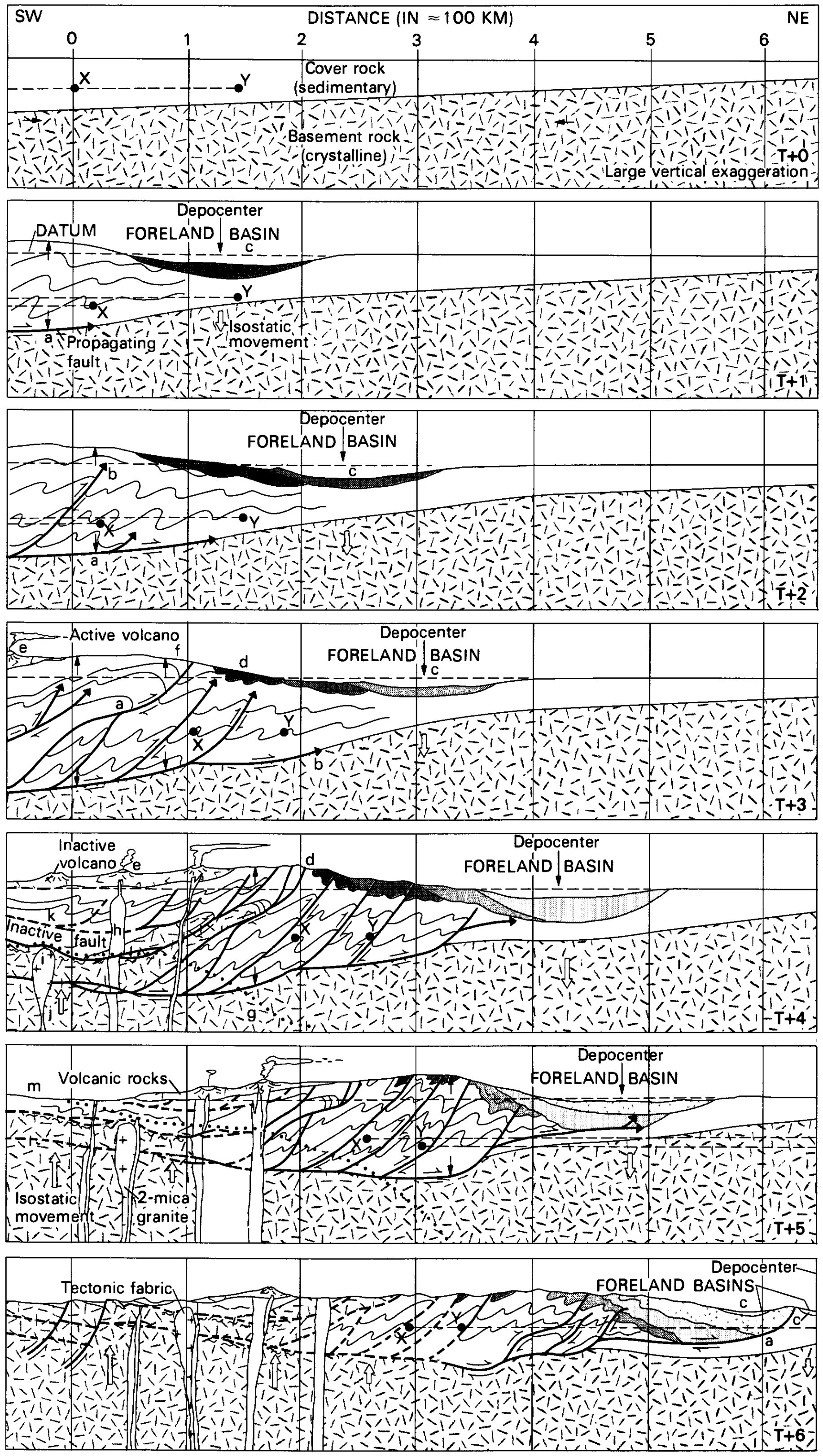


FIGURE 36.—Schematic tectonic development along the Texas-California transect of the Cordilleran orogen, shown in a series of profiles, T + 0 to T + 6, each following profile having occurred roughly 10 m.y. after the previous one. Bedding lines are shown only to illustrate styles of deformation and fault offsets and are not horizon specific or consistent throughout the sequence. See text for explanation of letters in each profile. Large vertical exaggeration; restored lengths along folds not true scale.

Y continue to converge; because of uplift and erosion of the central part of the orogen, area f. They may now lie on about the same level once more, and they have been transported eastward. The orogeny is in full bloom.

At time **T+4**, the processes already described continue but have shifted their loci still farther east. In the southwestern, more interior part of the orogen, magmas rise from an underlying source in the lower part of the crust or from the subducted edge of the western plate; the geothermal gradient, line g, rises to foster more plastic deformation and to facilitate regional metamorphism, acting as a catalyst for penetrative deformation. Early plutons, h, rise to levels where movement along thrust faults is still in progress; slightly later plutons cut across such faults after movement ceased or in the ductile zone 2-mica granite bodies may interact with the last fault movement, i. Some magmas breach the surface, adding a widespread cover, first of more andesitic volcanic rocks and then of more rhyolitic ones, e. Inasmuch as they form the cap on the upland, these volcanics provide much detritus for the depocenter currently being filled. With the cessation of magmatism to the west, erosion gradually reduces the thickness of the deformed pile, and consequently the crust begins to rise isostatically, j. Faults at shallow levels, in the hinterland, k, are the first to become inactive.

Continuing to time **T+5**, and focusing on the southwestern part of the orogen, the hinterland style of deformation is gradually exposed through deep erosion. Magmatism shifts eastward, and the volcanic record at the surface is largely destroyed, to give surface access to the plutonic and metamorphic record, m. Reference points X and Y have reversed their relative vertical position, have moved even closer together, and have been transported still farther east.

Finally, at time **T+6**, the main part of the orogen is no longer being compressed, but it continues to be uplifted and, consequently, is deeply eroded. Toward the foreland of this stage of development, at distance guide line 6, final minor responses to a dwindling compressive stress are still being made by faults, a. The depocenter system is dying out in a series of small cratonic basins, c. In the dormant western part of the orogen new tectonic regimes, such as an extensional one, may be initiated concurrently with the last compressive responses to the east. The orogeny has run its course.

The observations on the two reference points (X and Y) offers an insight to some of the complexities of movement within the orogenic belt. The site of the points were taken to lie within the same major plate; other site selections probably would result in more pronounced changes in their relative position through the course of orogenic development. In the present example, the following motions are deduced: (1) the points were transported toward the foreland; (2) they were pushed closer together; and (3) they were shifted vertically, at times independently and diversely, and perhaps at times in unison. Additionally, (4) the reference points were bypassed by the propagation of deformation of the orogenic belt toward the foreland; and (5) possibly they became tectonically dormant while the foreland deformation was still in progress.

As a final summary, an attempt is made in figures 37–40 to show the paleogeographic development of the region, using palinspastic reconstruction procedures. Of the assumptions that must be made for the reconstructions, the time of deformation, and direction of tectonic transport are better controlled than is the distance of tectonic transport. During the main phase of deformation, over most of the region a N. 70° E. transport direction is applied; only in the western reaches was the main movement direction more nearly due east; the more northerly movement direction of the Sierra Juárez area affects mainly the terranes south of the study area and involves a minor amount (10–20 km) of movement relative to what can be shown at the scale of the figures. At all stages represented by the reconstruction, the Colorado Plateau block is assumed to be stable and the other terrane mobile, although, in fact, both were mobile and converging. The post-orogenic extension of the Rio Grande rift and eastern Basin and Range province are likely to be minor—less than 10 km, which would account for rotation of blocks—and thus for them no adjustment was made. The effects of post-orogenic strike-slip movement along the lower reaches of the Colorado River and the Gulf of California fall largely outside the region of the reconstructions, and those fault splays that may enter the region are parallel to the major inferred structures and thus do not offset them. The effect of the post-orogenic extensional spreading of the western part of region is difficult to assess, partly because of the uncertainty of its magnitude and partly because of the diversity of opinion on whether shear and translation are involved or only ductile distensions without translation of surface features. For the present reconstruction this factor is set aside. The reader should note, however, that if the region was extended 100 percent, the magnitude of Cordilleran orogenic shortening is reduced to 50 percent. Geographic features, such as state borders and cities, are shown in their present sites for reference.

At the geographic scale of figures 37–40 and at the large incremental scale of time represented, only general features of paleogeography can be shown. During the oldest span of time (fig. 37), largely Early Cretaceous, but extending back into the Late Jurassic, a northeast-trending belt of compressive deformation was developed in parts of southern Nevada, southeastern California, and perhaps also Baja California. An initial plate, with southeast transport developed in the northwestern part of the region. Between it and the slightly younger main plates that moved eastward, the McCoy Mountains basin received at least 3,000 m of clastic and volcaniclastic

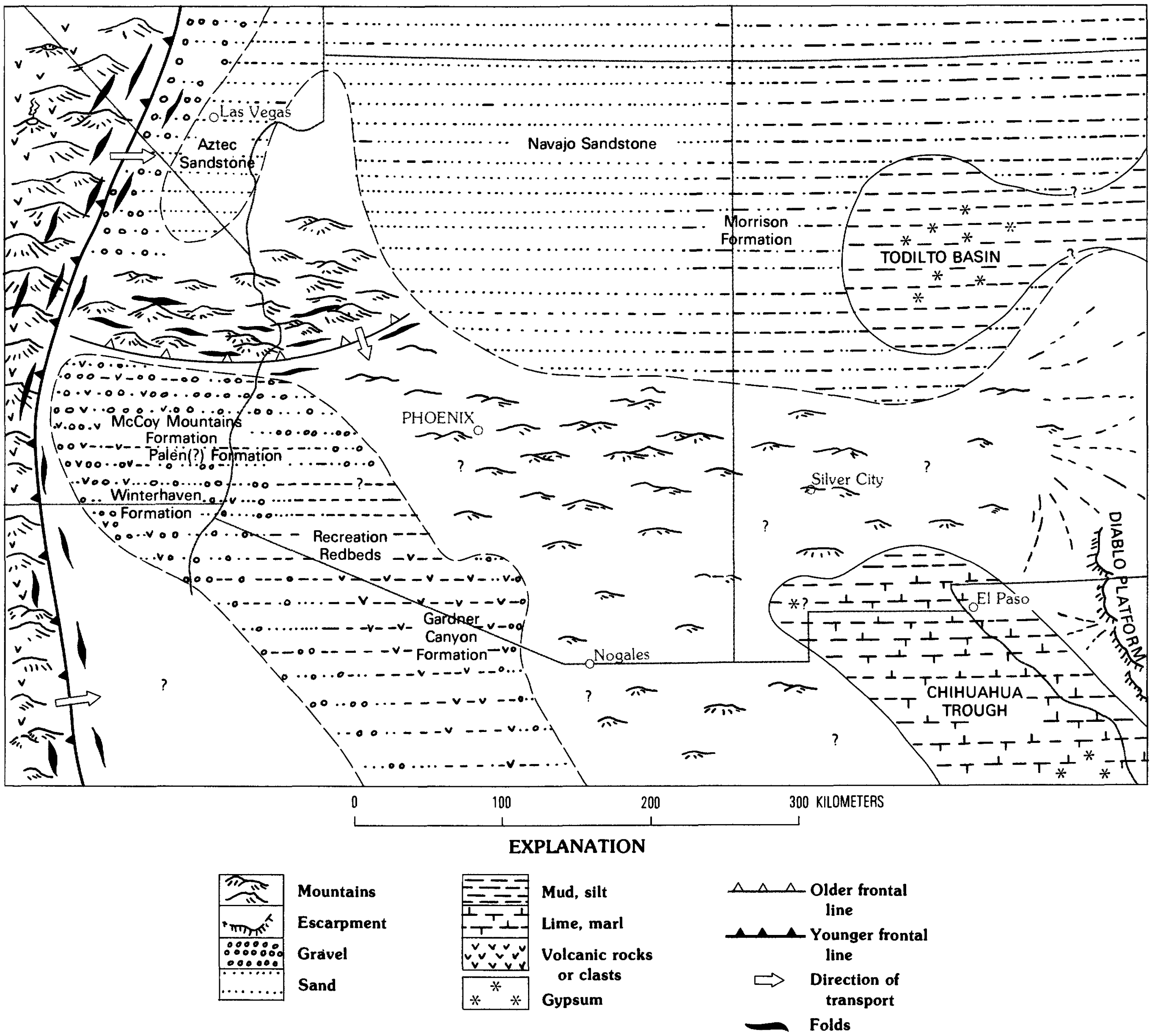


FIGURE 37.—Palinspastically reconstructed paleogeographic map, first stage (Late Jurassic and early Early Cretaceous, or 150–120 m.y. ago). Present geographic features shown for reference. Queries used where inferences are made with least control.

rocks. Farther southeast this, or similar basins, also received voluminous sedimentary and volcanic deposits, that all together may be viewed as an early stage of development of an orogenic foreland basin. Other basins of deposition lay farther east and northeast, apparently separated from each other and from the McCoy Mountains basin by emergent areas. Some of these basins, such as the Todilto basin of central New Mexico and the deposits of the Chihuahua trough, are noted for gypsum deposits. Between the Todilto basin and the orogenic front near Las Vegas, eolian sands and fluvial muds of the Aztec-Navajo Sandstones and the Morrison Formation, respectively, were widely deposited, reflecting relatively low, flat, arid conditions, perhaps in the lee of a mountainous terrane to the west that could be the topographic expression of the early stage of the Cordilleran orogenic belt.

At a second stage of development (fig. 38), the orogenic belt east of Las Vegas encountered the semistable mass of the Colorado Plateau block and started to wrap around its southwestern flank. Likely, the southwestern edge of the block was raised and the central and eastern parts were lowered to receive the westernmost incursions of the Cretaceous sea, whose

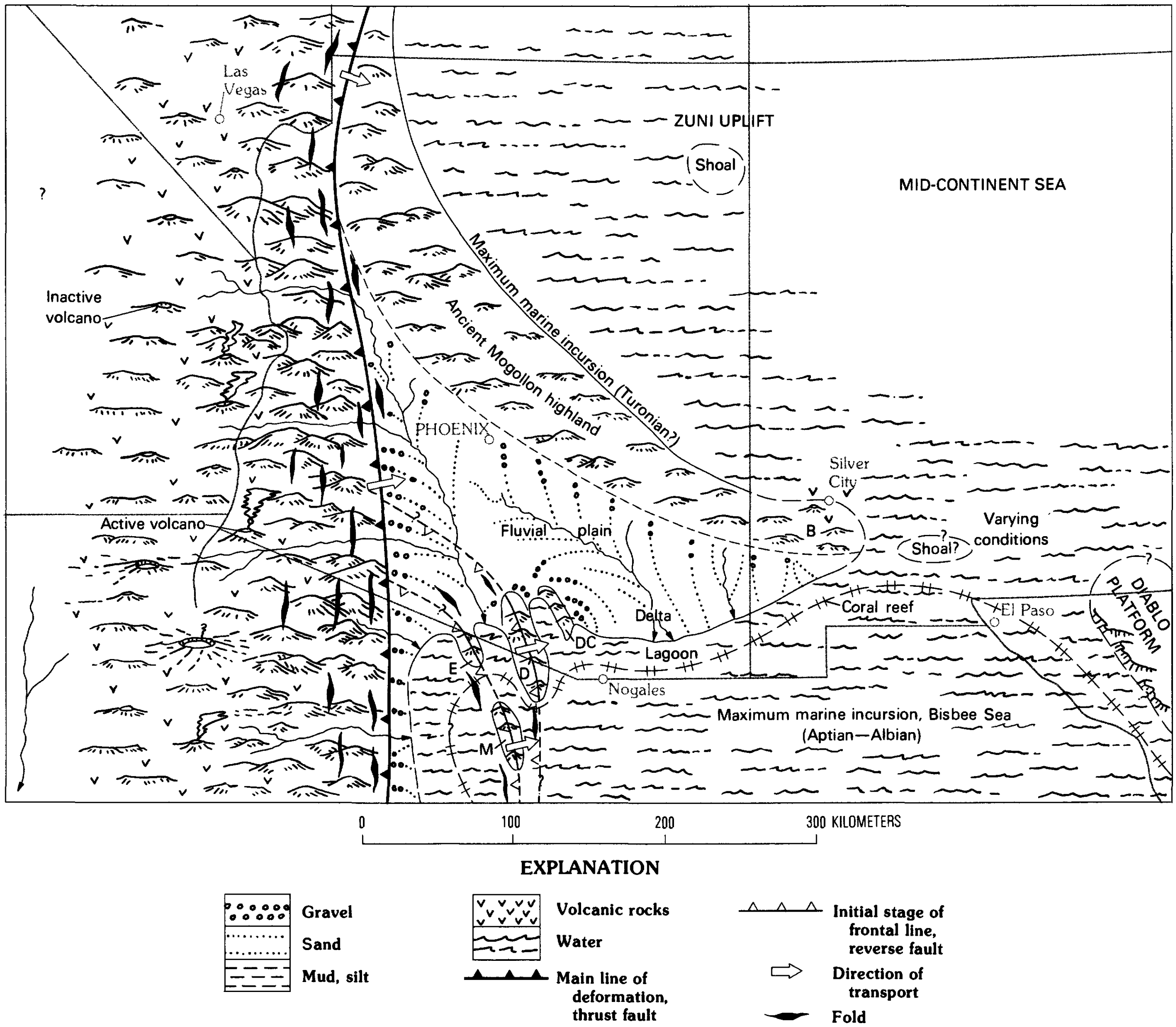


FIGURE 38.—Palinspastically reconstructed paleogeographic map, second stage (late Early Cretaceous and early Late Cretaceous, or 120–85 m.y. ago). Present geographic features shown for reference. Sites of present features: B, Burro uplift; D, Dragoon Mountains; M, Mule Mountains; E, Empire Mountains; and DC, Dos Cabezas Mountains.

progression is described by Cobban and Hook (1984). To the south of the raised edge of the block, which eventually appears to have developed, peninsula-like, southeast to the Burro uplift of southwestern New Mexico, the Bisbee sea extended northwestward across Chihuahua, northeastern Sonora, and into southeastern Arizona. A fluvial basin extended northeastward of the Bisbee sea, much as envisioned by Hayes (1970). Near, and west of the present site of Nogales, isolated sharp ridges were developed from which coarse detritus was shed locally to as late as perhaps early(?) Albian time (105–120 m.y. ago). The northern extension of the reefal development is shown wrapping around the peninsula-like southeastern end of one of these ridges, which represents the Dragoon Mountains terrane. That reef line may also have extended island-like north around another such ridge, perhaps an island in the Bisbee sea, that later became the terrane of the southern Mule Mountains. Other local ridges may mark the Empire Mountains and northwestern part of the Dos Cabezas Mountains terranes.

During this second stage of development, the main orogenic front lay to the west of the present site of Phoenix, but isolated sharp ridges were forced up east of the deformation front in a paleo-foreland zone. These

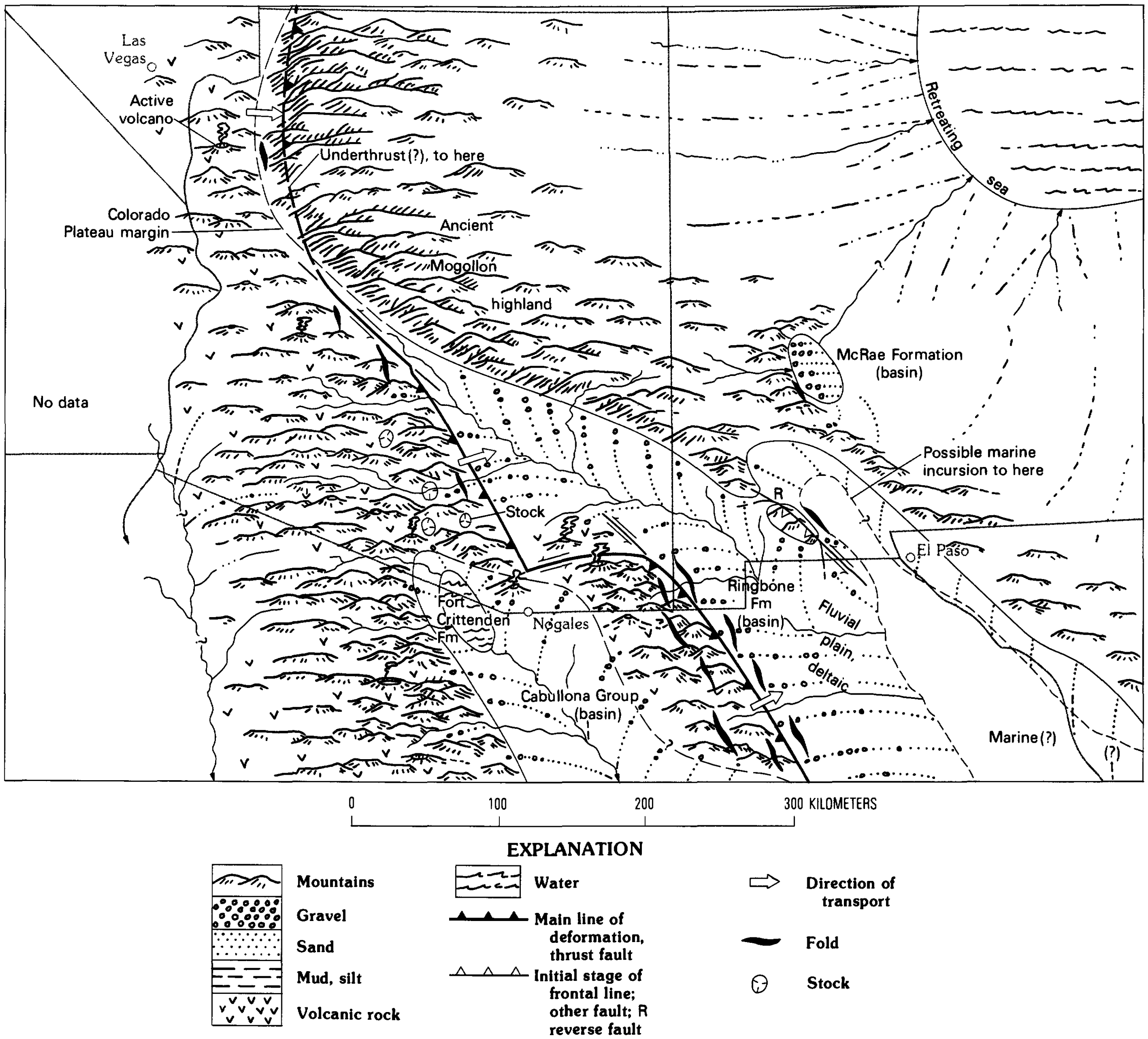


FIGURE 39.—Palinspastically reconstructed paleogeographic map, third stage (late Late Cretaceous and early Paleocene, 85–58 m.y. ago). Present geographic features shown for reference.

ridges were raised along local reverse faults resembling the fault of the southern San Andres Mountains, of the last stage of development. Alternatively, these ridges may have been raised along the most distal parts of sled-runner faults that propagated eastward beneath the western part of the Bisbee foreland basin and then upward, perhaps where the added friction of offsets across the pre-orogenic northwest-trending basement faults was encountered. Detritus from the andesitic terrane capping the interior part of the western orogenic terrane was carried first eastward to the fluvial basin and then ever more southeastward or southward into the Bisbee sea. Minor detritus came from carbonate rock or granite rock terranes appearing locally at the edges of the longer lived and more extensive orogenic upland, the ancient Mogollon highland, and some shorter lived, more local sharp ridges within the foreland basin.

Continuing to a third stage of development (fig. 39), of late Late Cretaceous and early Paleocene age, the orogenic belt propagated forward across southern Arizona. Some of the accommodation around the Colorado Plateau salient (fig. 2; at stage of development of figs. 38 and 39

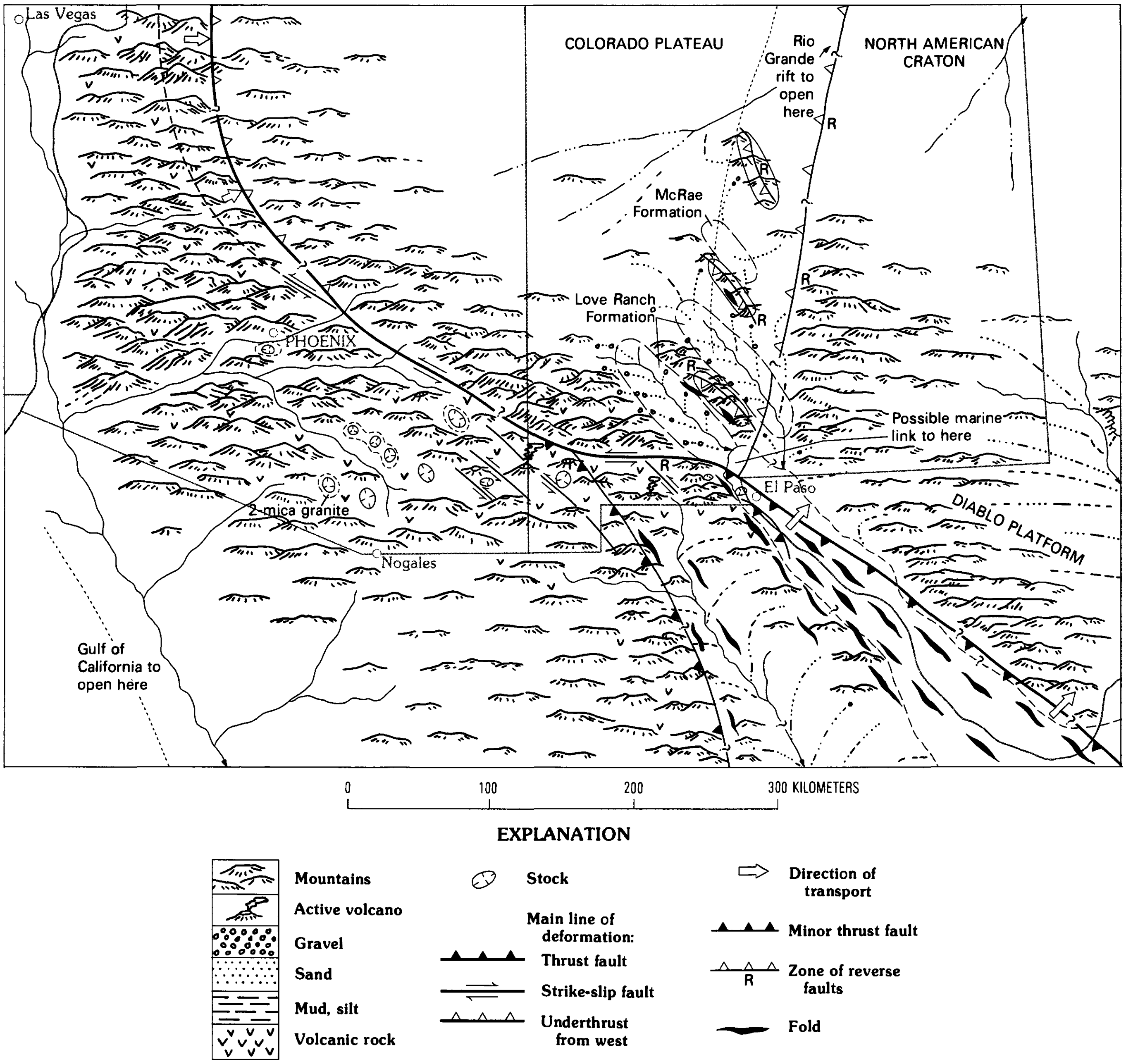


FIGURE 40.—Palinspastically reconstructed paleogeographic map, fourth stage (late Paleocene to early Eocene, 58–50 m.y. ago). Present geographic features shown for reference.

shown as western part of ancient Mogollon highland) took place by way of adjustments along strike-slip faults, and perhaps some by way of underthrusting beneath the edge of the salient. The Cabullona and Ringbone basin (or basins?) were the foreland receptacles of fluvial and lacustrine deposits. Their upward-coarsening characteristic suggests that they were strongly influenced—indeed, essentially overwhelmed—by the encroaching orogenic conditions. The Colorado Plateau margin remained a positive element, which spread eastward, to drive back the margin of the continental sea. The orogenic uplands of southern Arizona likewise propagated eastward. Magmatic activity increased; andesitic rocks spewed out at the surface, and granodioritic stocks were emplaced beneath it. The flood of effusive rocks may have produced a topographic barrier between two parts of a shrinking or southeastward-shifting foreland basin of southeastern Arizona. The western (Cabullona) basin may thus have been internal in the orogenic belt. This split may also have been aided by the separation of the mobile belt into a southeastern and a northwestern set of thrust plates.

During the fourth stage (fig. 40) of orogenic development, tectonism was waning, and uplift and erosion spread eastward and increased in tempo. The last tectonic movements—perhaps in amounts of 1–10 km—took place along the frontal zone along the northeastern part of Chihuahua, at scattered foreland zone sites in south-central New Mexico, and possibly beneath the western margin of the Colorado Plateau salient. (See "Structural Features of the Northeast Margin.") Monoclinal warps of the plateau block were possibly part of this stage of development. Finally, a separation between the main part of the North American craton and the slightly thinner Colorado Plateau block may have initiated north-trending crustal rupture through central New Mexico. At some sites the early phases of the 2-mica granites were emplaced, if they had not already appeared in the western terranes by this time. However, deep cover still overlay these stocks, and in the eastern and western intermediate zones some movement along faults still took place. Deep-seated interaction between stock emplacement and residual faulting formed the asymmetrically distributed intensities of deformation around some of the domes. To the northeast two basins of deposition received the Love Ranch deposits. Another(?) one northwest of El Paso (based only on data from an exploratory well) indicates that marine conditions of this age possibly were present. Currently (July, 1989), a reassessment of the paleontologic findings is in progress and, pending the results, a marine arm is not shown in figure 4. Andesitic deposits were extruded and granitic stocks intruded ever farther eastward, probably in much diminished volume (and variety?). Meanwhile, the tectonic and magmatic action had already ceased in southwestern Arizona and deep erosion followed the steady uplift of that region.

Following the cessation of the orogenic activity, during late Eocene and early Oligocene time, uplift and erosion took place. In south-central New Mexico, where the northern edge of the deformed terrane was only 3–4 km thick, erosion was deep enough to remove it, but in southwestern New Mexico and southeastern Arizona where the deformed terrane was 8 km thick, only the upper part of the terrane was eroded. Magmatism was restricted mainly to emplacement of late phases of the 2-mica granites, probably into rocks that now had left the more ductile level and entered the more brittle level, thus initiating development of topographic dome, as well as the local arching, tectonic denudation, and erosion over some domes. Gradually, a new tectonic regime characterized by extension spread across the region of the orogenic belt.

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