The Arizona Geological Society and the University of Arizona

1981 Symposium on

"RELATIONS OF TECTONICS TO ORE DEPOSITS

IN THE SOUTHERN CORDILLERA"

AGS/UA Mine Tour and Field Trip data for

TRIP #3

"Mylonitic Tectonites and Detachment Faults Southeastern Arizona"

March 16-18

Leaders: G. Davis, J. Hardy, S. Lingrey,

and P. Trever (U of A),

C. Thorman (USGS)

Field Trip Coordinator: Phil Gerla (GMRC).

ARIZONA GEOLOGICAL SOCIETY
P.O. BOX 40952, UNIVERSITY STATION
TUCSON, ARIZONA 85719

TO: Registrants for AGS Field Trip #3, Mylonitic Tectonites and Detachment Faults, Southeastern Arizona

The central theme of the symposium is tectonics and its relationship to ore deposits or absence thereof; however, as for specific trip content, this must necessarily remain in the able hands of each respective field trip leader. This particular excursion will take you into the metamorphic core complexes of the Graham and Rincon Mountains of southeastern Arizona. Although the Tertiary metamorphic core complexes of the southern Cordillera have repeatedly proven notoriously barren framework elements, modest quantities of base, precious, and lithophile metals have been produced primarily from décollement-zone microbreccia or adjoining fractured/brecciated footwall and hangingwall host rock. This empirical relatiosnhip when coupled with the intrigue associated with this newest element of the southern Cordillera tectonic framework, prompted the Society to include this trip for those registrants wanting an excellent overview as well as local definition of mylonite tectonics, detachment and concomittant listric(?) normal faulting, and their relavence to the metals explorationist.

Our appreciation is extended to Dr. George Davis and students (U of A) and Chuck Thorman (USGS) for agreeing to lead this trip and providing us with the excellent handout materials included herein. The continual efforts of the trip's logistic coordinator, Mr. Phil Gerla from Gulf Mineral Resources Co., is similarly appreciated.

Again, welcome to the American Southwest, happy rock knockin', and here's hoping that each of you will take the time to contribute to the trip's success by participating in many timely and stimulating discussions of the rocks at hand on their terms.

Sincerely,

Tom L. Heidrick
AGS Field Trip Chairman

TLH/gr

ITINERARY

3. MYLONITIC TECTONITES AND DETACHMENT FAULTS, SOUTHEASTERN ARIZONA

General Leader: George Davis (626-2365)
Coordinator: Phil Gerla (882-4030)

March 16, 1981

3:30PM	Leave U of A for Safford
7:00PM	AGS dinner banquet and trip overview (Sirloin Stockade in Safford)
Lodgings:	Desert Inn, Safford

March 17, 1981

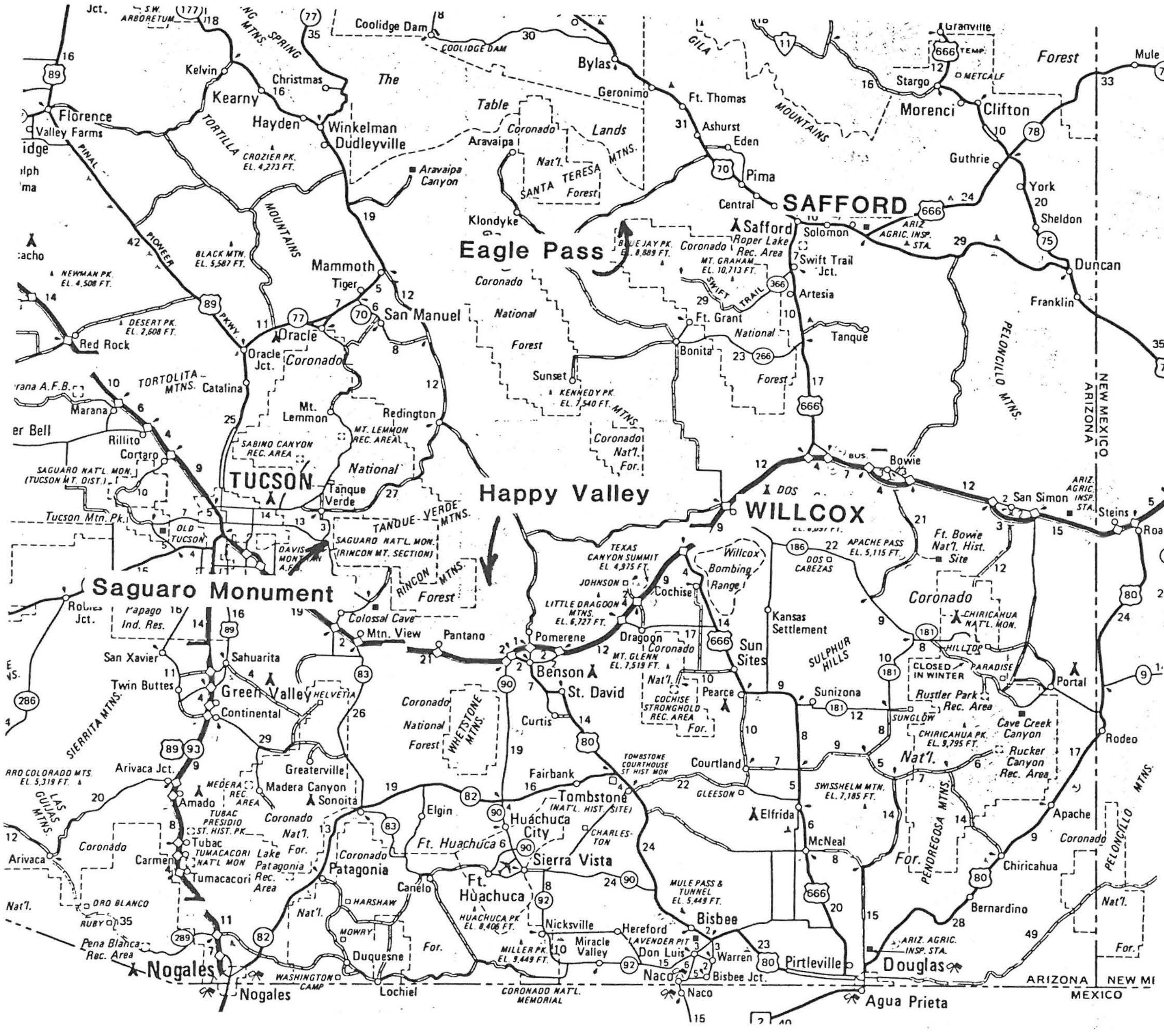
8:00AM	Leave Safford
morning	Stops in the northeastern Graham Mountains
12:00Noon	AGS lunch at Big Spring, near Eagle Pass
afternoon	Stops in the Eagle Pass area
6:00PM (approx.)	Arrive at Willcox
7:00PM	AGS dinner banquet and trip overview (Plaza Inn)
Lodgings:	Plaza Inn, Willcox

March 18, 1981

6:30AM	Depart Willcox for Happy Valley (eastern Rincon Mountains)
morning	Stops in Happy Valley
12:00Noon	AGS lunch before leaving Happy Valley
afternoon	Stop near Colossal Cave Stops in Saguaro National Monument East
5:30-6:00PM	Arrive at U of A

Drivers:
Dr. G. Davis
Barbara Phillips
Sandy Ballard
Steve Lingrey
Paula Trever
Phil Gerla-Ralph Rogers (support vehicle)

KEY MAP FOR FIELD TRIP 3



The Arizona Geological Society would like to take this opportunity to acknowledge the financial support provided by Southwestern Assayers & Chemists, Inc., who underwrote most of the expenses related to our bringing together, printing, and binding these handout materials for AGS field trip #3 registrants.

Geological Society of America
Memoir 153
1980

Structural characteristics of metamorphic core complexes, southern Arizona

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ABSTRACT

Metamorphic core complexes in southern Arizona may be subdivided into four elements: core, metamorphic carapace, decollement, and cover. Cores consist chiefly of mylonitic augen gneiss that is, for the most part, derived from Precambrian and Phanerozoic plutonic rocks of granitic composition. Foliation in the cores is characteristically low dipping and commonly defines large upright, doubly plunging foliation arches. The mylonitic augen gneisses everywhere display low-dipping mineral lineation of a cataclastic nature. Lineation within any given metamorphic core complex is generally remarkably systematic in orientation. In southeastern Arizona the lineation trends northeast to east-northeast; in south-central Arizona, it trends north-south. Ductile normal faults are locally abundant in the core rocks and always are oriented perpendicular to lineation. Core rocks in places are clearly transitional laterally or downward into undeformed protolith.

The metamorphic carapace consists of penetratively deformed younger Precambrian and Phanerozoic strata metamorphosed to greenschist-amphibolite grade. It forms a crudely tabular layer that locally overlies the crystalline rocks of the core. The contact is commonly so tight that the rocks of the carapace appear to be plated onto the crystalline rocks of the core. Within the metamorphic carapace, overturned to recumbent folds are ubiquitous, transposition is rampant, and boudins and pinch-and-swell features are commonplace. In spite of spectacular deformation, individual formations within the metamorphic carapace are arranged in normal stratigraphic order.

A decollement, marked by brittle low-angle faulting and shearing, separates carapace and cover or, where carapace is absent, core and cover. The surface thus separates rocks of remarkably contrasting deformational styles. Where the surface "caps" core rocks, a decollement zone is formed beneath the decollement and consists of a distinctive crudely tabular zone of crushed and granulated but strongly indurated mylonite, mylonitic gneiss, microbreccia, and chlorite breccia. Striking "younger-on-older" fault relations characterize the decollement. Decollements in this area typically separate orthogneisses, which were derived in part from Precambrian rocks, from unmetamorphosed upper Paleozoic, Mesozoic, or Tertiary strata. Overturned asymmetric folds, detached isoclinal folds, and unbroken cascades of recumbent folds are abundant in many of the cover sheets.

The metamorphic core complexes in southern Arizona are interpreted to be products of high-temperature extensional deformation, regional in extent, superseded by moderately ductile to moderately brittle tectonic denudation. Rocks in the augen gneissic core and metamorphic carapace were affected by profound ductile through brittle extension and flow in the direction of mineral lineation. Evolution of the decollement zones largely postdated the development of foliation and lineation.

Mechanics of strain are interpreted in the context of megaboudinage. Cores are pictured as parts of megaboudins imposed on heterogeneous crustal rocks. Profound thinning of younger Precambrian and Paleozoic sediments through flow during deformation had the effect of significantly decreasing the stratigraphic separation between individual Phanerozoic formations and the Precambrian basement. Intrafolial folds developed in the carapace as products of passive flow. The relatively brittle, massive crystalline basement responded to extension and flattening by ductile normal faulting and the development of penetrative foliation and lineation. The high-temperature extensional disturbance, which began in early Tertiary time and ended about 25 m.y. ago(?), was accompanied by moderately ductile to moderately brittle tectonic denudation and gravity-induced folding of cover rocks. Major listric normal faulting postdated the development of tectonite, shaped the internal fabric of the decollement zones, and effected final movements of the covers.

INTRODUCTION

Purpose

The purpose of this paper is to describe the characteristics of metamorphic core complexes as disclosed by attributes of core-complex terranes in southern Arizona. Emphasis is focused on structural elements and relationships that reveal important mechanical aspects of the kinematic and dynamic evolution of the core complex rocks.

Physical Model

Defining what constitutes characteristics of metamorphic core complexes is a scale-dependent process. Viewed as regional tectonic elements along the length of the western Cordillera from Sonora to southern Idaho, the metamorphic core complexes might be considered "small" outcrop areas of relatively high topographic relief that comprise arches of distinctively deformed and metamorphosed igneous and sedimentary rocks. They occur exclusively in the Basin and Range province. The deformed crystalline rocks are separated from unmetamorphosed country rock by decollement marking strikingly sharp thermal-strain gradients. Resting on the decollement are tiny thin plates of deformed but generally unmetamorphosed Phanerozoic strata. Commonly, the isotopic systems of core complex rocks disclose a mid-Tertiary thermal disturbance. The complexes are commonly co-spatial with Oligocene-Miocene continental sedimentary sequences and Miocene ignimbritic volcanic rocks.

Viewed macroscopically at the scale of several mountain ranges, individual metamorphic core complexes in southern Arizona may be subdivided into four elements: core, metamorphic carapace, decollement, and cover (left half of Fig. 1). Cores consist of either type A—mylonitic augen gneissic, granodioritic rocks with screens of mylonitic schist and metasedimentary metavolcanic rocks—or type B— cataclastically deformed metasedimentary and metavolcanic rocks intruded lit-par-lit by deformed granitic and pegmatitic layers. The core rocks in places are clearly transitional laterally or downward into nontectonite protoliths. Where cores are type A, a distinct metamorphic carapace can

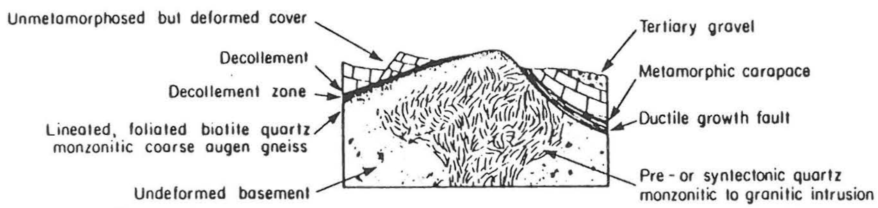


Figure 1. Schematic cross sections showing components of metamorphic core complex.

often be distinguished (center and rights parts of Fig. 1). The carapace consists of penetratively deformed younger Precambrian and Phanerozoic strata metamorphosed to greenschist or amphibolite grade. It forms a crudely tabular layer that overlies concordantly the crystalline rocks of the core. The contact itself is a tectonic contact, commonly so tight (that is, lacking a visible physical discontinuity) that the rocks of the metamorphic carapace appear to be "smeared," "welded," or "plated" onto the crystalline rocks. "Plated" will be used in this paper to describe such contact relationships, following the usage of Engel (1949) in the Adirondacks. In contrast to the tight contact between core and carapace, the contact between cover and carapace, or cover and core, is a decollement marked by brittle low-angle faulting and shearing. The decollement serves to separate two different distinctive strain styles. Typically, the decollement lies atop the core without significant intervening metamorphic carapace (left half of Fig. 1). Below such surfaces, mylonitic augen gneisses and metamorphosed strata of the cores are transformed into a low-dipping, resistant decollement zone of fine-grained cataclastically deformed rocks. The sharp, planar upper surface of this zone, the decollement itself, is overlain by semiconcordant thin sheets containing tectonic slices of unmetamorphosed Phanerozoic strata and Precambrian basement arranged in normal stratigraphic order (left half of Fig. 1).

Another way in which core, carapace, decollement, decollement zone(s), and cover can be bound together is shown in the right half of Figure 1. Where metamorphic carapace is preserved in significant thickness, the major detachment or decollement occurs at the top of the carapace. A resistant tabular decollement zone derived from intensive deformation of the metamorphosed carapace strata may locally develop directly below the detachment surface. But, additionally, a ductile fault zone may form just below the plated contact between core and carapace. The physical nature of the ductile fault zone in the crystalline rocks below the carapace is physically similar, although thinner, than the typical decollement zone that is derived from mylonitic gneiss.

At the mesoscopic scale, the array of structures is awesome in its diversity and pervasiveness: the cores contain low-dipping cataclastic foliation, systematically aligned low-plunging mineral lineation, slickensides, normal ductile faults, abundant boudins and pinch-and-swell features, and recumbent to overturned tight ptygmatic folds in aplites and pegmatities. Zones of contact between dissimilar augen gneisses are frequently marked by folded mylonitic schist. Rocks of the metamorphic carapace show extreme thinning and attenuation. The common array of structures includes tight isoclinal recumbent to overturned intrafolial folds, axial-plane cleavage, boudins, pinch-and-swell, flattened-stretched pebble metaconglomerate, lineation, and ductile faults. The metamorphic fabric is generally overprinted by pervasive fractures. The decollement zone, where it forms at the expense of mylonitic augen gneiss, is structurally a massive to moderately foliated resistant layer of microbreccia, chlorite breccia, mylonite, and mylonitic gneiss shattered by closely spaced fractures. It typically occurs at the upper surface of type A cores and is marked by a planar top. Although the mylonitic rocks are usually derived from augen gneiss, original lineation and foliation are rotated, overprinted, and masked by cataclastic granulation. Above the detachment surface atop the decollement zone, there may be some metamorphic carapace, but generally there are unmetamorphosed but deformed cover rocks characterized by overturned intraformational asymmetric folds, bedding-plane cleavage, and faults. Although in nor-

mal stratigraphic succession, formations represented in the cover rocks are markedly tectonically thinned or may be absent altogether.

Region of Study

The tectonic setting of southeastern Arizona is ideal for examining the structural geology of metamorphic core complexes and for interpreting their dynamic significance. Within this region, outside the metamorphic core complexes, granitic gneiss is insignificant volumetrically and, thus, not to be confused with gneisses in the core complexes themselves. Furthermore, in southeastern Arizona, no tectonites exist in the Precambrian and Phanerozoic rocks, except for 1.6- to 1.8-b.y.-old Pinal Schist, and, thus, the tectonites in the core complex terranes are unique to the region. This statement does not hold for the setting of core complexes west of Tucson in the Papago Indian Reservation. There, Mesozoic sedimentary and volcanic country rocks are metamorphosed to phyllite and schist. Finally, excellent exposures and great structural relief afford examination of a diversity of rocks at many different structural levels.

Southeastern Arizona may be an ideal setting in yet another, but more controversial, respect. Although rocks in southeastern Arizona have been affected by multiple, superposed deformations in the Phanerozoic, the deformations may not have produced at any time a regional low-angle overthrusting. If it is true that some core complexes of southeastern Arizona occur in a setting where major overthrusting *has not* occurred, then the descriptive characteristics of these complexes would have a significant bearing on dynamic interpretations of other complexes in the western Cordillera. The various models proposed for the nature of Sevier-Laramide deformation in southeastern Arizona will not be presented here. Papers that bear on the problem and that provide background for considering the metamorphic core complexes within their regional and historical framework include Cooper and Silver (1964), Davis (1977b, 1979), Drewes (1973, 1976a, 1978b), Gilluly (1956), Hansen and others (1978), Keith and Barrett (1976), Rehrig and Heidrick (1972), and Thorman (1977).

When compared with the normal country-rock terrane of southern Arizona, the rocks of the metamorphic core complexes are *striking* dynamo-thermal anomalies. The classic geologic terrane generally consists of (1) low-grade metasedimentary Pinal Schist (1.6 to 1.8 b.y. old) intruded by largely undeformed Precambrian (1.55 to 1.65 b.y. old) granite plutons (Ransome, 1903; Silver, 1955; Silver and Deutsch, 1961; Erickson, 1962; Cooper and Silver, 1964); (2) unfoliated, unlineated, 1.4- to 1.5-b.y.-old "anorogenic" Precambrian granitic rocks (Drewes, 1976b; Silver and others, 1977a, 1977b); (3) a thin (about 2,000 m) sequence of faulted, generally unfolded, commonly homoclinal, younger Precambrian and Paleozoic sedimentary rocks (Silver, 1955; Bryant, 1955, 1968; Butler, 1971); (4) a thick sequence of faulted, commonly wedge-shaped deposits of Triassic(?)-Jurassic volcanic and sedimentary rocks (Hayes and others, 1965; Simons and others, 1966; Hayes and Drewes, 1968; Hayes, 1970b; Cooper, 1971; Drewes, 1971); and (5) a thick sequence of gently to tightly folded Cretaceous sedimentary rocks (Hayes and Drewes, 1968; Drewes and Finnell, 1968; Finnell, 1970; Hayes, 1970a; Drewes, 1972). Transformation of these country rocks to tectonites in the sites of the metamorphic core complexes is believed to provide kinematic and dynamic clues to processes responsible for the evolution of core complexes in general.

The presently known metamorphic core-complex terranes in the Basin and Range province of southernmost Arizona are shown in Figure 2. Core-complex terranes are exposed in the Santa Catalina, Rincon, and Tanque Verde Mountains bounding Tucson on the north and east; in the Tortolita Mountains, Suizo Hills, Durham Hills, and southern Picacho Mountains in the corridor between Tucson and Phoenix; in the northern Pinaleno Mountains and eastern Santa Teresa Mountains (Jackson Mountain) west of Safford; and, to the southwest of Tucson, in the Coyote Mountains, the

outhe okiva ntain) Verc ntain of th bhabi ains, Blanca, Alvarez Mountain, and Kupk Hills. Noteworthy is the change in trend of the belt of metamorphic core complexes from N50°W in western and central Arizona to north-south from Tucson to Hermosilla, Sonora.

The Rincon–Santa Catalina–Tortolita metamorphic core complex contains well-exposed cataclastically deformed crystalline rocks and elegantly deformed metasedimentary and sedimentary rocks. Although the evolution of the rocks in these mountains is by no means completely understood, there exists a large and significant amount of data regarding the geologic and geochronologic relationships. Geologic maps of major parts of the Rincon–Santa Catalina–Tortolita complex have been made by Moore and others (1941), Pashley (1966), Drewes (1975, 1878a), Creasey and Theodore (1975), Davis and others (1975), Budden (1975), Banks (1976), and Banks and others (1977). Geochronologic data, including K-Ar, Rb-Sr, fission-track, and U-Pb, have been contributed by Damon and others (1963), Damon (1968a, 1968b), D. E. Livingston (unpub. Rb-Sr data, Laboratory of Geochronology, the University of Arizona), Shakel (1974), Creasey and others (1976), Shakel and others (1977), and Keith and others (this volume). The structural fabric of the rocks has been analyzed by Mayo (1964), Waag (1968), Peterson (1968), Davis (1973, 1975, 1977a), Davis and others (1974, 1975), and Davis and Frost (1976).

In contrast to the geology of the Rincon–Santa Catalina–Tortolita complex, the geology of the other ranges shown in Figure 2 is relatively little known. These terranes are dominated geologically by augen gneissic rocks and, unfortunately, are generally lacking in metasedimentary carapace rocks and unmetamorphosed but deformed cover rocks. Small-scale county geologic maps prepared by the Arizona Bureau of Mines provide some data (Wilson and others, 1957; Wilson and Moore, 1959; and Wilson and others, 1960). The Picacho Mountains have been mapped in reconnaissance fashion by Yeend (1976). Similarly, Haxel and others (1978) have mapped the Comobabi Mountains. Wargo and Kurtz (1956) mapped the geology of the Coyote Mountains. May and Haxel (1979) have recently mapped the Sells quadrangle. My own contributions to the geologic mapping include 1:24,000 mapping in the Tortolita Mountains (Davis and others, 1975; Banks and others, 1977), large-scale mapping in the Rincon Mountains (Davis and others, 1974; Davis and Frost, 1976), 1:62,500 mapping of the Coyote, Picacho, Sierra Blanca, Pozo Verde Mountains, Jackson Mountain, and the northeastern Pinaleno Mountains, and 1:20,000 mapping in Happy Valley (Davis and others, in prep.).

In this presentation, no attempt is made to outline the geology of the complexes, range by range. Rather the characteristics of the metamorphic core complexes are discussed from inside out, from augen gneissic core, through metasedimentary carapace, through decollement zone, to unmetamorphosed but folded and/or faulted cover rocks.

METAMORPHIC CORE

Augen Gneiss

Macroscopic Structures. By far the greatest volume (>90%) of rocks in the metamorphic core complexes of southern Arizona comprise the cores. These individual core zones underlie exposed surface areas of as much as 450 km² and, within single complexes, express an exposed vertical relief of as much as 1,800 m . These cores of mostly mylonitic augen gneiss undergird some of the highest summits in the Mountain subprovince of the Basin and Rane province (for example, Mount Lemmon in the Santa Catalina Mountains and Mica Mountain in the Rincon Mountains at about 2,700 m; Fig. 3).

Foliation in the individual mylonitic augen gneissic cores of the complexes is characteristically low dipping (see Fig. 2) and commonly defines large upright, doubly plunging foliation arches or antiforms,

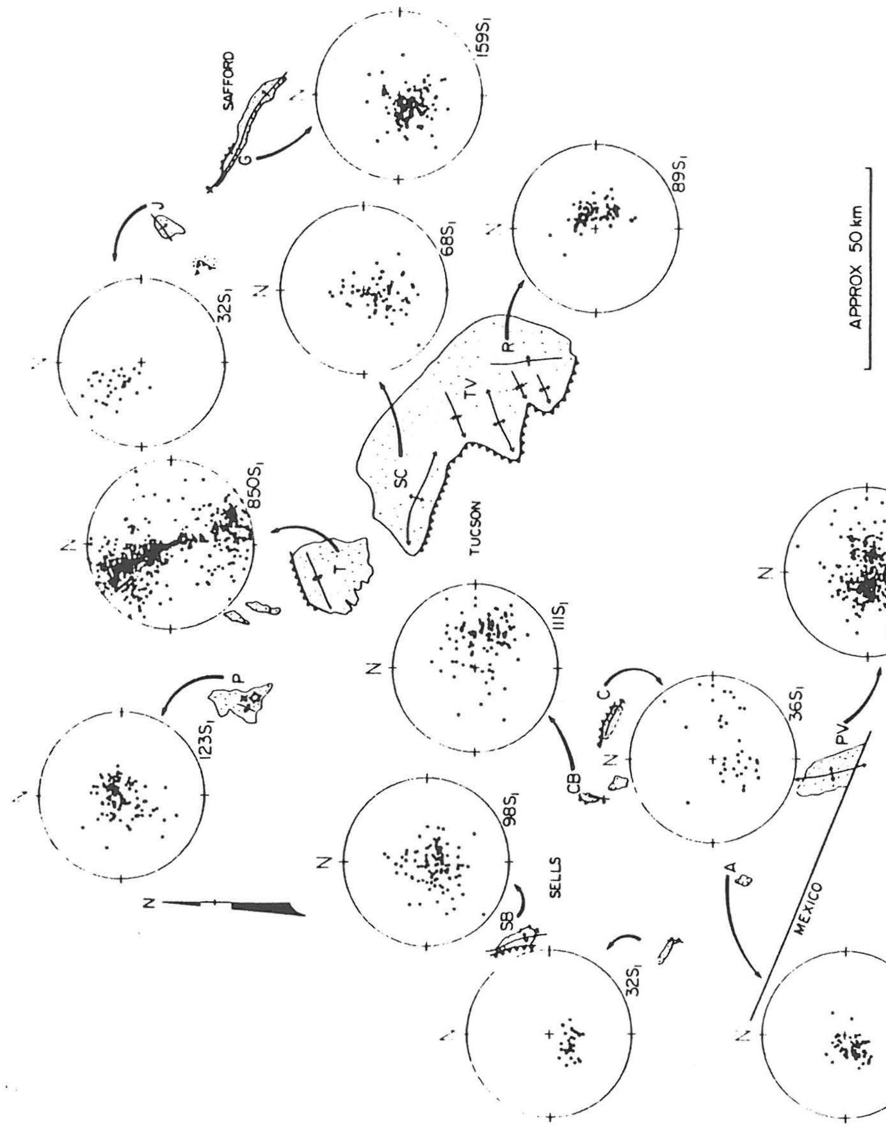


Figure 2. Map showing distribution of metamorphic core complexes in southern Arizona. Base used in preparation of figure is a palinspastic reconstruction of southern Arizona (prepared by Peter J. Coney, Richard Nielson, and me) that portrays the geology of southern Arizona after Laramide and before basin and range tectonism. The reconstruction assumes a conservative 10% to 15% extension in the last 16 m.y. J = Jackson Mountain in the eastern Santa Teresa Mountains; G = Graham Mountains (also known as Pinaleno Mountains); P = Picacho Mountains; T = Tortolita Mountains; SC = Santa Catalina Mountains; TV = Tanque Verde Mountains; R = Rincon Mountains; SB = Sierra Blanca; NC = northern Comobabi Mountains; C = Coyote Mountains; PV = Pozo Verde Mountains; A = Alvarez Mountains; K = Kupk Hills. Sawtooth symbol = decollement zone. Lower-hemisphere projections show poles to foliation in rocks of the core and metamorphic carapace rocks.

half-arches, upright synforms or troughs, and irregular ameboid domical structures. Some of the doubly plunging arches, like the Tanque Verde antiform (Fig. 4A) and Sierra Blanca (Fig. 4B), have exceptional physiographic expression. Tanque Verde antiform merges with two other antiforms and a synform in the Rincon Mountains to define a crudely ameboid-shaped domical structure (Moore and others, 1941; Pashley, 1966; Drewes, 1978a). Other archlike structures are found in the forerange of the Santa Catalina Mountains (Pashley, 1966; Peterson, 1968) and the Pozo Verde Mountains (Fig. 4C). Quaquaversal dips characterize the low-dipping foliation in the Picacho Mountains (Fig. 4D), and these define a crude, irregular dome. Half-arches are the dominant macroscopic structural form of the Coyote Mountains (Fig. 4E), part of the Comobabi Mountains (Fig. 4F), Jackson Mountain (Fig. 4G), and the northeastern Pinaleno Mountains (Fig. 4I). The internal structure of foliation in the central Tortolita Mountains appears to be archlike (Fig. 4H), but mapping reveals that no actual closure exists.

In general, the penetrative mineral lineation within the arched augen gneissic rocks is coaxial with the axis (axes) of arching. The only significant exception is the forerange of the Santa Catalina Mountains (Peterson, 1968). There, formation of the east forerange fold quite clearly resulted in rotation of the original undirectional penetrative lineation.

Lithology

Coarse-Grained Quartz Monzonitic Augen Gneiss. The mylonitic gneisses in the cores of some of the terranes are derived in part from coarse-grained Precambrian quartz monzonite, approximately 1.4 to 1.5 b.y. old. This interpretation has been offered by U.S. Geological Survey geologists in the Santa Catalina–Rincon–Tortolita complex for coarse-grained, dark-colored, biotite augen gneiss and schist (Fig. 5). This rock has been mapped as such by Drewes (1975, 1978a) in the Rincon and Tanque Verde Mountains, by Creasey and Theodore (1975) in the eastern part of the Santa Catalina Mountains, and by Davis and others (1975) and Banks and others (1977) in the Tortolita Mountains. To the north of Tucson, Banks and others (1977) mapped lithologically similar rock in the Durham and Suizo Hills, and I mapped it in parts of the Picacho Mountains. The progressive transformation of the recognizable nonfoliated Precambrian porphyritic quartz monzonite to coarse-grained augen gneiss, and even to mylonitic schist, can be observed at a number of locations in southern Arizona (Davis and others, 1975; Banks and others, 1977; Banks, this volume). The most recently discovered locale in which this transition can be observed is in the northeastern Pinaleno Mountains (Fig. 4I). Isotopic data also have confirmed that at least part of the coarse-grained augen gneiss was derived from Precambrian porphyritic quartz monzonite. Shakel and others (1977) reported that zircons from coarse-grained biotite augen gneiss in the Santa Catalina forerange indeed yield an age of 1,440 ± 10 m.y.

The microscopic petrography of the coarse-grained quartz monzonitic augen gneiss has been described in detail by a number of workers, especially Peterson (1968) and Sherwonit (1974), and they are described in this volume by Banks. In outcrop, the augen gneiss is relatively dark colored and cut by the penetrative flat to moderately dipping cataclastic foliation (Fig. 6A). Concordant aplite and pegmatite veins serve to enhance the foliation, and such sills and veins commonly display abundant boudinage and pinch-and-swell. Aplites and pegmatites that cut the foliation at high angles are almost always ptygmatically folded (Fig. 6B). These folds are axial planar with respect to the low-dipping penetrative foliation and, thus, are typically strongly overturned to recumbent.

Medium-Grained Quartz Monzonitic Augen Gneiss

In addition to the mylonitic gneisses derived from the Precambrian quartz monzonite, all of the Arizona core complexes expose abundant mylonitic augen gneisses that are derived from quartz

Figure 3. Northwest-directed U-2 oblique aerial photograph showing the Santa Catalina Mountains and Tanque Verde–Rincon Mountains, north and east of Tucson, respectively. Ranges are composed dominantly of foliated, lineated quartz monzonitic augen gneiss. The gneisses are separated from unmetamorphosed but deformed sedimentary cover rocks by a curviplanar decollement zone that marks the basin-mountain interface on the south side of the Santa Catalina Mountains and the western and southern sides of the Tanque Verde–Rincon Mountains. Rocks of the metamorphic carapace are preserved along the very summit of the Santa Catalina Mountains and on the eastern side of the Rincon Mountains. The northern part of the Pinaleno Mountains (dark range in right background), 95 km northeast of the Rincon Mountains, displays deformation of the metamorphic core complex as

Figure 4 (A through I). Maps showing internal structure of core rocks in nine ranges. Structural symbols show strike and dip of foliation and trend and plunge of lineation.

A

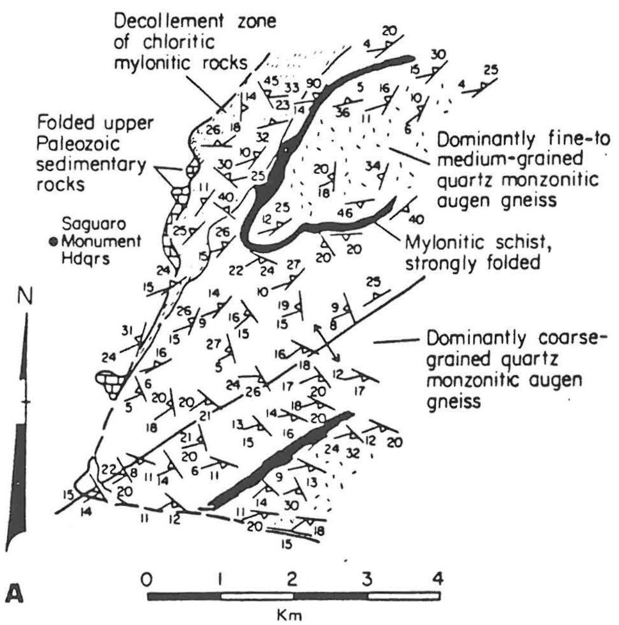


Figure 4A. Tanque Verde Mountains.

B

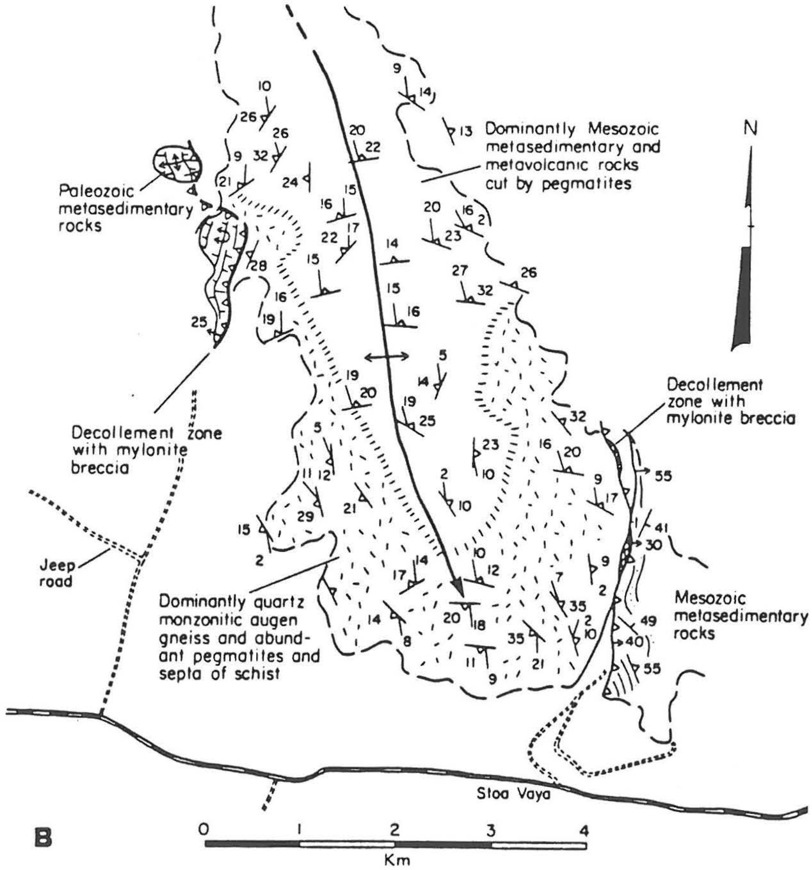


Figure 4B. Sierra Blanca.

C

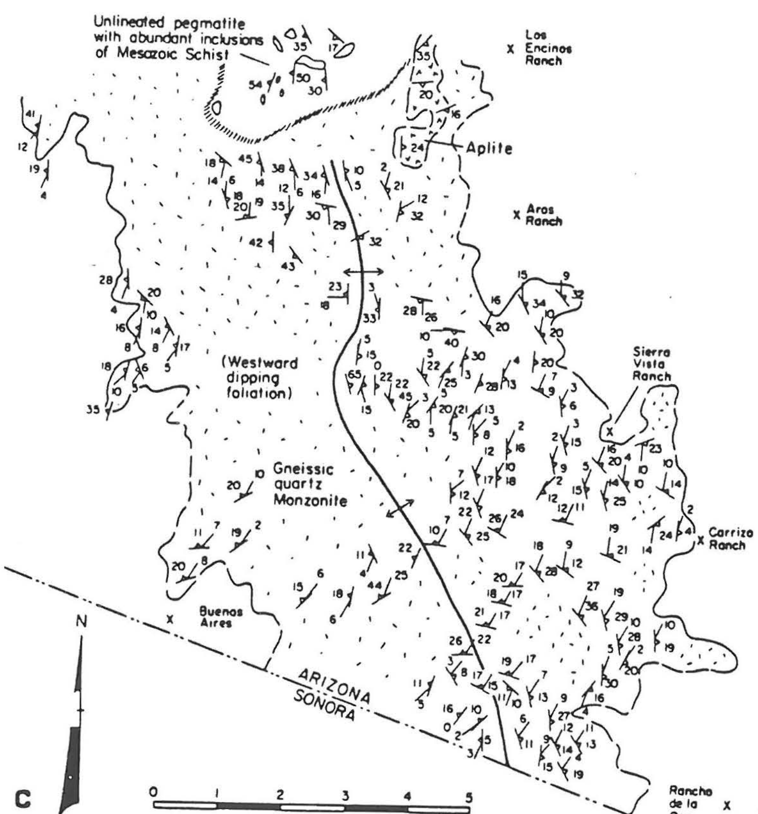


Figure 4C. Pozo Verde Mountains.

D

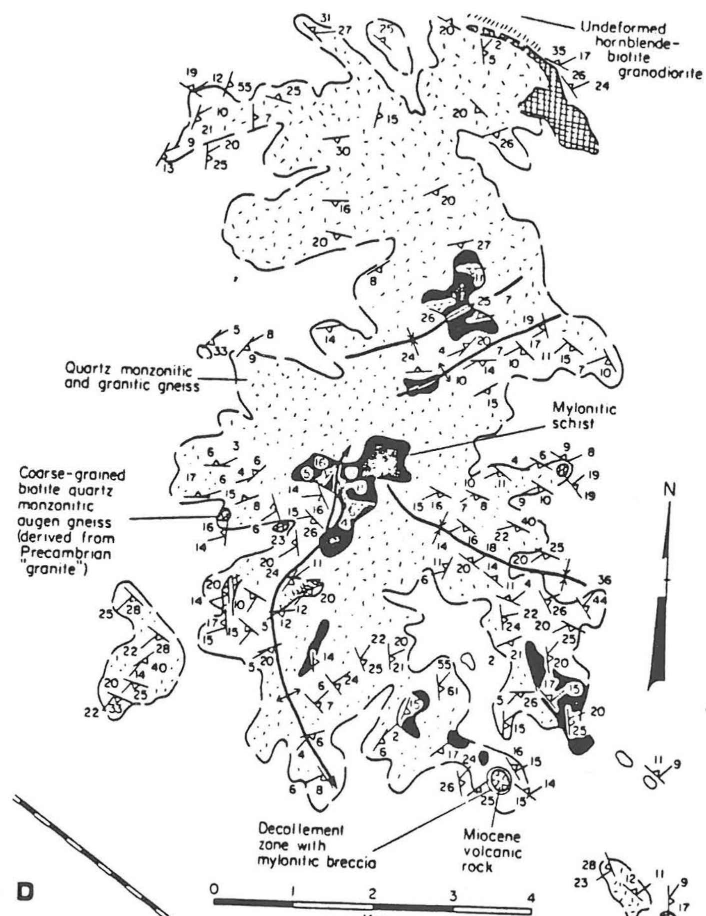


Figure 4D. Picacho Mountains.

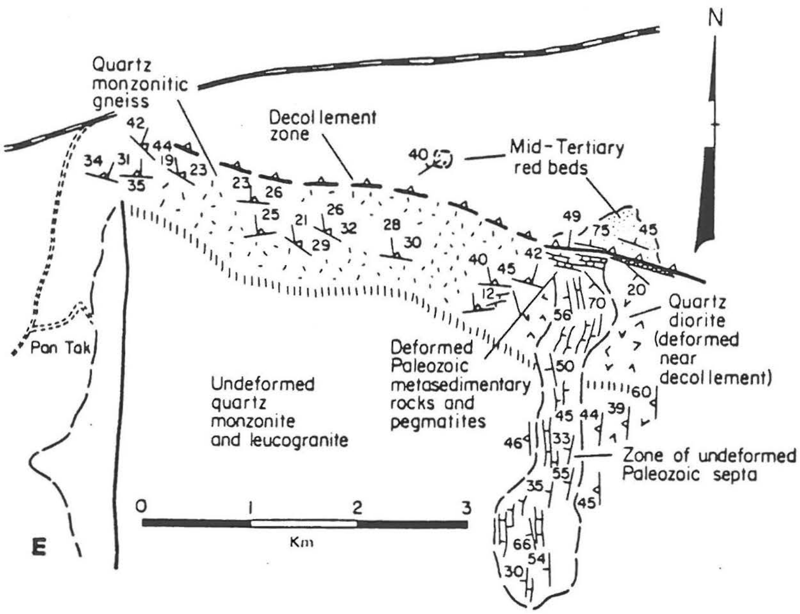


Figure 4E. Coyote Mountains.

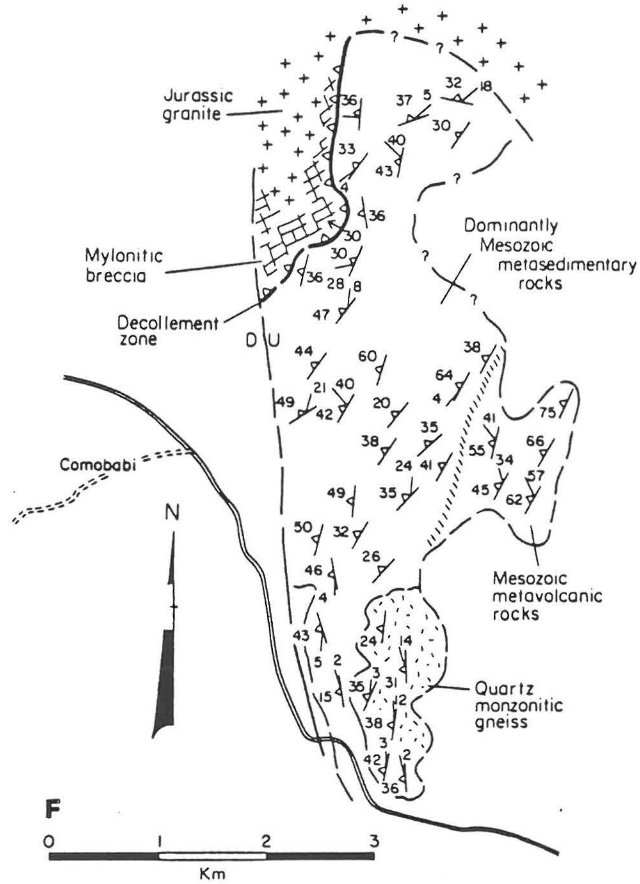


Figure 4F. Comobabi Mountains.

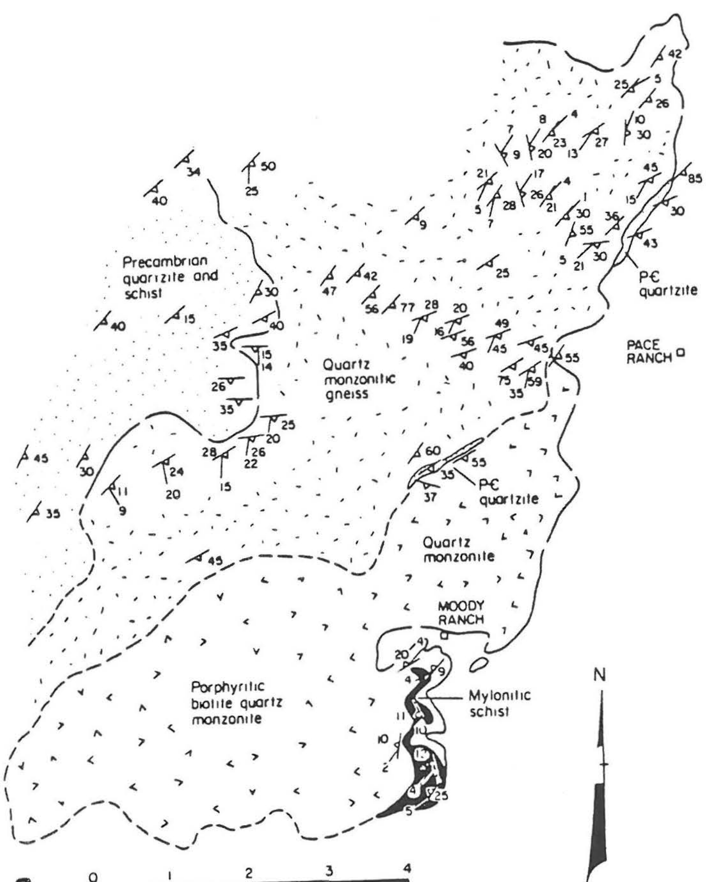


Figure 4G. Jackson Mountains.

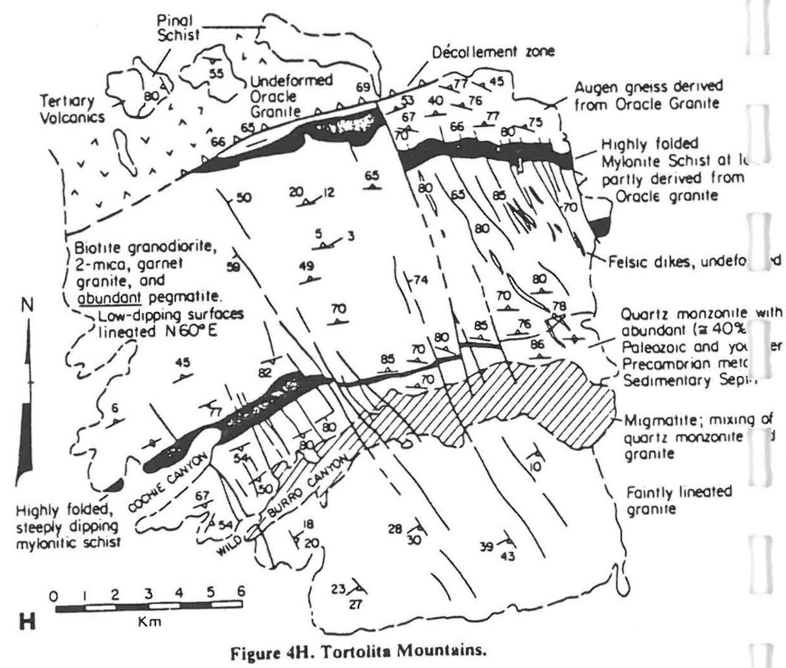


Figure 4H. Tortolita Mountains.

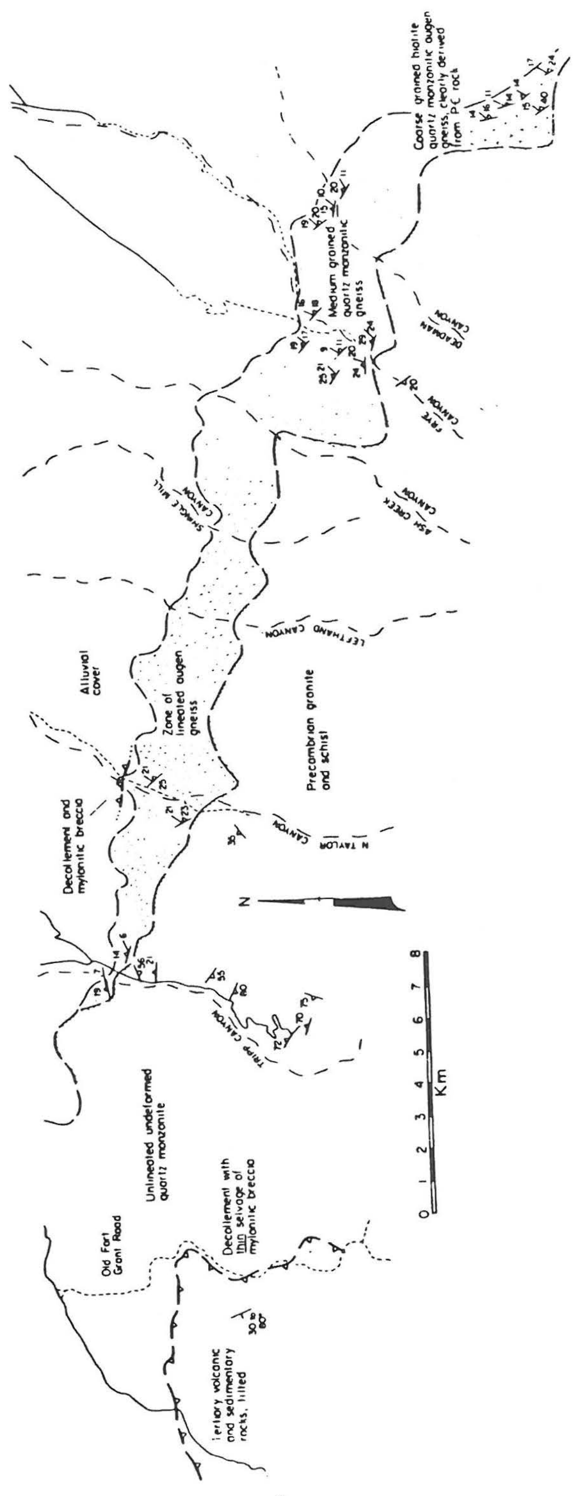


Figure 41. Pinaleno Mountains.

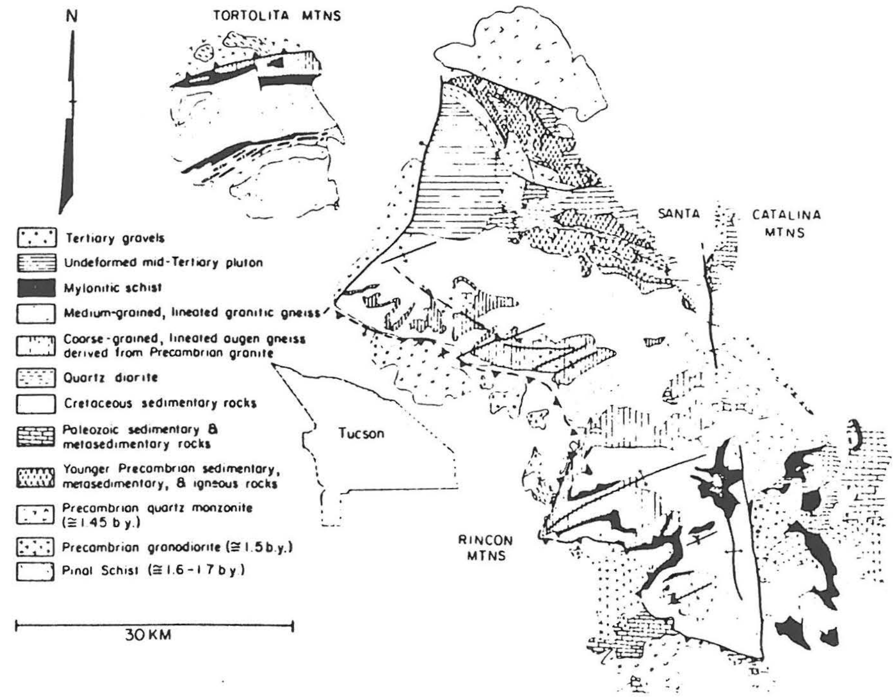


Figure 5. Generalized structure map of the Rincon–Santa Catalina–Tortolita complex. Adapted and interpreted from the maps of Banks (1974), Creasey and Theodore (1975), Drewes (1975, 1978a), Davis and others (1975), and Banks and others (1977).

monzonite, granodiorite, granite, and alaskite of Phanerozoic age. Many textural and compositional varieties of these rocks exist, but most tend to be medium grained and quartz monzonitic in composition. Garnet-bearing, two-mica granites are not uncommon. As in the case of the coarse-grained augen gneiss, a distinguishing structural characteristic of these rocks is low-dipping cataclastic foliation and penetrative lineation.

In the Rincon Mountains, the medium-grained augen gneisses have been mapped by Drewes (1975, 1978a) as belonging to the Precambrian(?) Wrong Mountain quartz monzonite. In the Santa Catalina Mountains, Creasey and Theodore (1975), Banks (1976), and Creasey and others (1976) have included the medium-grained augen gneisses as components of a mid-Tertiary composite batholith (quartz monzonite of Sanmaniego Ridge). In the Tortolita Mountains, Davis and others (1975) and Budden (1976) mapped three main phases of medium-grained augen gneiss that range from biotite granodiorite to granite in composition.

The dominant rocks in the Picacho Mountains, Pinaleno, and Jackson Mountain core-complex terranes are fine- to medium-grained quartz monzonitic augen gneisses. These display a striking variety of textures, with deformational fabric ranging from barely recognizable to mylonitic and schistose. Similar rocks dominate the terrane in the southern Pozo Verde Mountains, Alvarez Mountains, and Kupk Hills. To the north in the Coyote Mountains, alaskite, biotite quartz diorite with septa of Paleozoic rocks, and quartz monzonite were mapped by Wargo and Kurz (1956). In the northern Coyote Mountains these phases are augen gneissic and marked by penetrative lineation and low- to moderate-dipping cataclastic foliation. Pegmatites are abundant. In the Sierra Blanca and part of the northern Comobabi Mountains, lineated, foliated fine- to medium-grained quartz monzonite augen gneiss, alaskite, pegmatite, and minor biotite granodiorite invade a terrane of metamorphosed Mesozoic sedimentary and volcanic rocks. Mapping, dating, and correlating all of the fine- to medium-grained augen gneiss at a 1:24,000 scale constitutes a major regional geologic problem!

The petrography of the medium-grained augen gneisses has been examined most carefully in the Santa Catalina Mountains by Pilkington (1962), Peterson (1968), Sherwonit (1974), Shakel (1975), and Banks (this volume). In outcrop the rocks are light colored, hypidiomorphic, equigranular, and generally homogeneous. As in the case of the coarse-grained augen gneiss, foliation is typically defined by thin laminae of quartz, parallelism of micas, and aligned augen of cataclastically deformed feldspar. The rocks are cut by abundant pegmatite, aplite, and flat joints. The striking physiographic expression of the penetrative low-dipping foliation and jointing calls attention to these core-complex gneisses from afar.

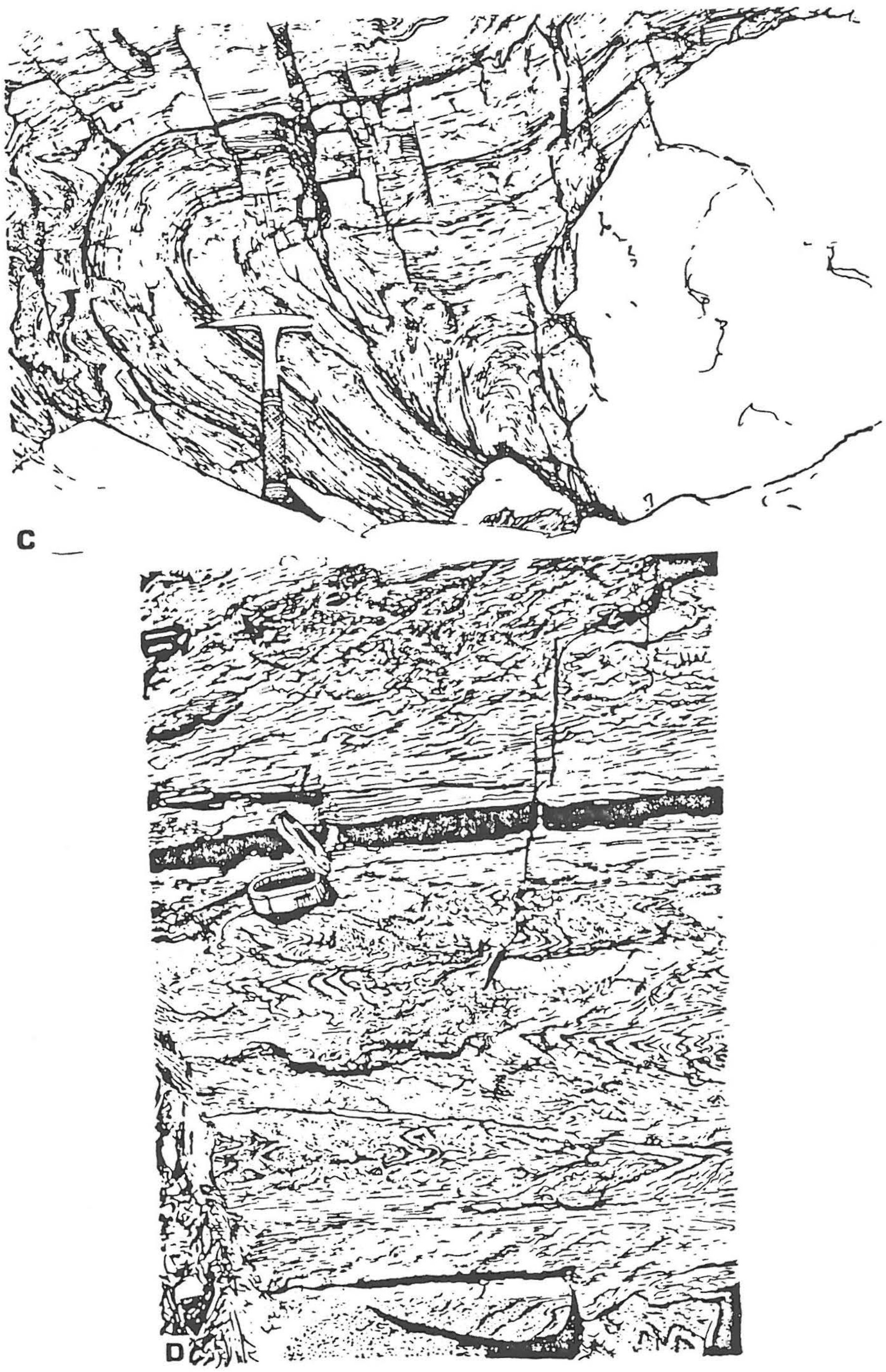
The general distribution of the medium-grained augen gneisses in the Rincon–Santa Catalina–Tortolita complex is shown in Figure 5. Establishing the exact age of these gneisses has been difficult, for the rocks are characterized by profoundly disturbed Rb-Sr isotopic systems. K-Ar and fission-track ages (from the work of Damon and his associates of the University of Arizona, and Creasey and others of the U .S. Geological Survey) typically range from 30 m.y. to 24 m.y. for the quartz monzonitic gneisses and associated pegmatites and aplites. Through the years, these dates have been interpreted variously as cooling ages by some and as emplacement ages by others. For example: (1) Creasey and others (1976) and Banks (this volume) regard the entire body of fine- to medium-grained augen gneisses as a composite batholith 25 m.y. old that was penetratively deformed at the time of emplacement. (2) Shakel and others (1977) reported that part of the so-called composite batholith is indeed 27 m.y. old, but that monzonites from a two-mica, garnet-bearing part of the mass known locally as the Wilderness granite gives U-Th-Pb ages of 44 to 47 m.y. They interpreted the Wilderness granite to be undeformed and concluded that the last major gneiss-forming event was no younger than 44 m.y. (3) Drewes (1978a) considered the mass of fine- to medium-grained quartz monzonite augen gneiss in the Rincon Mountains to be Precambrian in age or to have been derived from Precambrian rocks.

The article by Keith and others (this volume) provides a useful summary of the geochronological problems and offers some new interpretations. It is becoming increasingly well documented that the core-complex terranes contain both mid-Tertiary *and* pre-mid-Tertiary Phanerozoic plutons. Recent isotopic data suggest that some of the mid-Tertiary bodies have been affected by low-dipping foliation and penetrative lineation (Rehrig and Reynolds, 1978, and this volume). The Tertiary event, whatever its dynamic nature, imposed its thermal-tectonic signature on all core rocks, regardless of whether they existed before the event or were emplaced during the event.

Mylonitic Schist

The contact zones between the coarse- and the medium-grained augen gneissic phases are commonly marked by strongly foliated, folded rocks that herein are referred to as mylonitic schist. Such highly deformed, cataclastic rocks are well exposed in Tanque Verde Mountain, the forerange of the Santa Catalina Mountains, and in the Little Rincon, Tortolita, Picacho, and Sierra Blanca Mountains. These rocks are fine to very fine grained and range in color from brown to steel gray to black (Fig. 6C). White broken feldspar chips, irregular in size and shape, commonly "float" in the fine-grained, mylonitic matrix. Deformed concordant aplite and pegmatite layers and veins accentuate the foliation. Penetrative folding is a characteristic of these rocks and presumably reflects movements that helped to shape the mylonitic, foliate fabric. The folds are intrafolial, tight to isoclinal, overturned to recumbent structures whose axes lie in the low-dipping foliation. Orientations of folds in mylonitic schist in a

Figure 6 (facing pages). Tracings of photographs of structures in coarse-grained quartz monzonitic augen gneiss (A) Foliation as seen in outcrop. (B) Ptygmatically folded aplite dikes cutting augen gneiss. (C) Folded myloniti schist. (D) Transposed foliation in coarse-grained quartz monzonitic augen gneiss.



number of ranges in southern Arizona are portrayed stereographically in Figure 12. Foliation is essentially concordant with that of the adjacent augen gneissic phases. Foliation surfaces in the mylonitic schist are not always marked by mesoscopically recognizable mineral lineation, but where lineation occurs, it too is parallel to that of the augen gneissic phases.

It is not clear in all cases whether the mylonitic schist has evolved cataclastically from the coarse-grained quartz monzonitic augen gneiss, the fine- to medium-grained quartz monzonitic augen gneiss, or both. Adjacent coarse-grained augen gneiss is commonly marked by transposition and folding of the penetrative low-dipping foliation (S_1), producing zones of moderately dipping crenulation foliation (S_2; Fig 6D). The fine- to medium-grained quartz monzonitic augen gneisses and associated pegmatites in contact with the mylonitic schist commonly display rather remarkable overturned to recumbent folds.

A significant field problem is distinguishing the mylonitic schist derived by cataclasis of augen gneiss from actual Pinal Schist. Mayo's (1964) investigation of folds in the schist in the forerange of the Santa Catalina Mountains is, in my judgment, an analysis of structures in mylonitic schists cataclastically derived from orthogneiss. The folds there occur in a deformed zone between the coarse- and the medium-grained augen gneissic phases. Davis and others (1975), in mapping the geology of the Tortolita Mountains, noted the transformation of undeformed porphyritic quartz monzonite to coarse-grained quartz monzonitic augen gneiss. Banks demonstrated to us that even the very fine grained, folded schistose rocks that we had mapped as Pinal Schist may be in large part derived from quartz monzonite (Banks and others, 1978). Drewes's (1978a) geologic map of the Rincon and Tanque Verde Mountains shows Pinal Schist in narrow bands, generally occurring in the contact zone between the coarse-grained and the fine- to medium-grained augen gneissic phases. Some of this rock is indeed Pinal Schist but much of it was derived from orthogneiss (Fig. 4A). Wright (1978) has prepared a large-scale map of part of the mylonitic schist zone on the northwest flank of Tanque Verde Mountain showing its relation to augen gneissic phases (Fig. 7). Yeend (1976) mapped undifferentiated schist in the Picacho Mountains and suggested (in the explanation of his map) that these may have been derived from sedimentary rocks. These schists, for the most part, cap the mountain and closely resemble mylonitic schists found in other ranges. Some of the schists that occur at midslope on the western flank are located at the contact between the coarse-grained and the fine- to medium-grained augen gneiss as are those in other ranges.

At the scale of individual outcrops, mylonitic schist forms thin bands and layers at contacts between the medium-grained augen gneiss and concordant aplite and pegmatite. Such mylonitic schists are especially well developed in the Sierra Blanca. Black mylonites form at the margins of the abundant pegmatite-aplite sills. Such relations draw attention to the role of ductility contrast in controlling development of the mylonitic schist.

Lineation

The augen gneisses everywhere display low-plunging mineral lineation of a cataclastic nature. The lineation generally is penetrative on the scale of hand specimen, with a delicacy of development dependent in large part on grain size of the host rock. Figure 8A reveals the physical nature of the lineation in coarse-grained augen gneiss in outcrop. The lineation has a variety of forms, but generally is expressed in the plane of foliation of the augen gneisses by the alignment of long directions of inequant feldspar and quartz-feldspar augen, striae and slickensides, elongate aggregates and streaks of crushed minerals, and crenulations on quartz ribbons. In the Santa Catalina Mountains alone, the lineation pervades approximately 2,000 m of augen gneiss. Lineation is by no means restricted to the augen gneisses, but occurs penetratively on concordant or discordant *low-dipping* aplite and pegmatite layers, mylonite zones, transposition foliation surfaces, and low-angle normal faults. Even in some

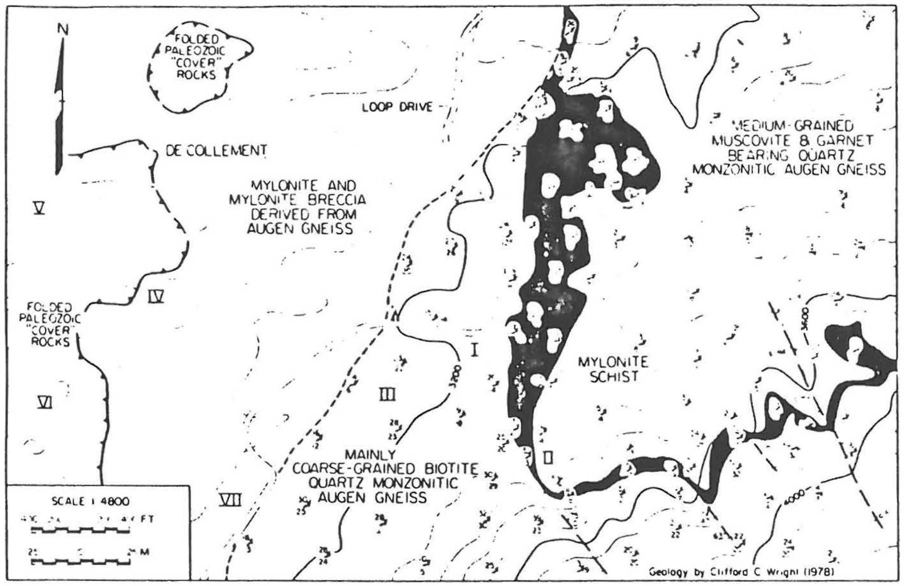


Figure 7. Structural geologic map of part of Saguaro National Monument showing relationship of mylonitic schist to augen gneisses and the decollement zone (from Wright, 1978).

equigranular hypidiomorphic quartz monzonitic bodies lacking penetrative lineation, such as those in the central Tortolita Mountains or the southern part of Jackson Mountain, the lineation locally occurs on shallow-dipping aplite and pegmatite layers that have served to localize tectonic movements.

Within any given metamorphic core complex the lineation is generally remarkably systematic in orientation (Fig. 8B). Lineation in the Picacho, Tortolita, and most of the Santa Catalina and Rincon Mountains trends N50°E to N70°E. Departure from the modal N60°E trend is evident between the Santa Catalina and Rincon Mountains, where lineation is approximately north-south, and near the crest of the Santa Catalina Mountains where it trends north-northeast. In Jackson Mountain and the northern Pinaleno Mountains, the trend of lineation is on the average N30°E and N40°E, respectively. Southwest of Tucson, lineation trends about N10°W in the Sierra Blanca, Comobabi Mountains, and Kupk Hills; north-south in the Coyote and Alvarez Mountains; and N30°E in the Pozo Verde Mountains. This systematic swing in lineation trend occurs in the inner arc of the bend in the belt of metamorphic core complexes in southern Arizona. In essence, three regional structural domains can be distinguished in southern Arizona on the basis of homogeneity of penetrative lineation (Fig. 8C): a western, "Papago" domain characterized by north-trending lineation, a central, "Catalina" domain of east-northeast–trending lineation, and an eastern, "Pinaleno" domain characterized by northeast-trending lineation. Axes of the folds in mylonitic schists in the core rocks are generally oriented parallel to lineation. Axes of boudinage and pinch-and-swell structures, abundant in the lineated core rocks, are more diffuse but tend to lie at right angles to lineation. .

Ductile Normal Faults

Ductile normal faults (Fig. 9A), first discovered by Davis and others (1975) in the Tortolita Mountains, are associated with mylonitic, lineated rocks and typically result in impressive local thinning of

Figure 8A. Penetrative lineation in coarse-grained quartz monzonite augen gneiss.

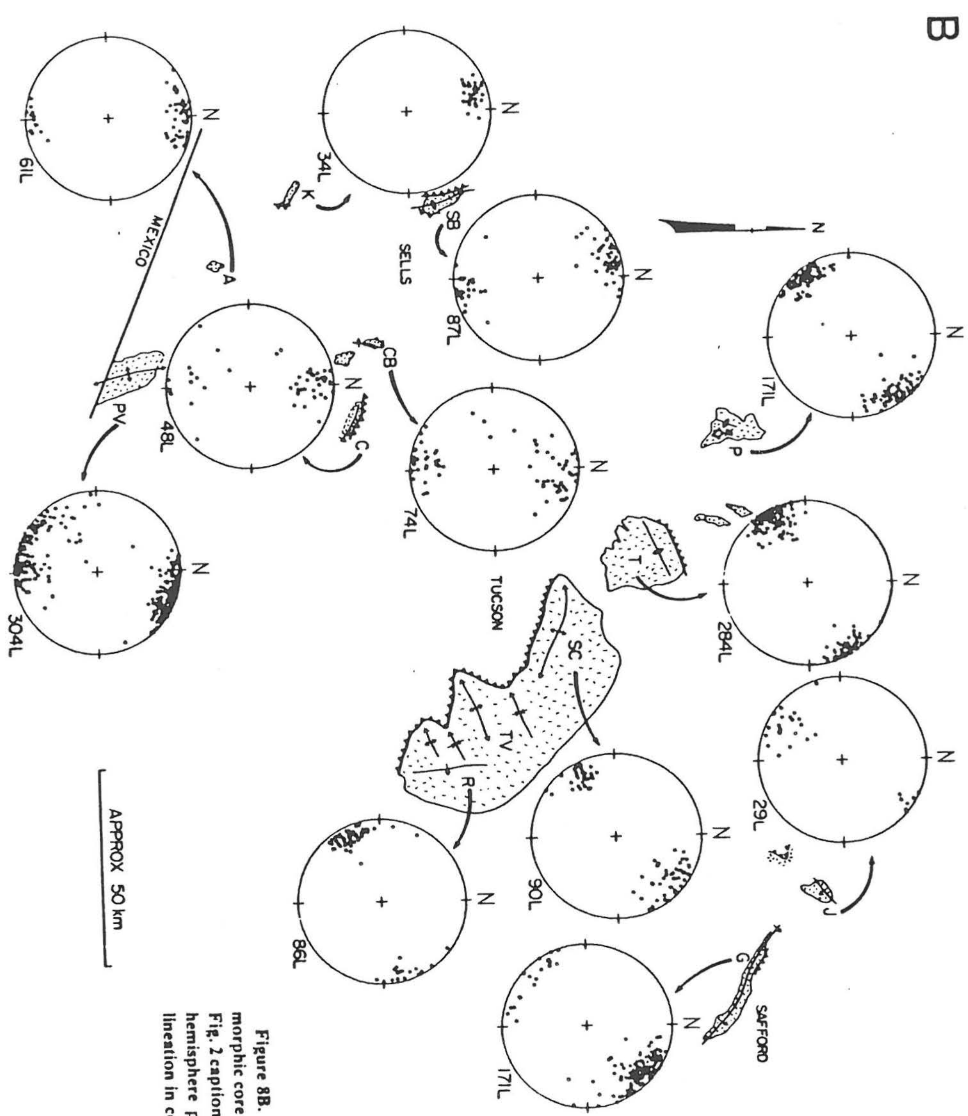


Figure 8B. Map showing distribution of metamorphic core complexes in southern Arizona (see Fig. 2 caption for explanation of symbols); lower-hemisphere projections disclose orientations of lineation in core rocks.

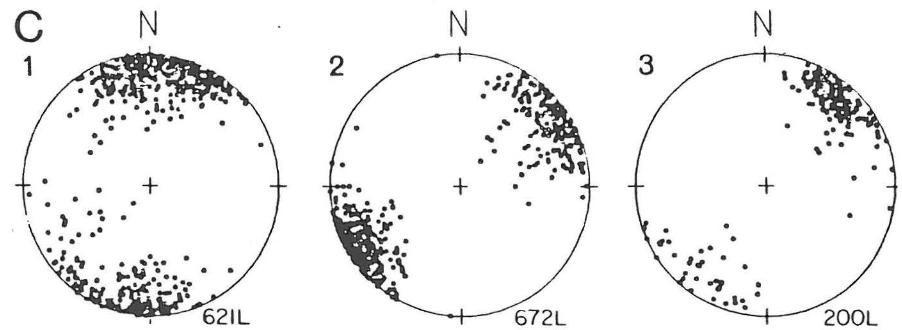


Figure 8C. Lower-hemisphere equal-area projections of lineation orientations in (1) the "Papago" domain (Sierra Blanca, Comobabi, Coyote, Pozo Verde, Alvarez, and Kupk Mountains), (2) the "Catalina" domain (Rincon, Tanque Verde, Santa Catalina, Tortolita, and Picacho Mountains), and (3) the "Pinaleno" domain (Jackson and Pinaleno Mountains).

these rocks within the core. The ductile normal faults are always oriented at right angles to the lineation, regardless of its absolute orientation. I have observed them in all of the ranges shown in Figure 2 except the Pozo Verde Mountains, Alvarez Mountains, and Kupk Hills. Resultant folds that naturally arise from the ductile faulting are superimposed on folds in mylonitic schist. Where examined by Davis and others (1975) in the Tortolita Mountains, the antiforms and synforms are gentle to open, characterized by wavelengths and amplitudes of approximately 6 to 3 m, respectively. Axial surfaces are moderately steep to steeply dipping and strike at right angles to the lineation. In profile, the associated folds are seen to produce marked thinning in the zones of maximum inflection. In some cases, the hinge zones of the folds lack visible offset, and ductile flexing appears to be superseded by actual normal faulting (Fig. 9B).

METASEDIMENTARY CARAPACE

General Relations

In southern Arizona metamorphic core complexes, the metasedimentary carapace, where present, consists of younger Precambrian and lower Paleozoic metasedimentary rocks that rest concordantly atop type A core rocks. Strata in the carapace are commonly metamorphosed to upper greenschist and amphibolite grade and form a relatively thin, tabular sheet. The rocks appear to be concordantly welded or *plated* to underlying crystalline rocks. For example, in Happy Valley east of the Rincon Mountains, there are many outcrops where not even a crack separates younger Precambrian quartzite or marble of the metamorphic carapace from underlying, cataclastically foliated crystalline basement. Along the crest of the Santa Catalina Mountains (Fig. 5), in the vicinity of Mount Lemmon and Mount Bigelow (Waag, 1968), weakly foliated and lineated medium-grained augen gneiss concordantly intrudes the basal part of the metasedimentary carapace that consists of younger Precambrian Apache Group rocks, there converted to phyllite, amphibolite schist, quartz-sericite schist, quartzite, marble, and metaconglomerate. These are overlain (still within the metamorphic carapace) by marble, calc-silicate, and phyllitic rocks derived from Paleozoic formations. The carapace generally rests in low-angle contact on either (1) medium-grained quartz monzonitic augen gneiss of Tertiary(?) age or (2) moderately deformed coarse-grained augen gneiss derived from 1.4- to 1.5-b.y.-old porphyritic quartz monzonite (Banks, 1977).

In Happy Valley east of the Rincon Mountains (Fig. 5), the rocks of the metasedimentary carapace

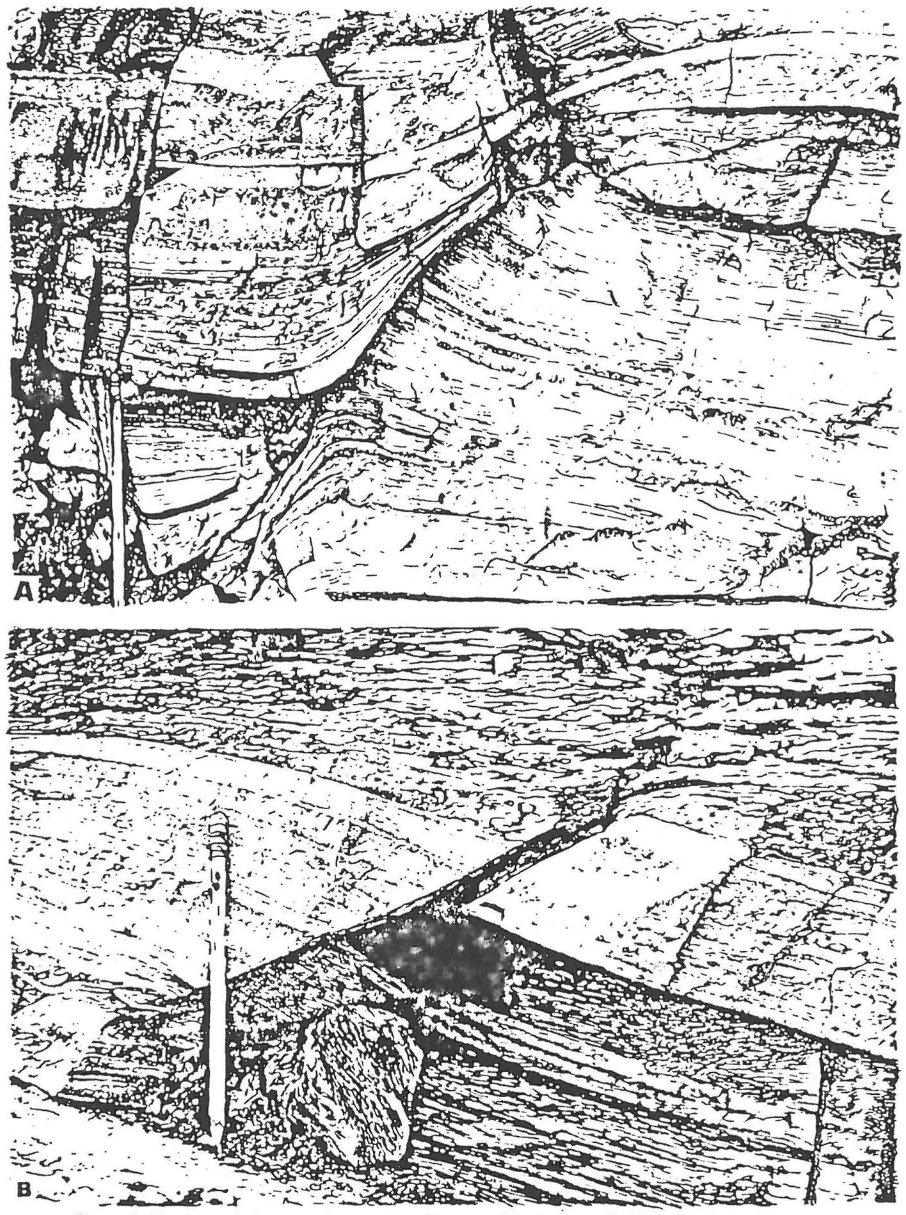


Figure 9. Tracings of photographs. (A) Ductile normal fault zone. (B) Normal-slip fault.

are exposed over large areas (Drewes, 1975). They include unequivocal metamorphosed Apache Group and lower Paleozoic formations. Upper Paleozoic formations also appear to be present in great volume but are harder to distinguish. Dominant and/or distinctive rock types include quartzite, marble, calc-silicate and marble sequences, phyllite, and minor flattened pebble metaconglomerate. These rocks rest in concordant, low- to moderate-dipping contact on Precambrian granitic basement rocks thought to be deformed 1.4- to 1.5-b.y.-old granodioritic quartz monzonitic rocks, and 1.55- to 1.65-b.y.-old granite (Drewes, 1975). Although the metamorphic carapace of the Santa Catalina Mountains rests directly on a vast thickness of penetratively lineated and foliated augen gneiss, in Happy Valley 35 km to the southeast the basement is generally unlineated and gently to moderately foliated. The rocks of the metamorphic carapace are separated from granitic basement by a thin (<20 m) zone of microbreccia, mylonite, and mylonitic gneiss. Thin tectonic slices (<10 m) of the mylonitic gneiss occur from place to place within the lowermost part of the metasedimentary carapace. On the mesoscopic scale, the base of the metasedimentary carapace is planar and strictly concordant with the foliation in the gneiss immediately below. Viewed macroscopically, the surface is smooth and gently warped into systematic upright antiforms and synforms. Locally, the surface is deformed into surprisingly tight antiforms that project upward in almost tent-like fashion. The vertical relief on such sharp, tight structures is at least 10 to 20 m. At many locations in Happy Valley, granite gneiss below the contact displays tight to isoclinal recumbent folds.

Except for the Rincon–Santa Catalina–Tortolita complex, the characteristic features of a metamorphic carapace are seldom exposed in southern Arizona. In the Tortolita Mountains, metasedimentary carapace rocks derived from Apache Group and lower Paleozoic rocks are found in the west-central part of the range (Fig. 5), where they rest in low-dipping contact on lineated, foliated, fine- to medium-grained quartz monzonitic augen gneiss. Homoclinal but internally deformed quartzite of the Pinal Schist is plated onto fine- to medium-grained quartz monzonitic augen gneiss in Jackson Mountain (Fig. 2). In the Coyote Mountains, highly foliated and attenuated quartzite, possibly Cambrian Bolsa quartzite, rests concordantly on thick, lineated, pegmatitic intrusions, which in turn cap foliated and lineated leucocratic augen gneiss. The foliated, folded, and lineated metasedimentary and metavolcanic rocks (Mesozoic?) in the Sierra Blanca are not considered strictly "metamorphic carapace," for they do not clearly rest on a discrete concordant augen gneiss substructure. Rather, they are invaded concordantly and discordantly by pegmatite, alaskite, quartz monzonite, and biotite granodiorite, which are foliated and lineated. Anderson and others (this volume) describe similar relations in northern Sonora.

Mesoscopic Structures

Rocks of the metasedimentary carapace, wherever found, display very distinct structural characteristics (Davis, 1975; Schloderer, 1974; Waag, 1968; Frost and Davis, 1976; Frost, 1977). Strongly overturned to recumbent folds are ubiquitous. Protoliths for the phyllite, schist, marble, and quartzite have been deformed by transposition into tectonites characterized by intrafolial, commonly rootless, tight to isoclinal folds (Figs. 10A, 10B, 10C). Depending on mechanical characteristics of the original rocks, the folds may form by passive flow, passive slip, and/or flexural flow. The passive folds typify homogeneous domains of quartzite or marble. The most outstanding flexural forms occur in interlayered calc-silicate and marble. Where the rocks are dominantly marble and contain only thin brittle calc-silicate or quartzite struts, the "competent" layers are typically distended, attenuated, and transposed (Figs. 10C, 10D, 10E). Fragments of the originally continuous layers form boudins, isolated fold hinges, and assorted tectonic inclusions in the marble matrix (Fig. 10F). In the plane of foliation and lithologic layering of calc-silicate and marble, gash fracturing and orthogonally disposed mineral lineation are locally strongly developed (Fig. 10G).

Conglomerates within the metamorphic carapace are transformed into subhorizontally flattened quartzite-pebble units. The flattened pebbles are typically profoundly elongated parallel to the lineation in underlying augen gneiss. Where the pebbles are flattened but not elongate, underlying cataclastically deformed augen gneiss is foliated but not lineated. In the Tortolita Mountains, Davis and others (1975) measured the orientation and dimensions of 86 flattened elongate quartzite pebbles within Barnes metaconglomerate and compared these to undeformed clasts at a nearby locality in the Santa Catalina Mountains (Fig. 11). Axial ratios were computed to be 9:2:1, the plane of flattening is subhorizontal, and the direction of elongation (N58°E) is identical to that of the adjoining augen gneiss.

Within the metasedimentary carapace, foliation and lithologic layering are generally shallow dipping and strongly developed. Axial planes of tight to isoclinal, overturned to recumbent folds are generally parallel to foliation and layering. The folds are commonly reclined relative to layering and foliation. Seen on the scale of large single outcrops, the foliation and layering are marked by pinch-and-swell and boudinage, with graceful gentle changes in attitude. Predictably, boudinage and pinch-and-swell are best developed where rocks of contrasting ductilities are juxtaposed.

Stratigraphic Relationships

Individual formations within the metamorphic carapace are arranged in normal stratigraphic order, but they are generally tectonically thinned or locally thickened. Specific estimates of the change in thickness are very difficult to make because of the uncertainties inherent in correlating the strongly deformed lithotectonic units with southeastern Arizona stratigraphy. In many places in Happy Valley, the thinning appears to exceed 75%.

Thinning and thickening within the younger Precambrian and Paleozoic sequences have been achieved by passive flowing, including transposition (Frost, 1977) of units rendered ductile during the deformational process. On the mesoscopic scale, the mode of thinning is explicitly displayed in the stretched-flattened pebble metaconglomerates, passively folded marble and calc-silicate rocks, transposed schists, and dismembered, attenuated quartzite layers in ductile matrix. Detailed mapping reveals interesting macroscopic adjustments to the "thinning process," mainly a heterogeneous distribution of lithologic units of contrasting mechanical properties. Mapping by Plut (1968), Drewes (1975), and G. H. Davis and others (in prep.) in Happy Valley has revealed that the massive quartzites of the Apache Group and Cambrian Bolsa Quartzite only locally rest (in parautochthonous contact) on basement. More commonly, these quartzites have been faulted out entirely, and marbles inferred to be of early to mid-Paleozoic age rest directly on basement. The map pattern of such relations demonstrates lateral movement and concentration of ductile materials as a response to thinning, but carried out so that young rocks are always moved over older rocks.

Kinematic Significance

The structural fabric of the metasedimentary rocks is intimately coordinated with that of the underlying augen gneiss. Fabrics of both structural units are marked by profound flattening perpendicular to subhorizontal layering and foliation (see Fig. 2) and by extension parallel to lineation (see Fig. 8B) as denoted by boudinage, deformed pebbles, and the ductile normal fault zones. Davis and others (1975) analyzed structures in the Tortolita Mountains and showed that the lineated surfaces within fine- to medium-grained quartz monzonite augen gneiss accommodated vertically directed flattening and profound east-northeast extension; they also showed that the fold and stretched-pebble fabric in the immediately adjacent metamorphic carapace formed as a response to the same deforming process. Critical to that analysis was recognition of the coordinated nature of the augen gneiss fabric to

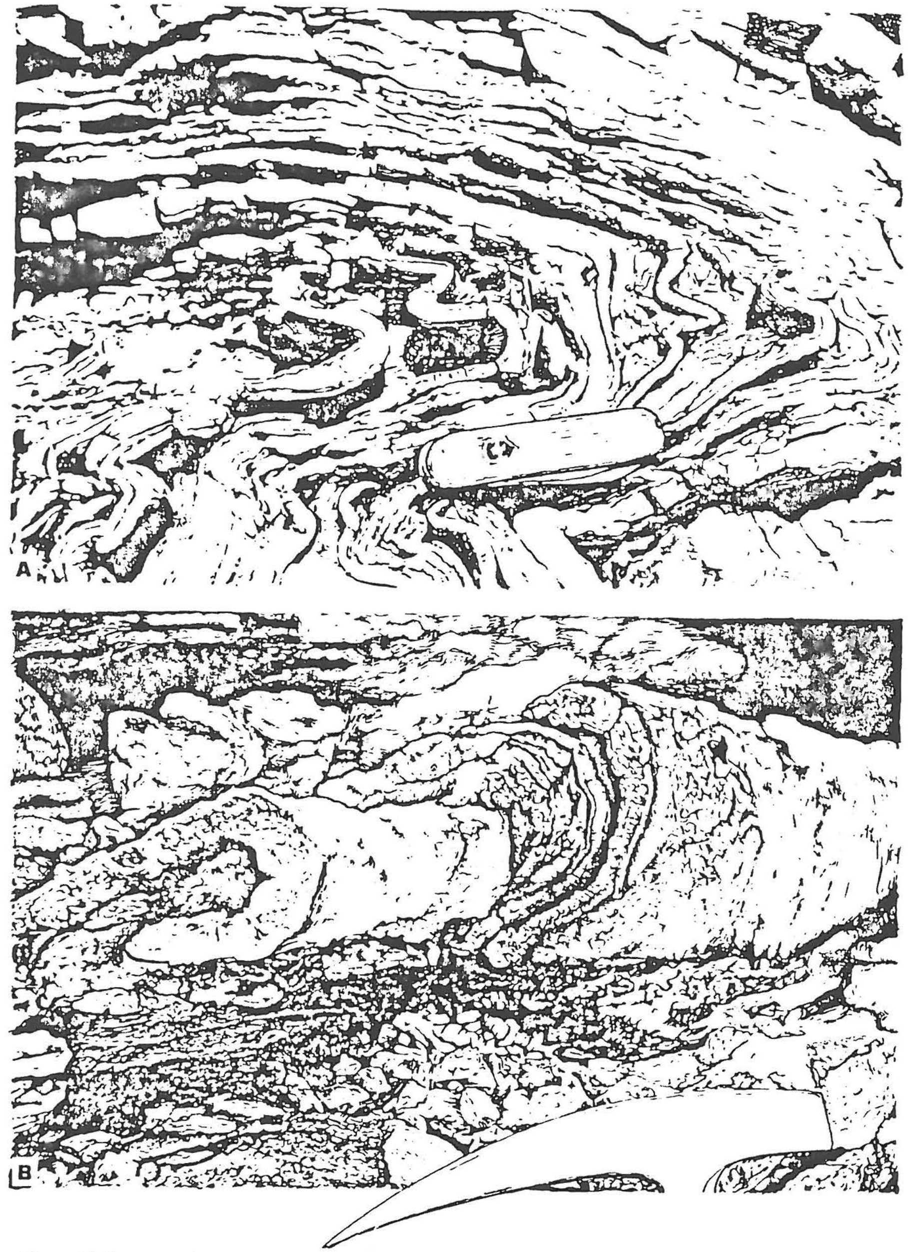
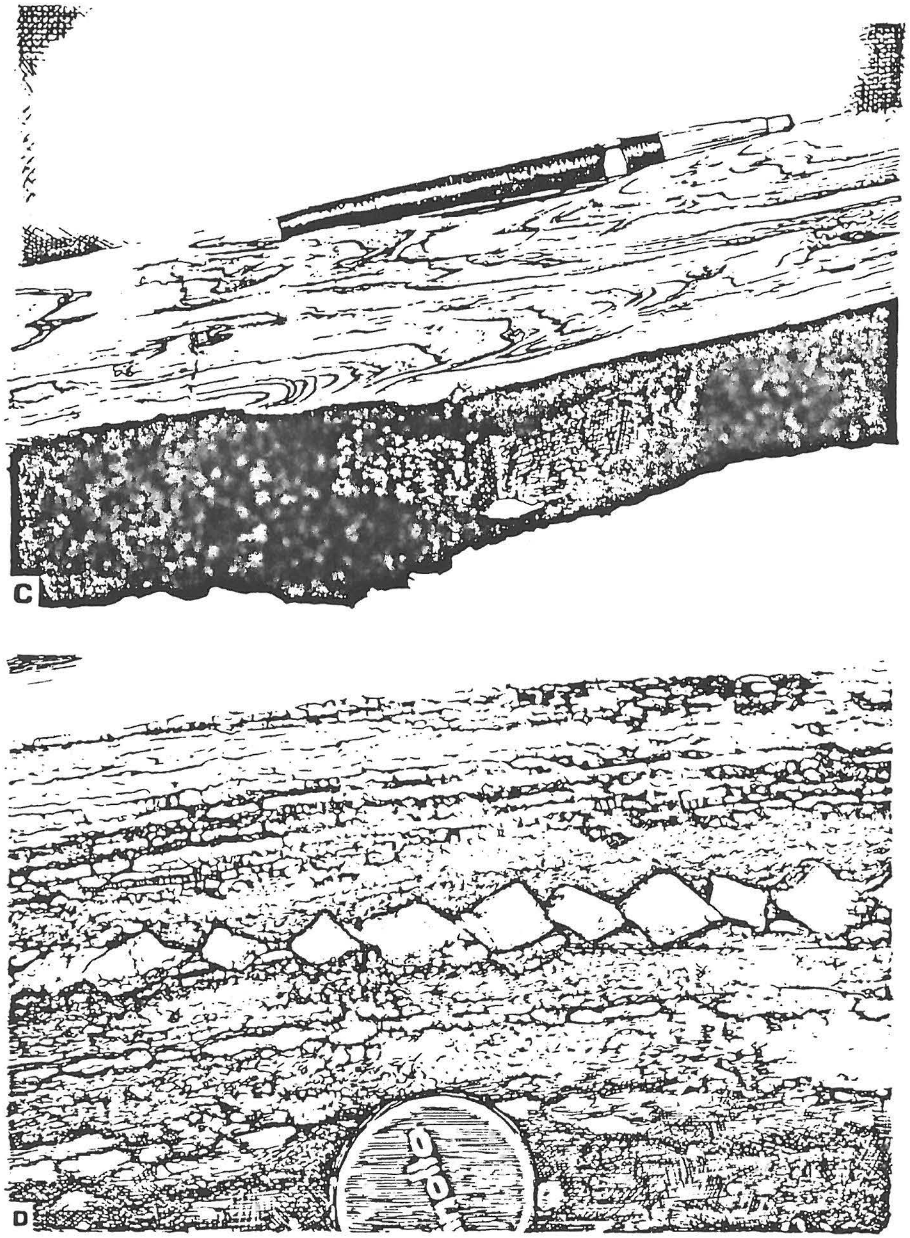


Figure 10. Structures in metamorphic carapace. (A, B, C) Folds in calc-silicate–marble sequences. (D) Lozenge-shaped boudins (traced from photo taken by Eric G. Frost). (E) Tectonic inclusions derived from folded, attenuated calc-slicate and quartzite layers. (F) Boudin of calc-silicate rock in marble. (G) Penetrative gash fractures. (Continued on following three pages.)



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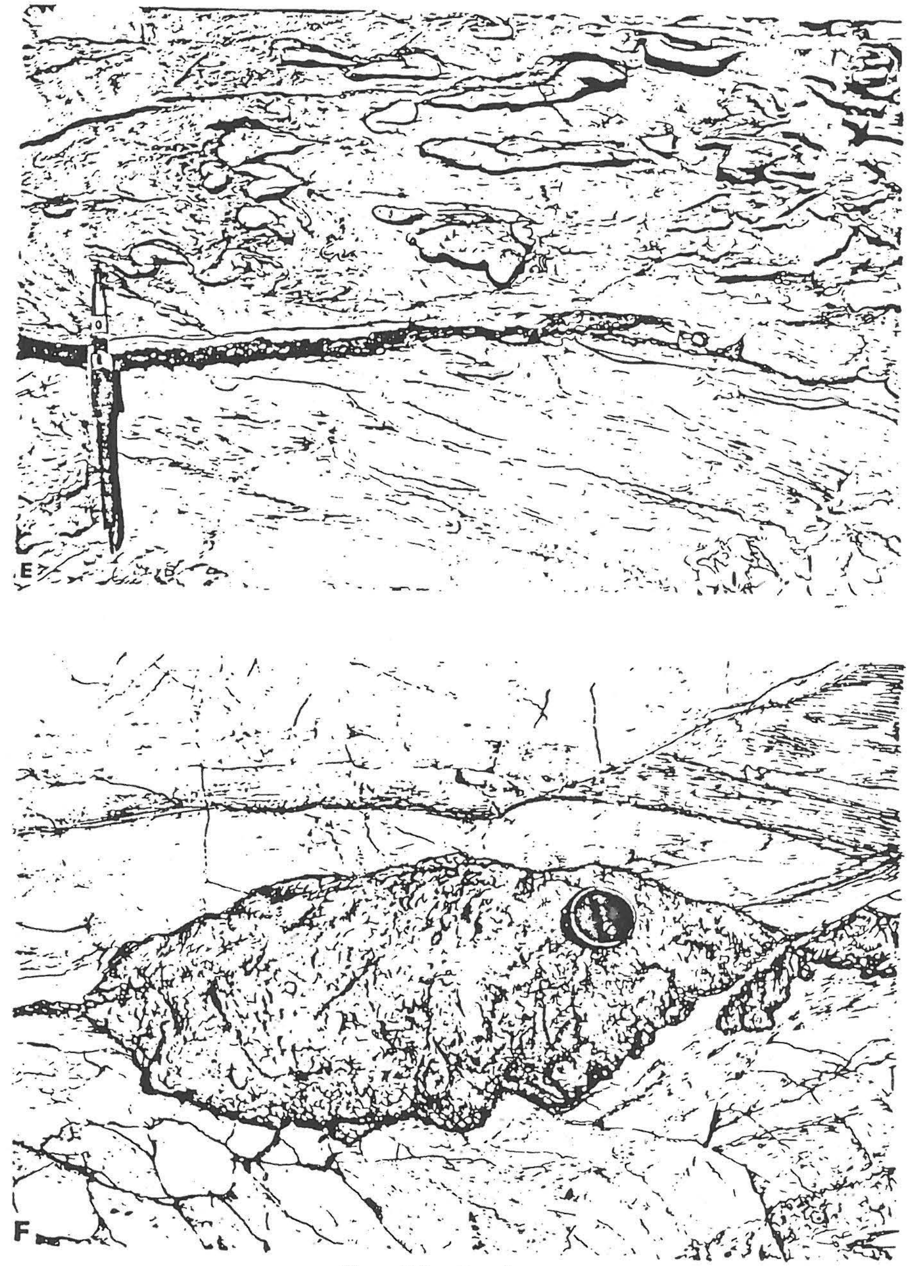

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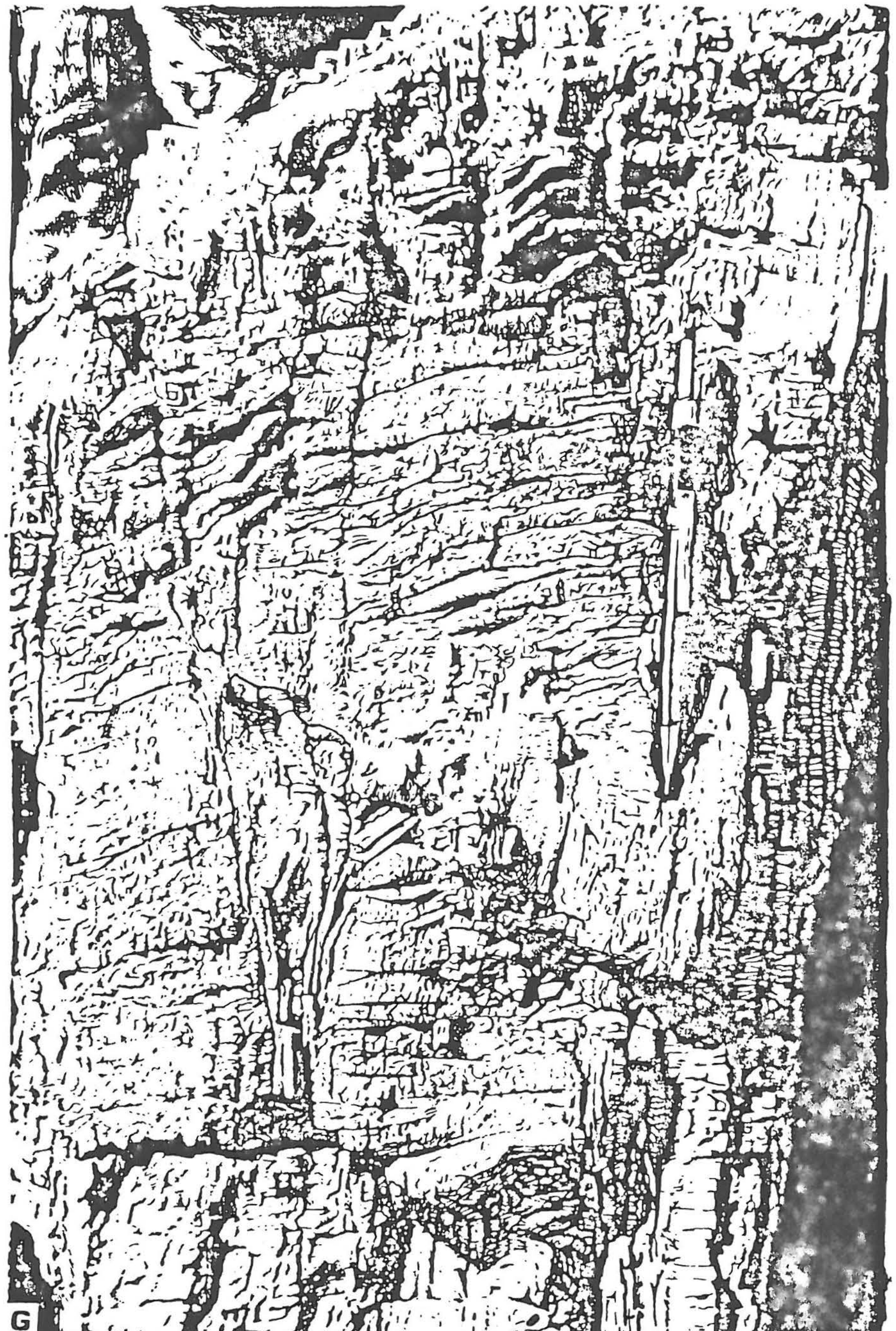

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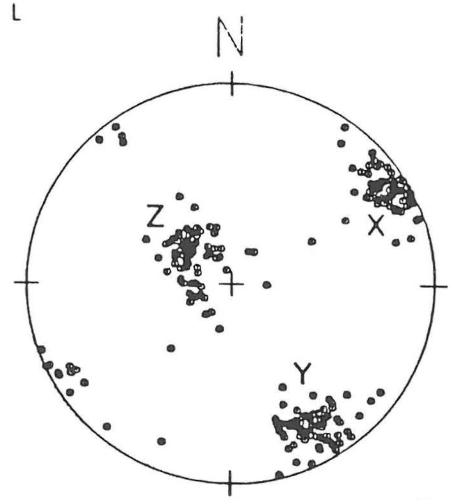


Figure 11. Lower-hemisphere equal-area projection showing long (X), short (Z), and intermediate (Y) axes of 86 deformed quartzite pebbles in the Tortolita Mountains (from Davis and others, 1975).

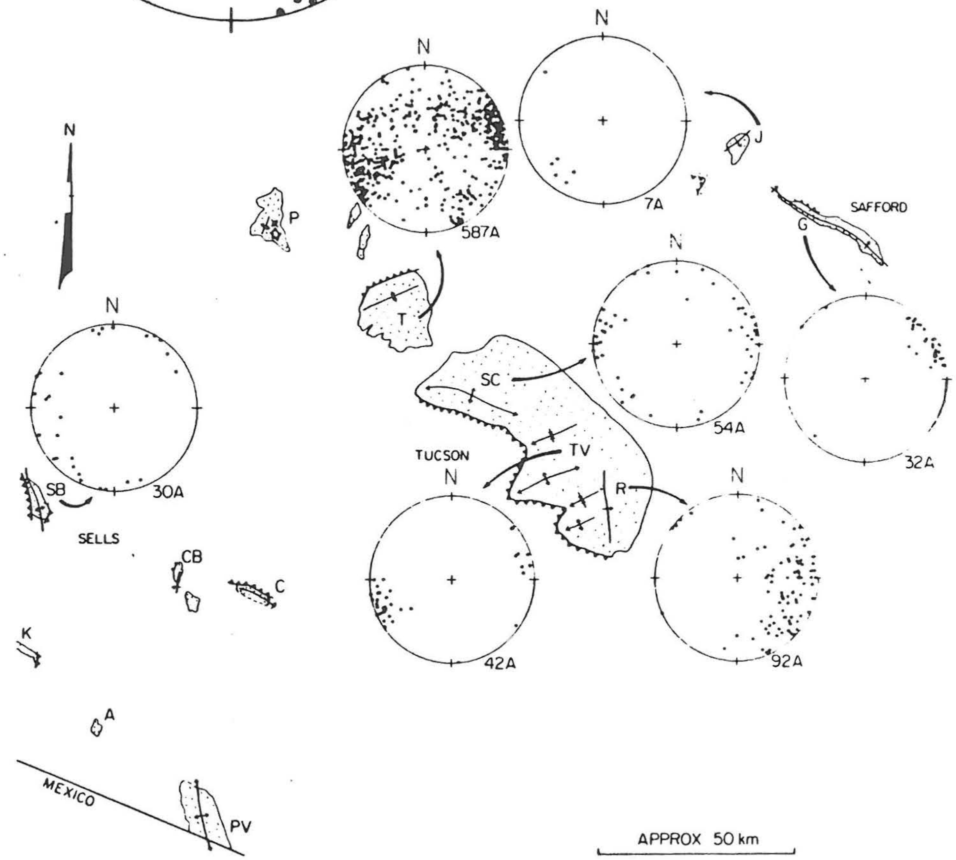


Figure 12. Map showing distribution of metamorphic core complexes in southern Arizona and lower-hemisphere projections of orientations of axes in metamorphic carapace (R, SC, T, J, and SB) and mylonitic schist (G, P, T, and TV).

that of the metamorphic carapace: (1) the modal long axis of pebbles is strictly parallel to lineation in adjacent augen gneiss, (2) the fold axes in the metamorphic carapace parallel the lineation orientation, and (3) ductile normal faults are orthogonal to lineation. On the basis of the strict compatibility of the normal-slip ductile and brittle faults to inferred principal strain directions inferred from the stretched-pebble data, the "stretching" of the quartzite pebbles and the development of normal-slip faults were interpreted as the early and late stages, respectively, of a ductile-to-brittle continuum of extensional deformation.

Fold axes in rocks in the metasedimentary carapace are generally difficult to evaluate in regard to slip-line direction. The fundamental problem is establishing for certain that specific asymmetric folds are indeed first-order folds within the transposed sequences. Furthermore, the fold axes in the metasedimentary rocks generally display a broad range of orientation within the plane of slip or flow (Fig. 12). Although details of contrasting interpretations of kinematics and dynamics could be cited, I believe it is most important to emphasize that (1) within the array of variably oriented overturned to recumbent folds, reclined folds are commonly the preferred mode, (2) the axes of reclined folds tend to be coaxial with lineation in ductile rocks like marble, (3) mineral lineation in marble is essentially orthogonal to penetrative gash fracturing in mechanically suitable lithologies in the metamorphic carapace, and (4) lineation in underlying augen gneiss is parallel to that in marble within the metamorphic carapace.

The above observations and inferences are consistent with the interpretation that both the augen gneiss and metamorphic carapace were affected by flattening and extension and that these processes resulted in profound thinning through flow of the ductile, originally bedded, carapace strata. Flow in the metamorphic carapace was parallel to mineral lineation, and during progressive deformation fold hinges rotated partly or wholly into alignment with the flow direction. The degree of rotation was partly related to the mechanical properties of the deforming sequence.

DECOLLEMENT ZONES

A decollement separates unmetamorphosed cover rocks above from tectonites of the core and/or carapace below. The surface separates rocks of remarkably contrasting deformational styles. Most often in southern Arizona examples, the decollement marks the top of augen gneissic quartz monzonitic core rocks and the base of unmetamorphosed cover; metamorphic carapace rocks are generally absent (Figs. 13, 14). Below the decollement surface, a "decollement zone" is usually conspicuous and consists of a distinctive crudely tabular zone of crushed and granulated but strongly indurated fine-grained mylonitic rocks (mylonite, mylonitic gneiss, microbreccia, and chlorite breccia). The decollement zones are distinctive in their physiographic and structural expression, for the mylonitic rocks form resistant benches or cliffs that are capped by planar upper surfaces (Figs. 13, 14). These so-called decollement zones occur on one or two flanks of the gneissic cores of individual complexes (Fig. 2). Where metamorphic carapace is present, the decollement separates carapace from cover, and a resistant decollement zone of intensely deformed metasedimentary rocks may cap the top of the carapace.

The decollement zones commonly display striking "younger on older" fault relations involving tens to hundreds of metres of stratigraphic separation. They typically separate orthogneisses, derived in part from Precambrian rock or from unmetamorphosed upper Paleozoic or Mesozoic or Tertiary deformed cover rocks. This array of structural and petrologic characteristics has prompted many workers to interpret them as thrust faults (Thorman, 1977; Drewes, 1978b).

The decollement zone in the Santa Catalina and Rincon Mountains, known locally as the Catalina fault, crops out along a sinuous trace, tens of kilometres in length, on the south and west flanks of the

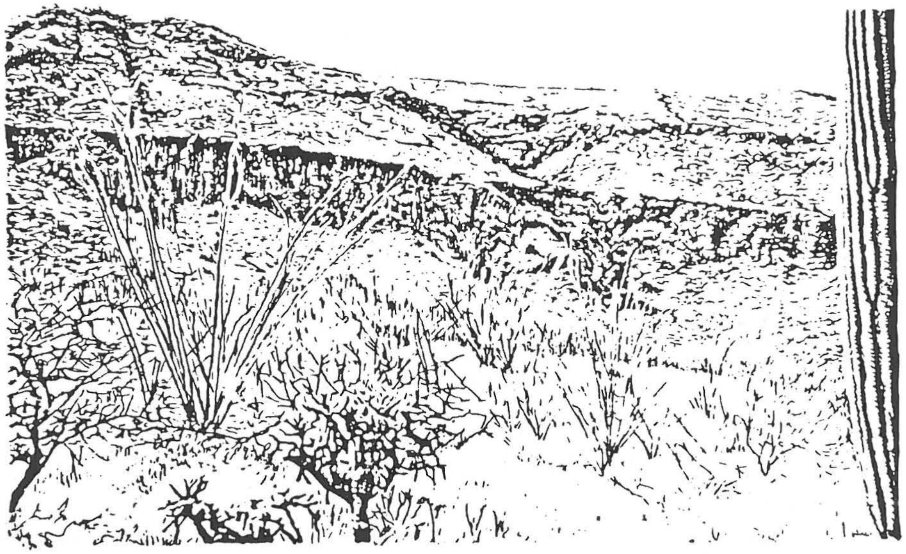


Figure 13. Decollement zone (resistant ledge) in the southern Rincon Mountains. Upper planar surface of decollement zone is the so-called Catalina fault. Overlying the surface is the Precambrian Rincon Valley granodiorite and unmetamorphosed Permian strata.

complex (Figs. 2, 5). The decollement zone separates augen gneiss (below) from a variety of deformed but generally unmetamorphosed cover rocks, including Paleozoic limestone, sandstone, and shale, Mesozoic shale, dolomite, and conglomerate, and Oligocene-Miocene red beds (Fig. 14). The dip of the decollement zone is generally less than 15° or 20°. The only exposures are commonly confined to the pediment-mountain interface (Fig. 3); at no place is the zone known to crop out at a high level on the mountain flank.

Viewed at the scale of the complex, the decollement zone forms a smoothly arcuate surface conforming to the macroscopic structural geometry of the augen gneiss. Viewed at the mesoscopic scale, the zone may be concordant or discordant to foliation in underlying augen gneiss (Fig. 4A).

Although the contact of the decollement zone with the overlying deformed cover rocks is sharp and smooth, the contact with the underlying augen gneiss is generally gradational and ill defined. The thickness of the zone varies from several metres to several tens of metres.

In outcrop, rocks of the decollement zone are brown to brownish-green, chloritic, fine-grained mylonites. They are pervasively overprinted by shattering along closely spaced fractures. Where strongly foliated, rocks of the decollement zone are deformed by kink folding. Within the highly deformed rock suite, lineation and augen gneissic foliation fabrics can be recognized, but these are generally masked by the mylonitization. It seems evident that the rocks of the decollement zone were produced at the expense of already lineated coarse-grained augen gneiss, fine- to medium-grained augen gneiss, and pegmatites and aplites.

The internal structure of the decollement zone in the Santa Catalina–Rincon complex is highly variable. In places, the zone shows homoclinal foliation, with or without northeast-trending lineation. In other places, for example, on the northwest flank of Tanque Verde Mountain, the relict lineation and foliation are systematically rotated.

Decollement zones of this type are by no means restricted to the Santa Catalina and Rincon Mountains. Such zones crop out along the north end of the Tortolita Mountains, at the southeastern

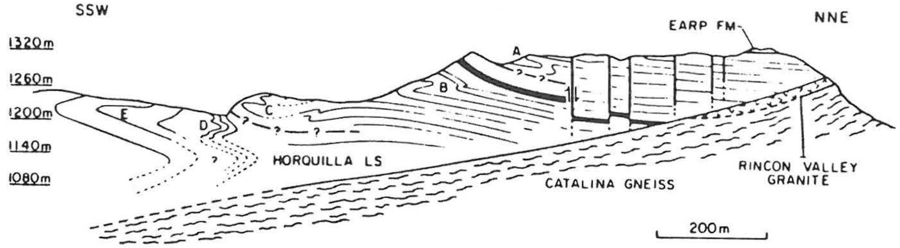


Figure 14. Cross section of decollement zone and upper plate rocks on the southern flank of the Rincon Mountains (from Davis and others, 1974).

end of the Picacho Mountains, along the north end of the Coyote Mountains, and on the west flank of Sierra Blanca (Fig. 2). Also, one exposure of the otherwise concealed decollement zone on the east flank of the Pinaleno Mountains was discovered (Fig. 41).

At the north end of the Tortolita Mountains, the decollement zone strikes east-west, dips gently north, and separates undeformed (1.4- to 1.5-b.y.-old) porphyritic biotite quartz monzonite and normal Pinal Schist on the north from cataclastically deformed porphyritic biotite quartz monzonite of Tertiary(?) age on the south. Cataclasis has transformed the quartz monzonite to augen gneiss and schist.

A low-dipping chloritic, mylonitic layer defines a decollement zone at the southeastern end of the Picacho Mountains. The zone marks the deformed upper surface of exposures of medium-grained quartz monzonitic augen gneiss. The upper surface of the zone is planar; it separates the cataclastically deformed augen gneisses from overlying Tertiary (23-m.y.-old) basalt. The interface of contact is itself a fault contact, marked by slickenslided surfaces that overprint the mylonitic fabric of the decollement zone. Slickensides plunge gently southeastward. The decollement zone grades downward over a 10- to 20-m interval into typical lineated, foliated augen gneiss.

In the Coyote Mountains southwest of Tucson (Fig. 2), a decollement zone expressing physical properties identical to the Catalina fault crops out along the northern pediment of the mountain. Its strike is east-west, and it dips northerly from 20° to 35°. Brown-green cataclastically deformed, chloritic, resistant rocks occur on the "footwall" of the zone, including biotite quartz diorite, marble and calc-silicate rocks of a metasedimentary carapace, and leucocratic granitic and quartz monzonite augen gneiss. Overlying the decollement zone, in sharp planar contact, are red beds inferred to be Oligocene-Miocene in age.

On the west flank of Sierra Blanca, unmetamorphosed to slightly metamorphosed Permian sedimentary strata are separated from the crystalline core rocks by a pronounced, gently west-dipping decollement zone. Again, the rock of the decollement zone is chloritic, highly fractured but highly indurated mylonite, mylonitic gneiss, and microbreccia. Its thickness is about 10 m, although the base of the zone grades into underlying Mesozoic schists and biotite granodiorite.

UNMETAMORPHOSED BUT DEFORMED COVER ROCKS

The so-called cover rocks that overlie the augen gneissic core and metamorphic carapace of the core-complex terranes are interesting and instructive in their gross configuration and internal structure. For the most part, the cover rocks directly overlie well-developed decollement zones of the type already described. Cover rocks above the decollement zone (Catalina fault) in the Santa Catalina and

Rincon Mountains form thin plates that include slices of Paleozoic, Cretaceous, and Tertiary formations (Fig. 5). They include shattered Precambrian granite as well. Along the front of the Santa Catalina Mountains, the zone separates augen gneiss from the Oligocene-Miocene continental red beds, notably mudstone, siltstone, sandstone, and conglomerate. Clasts within the lower part of the sequence bear no resemblance to rocks within the core complex itself, except in the upper beds where the "influx" of clasts of limestone and lineated augen gneiss document an unroofing of the complex (Pashley, 1968).

Miocene beds locally rest on the decollement zone on the south and west flanks of the Rincon Mountains; additionally, Precambrian, Paleozoic, and Mesozoic strata lie atop the zone as well (Drewes, 1978a). Completeness (incompleteness?) of stratigraphy in these sections is highly variable. Sedimentary rocks on the west side of the Rincon Mountains within the Saguaro National Monument (East) consist of limestone, dolomite, and shale of Permian age, as well as remnants of Mississippian(?), Pennsylvanian, and Cretaceous formations. Along the southeast flank of the Tanque Verde antiform, Cretaceous shale and limestone, with interbedded siltstone, dolomite, and limestone conglomerate, lie directly on the decollement. The thickness of the sheet is less than 90 m. Sedimentary rocks near Colossal Cave on the south side of the Rincon Mountains form a sheet approximately 150 m thick that rests on the decollement zone. The rocks consist of limestone interbedded shale and include formations of Cambrian through Permian age. At the southeasternmost corner of the Rincon Mountains, a 75-m sheet of Paleozoic rocks rests on the decollement zone. Although formations from Cambrian to Permian time are represented, the thickness of the sequence is less than 10% of the full Paleozoic section exposed in the Whetstone Mountains only 20 km south.

Structures in such deformed cover rocks are described in some detail in Davis (1975), Liming (1974), and Davis and Frost (1976). The individual sheets of Phanerozoic strata range from about 40 to 120 m in thickness. Strata within each sheet are generally unmetamorphosed, except near the base where limestones are commonly marbleized over thicknesses of 10 m or so. However, in the Martinez Ranch area, low-grade marbles can be found 60 m above the decollement zone. The sheets of deformed cover rock are grossly concordant with the underlying decollement zone, but in detail it can be demonstrated that discordance exists between layering in the sheets and foliation in the underlying gneisses.

Overturned asymmetric folds, detached isoclinal folds, and unbroken cascades of recumbent folds characterize the sheets of the Paleozoic and Mesozoic cover rocks (Fig. 15). Such folds by no means characterize normal Paleozoic-Mesozoic sedimentary country rock in southeastern Arizona (Davis, 1979). Most of the folds are transitional between ideal parallel and ideal similar folds and, thus, are characterized by some hinge-zone thickening (Davis, 1975). The scarcity of axial-plane cleavage, the abundance of bedding-plane cleavage, and the obvious influence of layering on the morphology of folds indicate that the folds evolved through slippage between layers and flow within layers.

Cover rocks in the Coyote Mountains include Mesozoic volcanic rocks and Oligocene-Miocene red beds. The small patch of cover at the southeast end of the Picacho Mountains is Miocene basalt. In the Sierra Blanca, cover rocks are folded upper Paleozoic strata.

INTERPRETIVE REMARKS

The metamorphic core complexes in southern Arizona are fundamentally products of high-temperature extensional deformation, regional in extent, superseded by moderately ductile to moderately brittle tectonic denudation. During the disturbance, rocks in the augen gneissic core and metamorphic carapace were produced by profound ductile through brittle extension and flow in the direction of mineral lineation, accompanied by vertical flattening and transposition. Tectonites developed at the expense of Precambrian crystalline basement; Phanerozoic quartz monzonite, granodiorite, and granite bodies that were emplaced both before and during the disturbance; and younger Precambrian, Paleozoic, and Mesozoic sedimentary rocks that were metamorphosed during the disturbance. Folds rotated in the plane of flow during deformation; this produced a broad range of axial trends in general, except in zones of mylonitic schist where movement was localized and fold hinges rotated into strict parallelism with lineation. Decreasing ductility through time is recorded by features such as folding of lineation and refolding of folds along ductile normal fault zones and superimposition of brittle normal-slip faults upon the ductile faults.

Evolution of the decollement zones largely postdated the development of foliation and lineation fabric in the augen gneiss and metamorphic carapace. Progressive cataclastic granulation of augen gneiss or metasedimentary rocks in the decollement zones was accompanied by rotation of the original

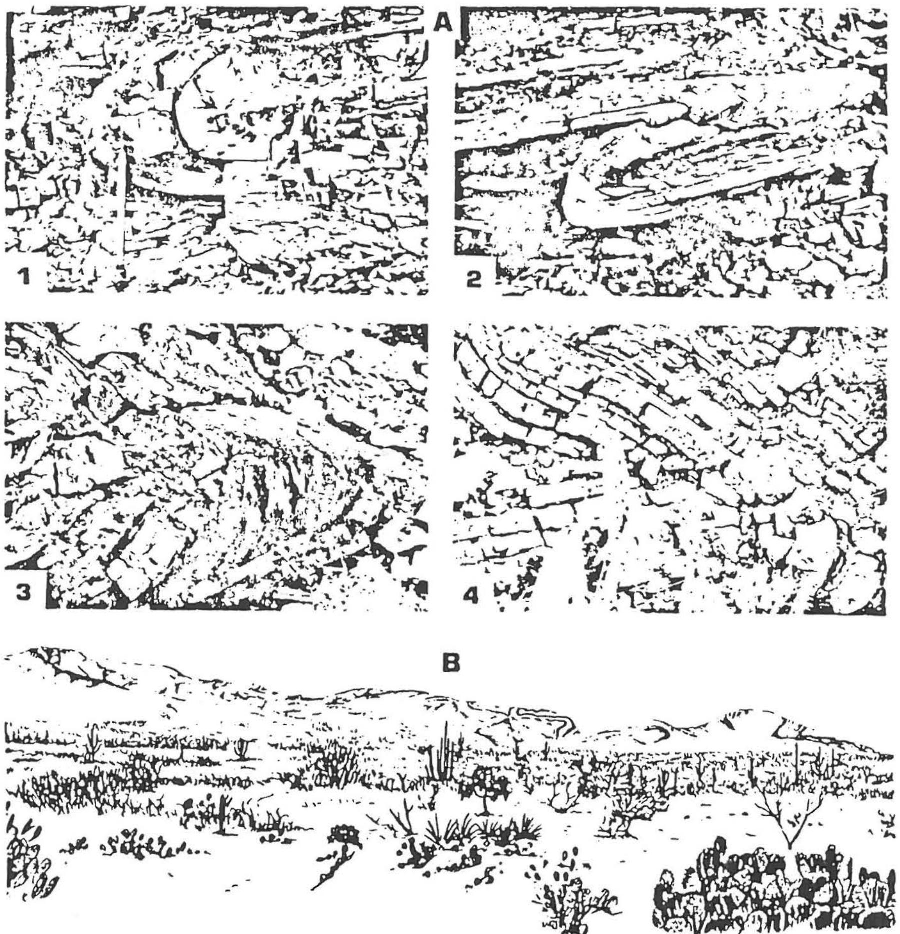


Figure 15. (A) Recumbent and overturned folds in mudstone: 1, dolomite; 2, 3, limestone; 4, Cretaceous strata on the south flank of Tanque Verde Mountain. (B) Overturned to recumbent large-scale folds in Paleozoic strata on the south flank of the Rincon Mountains (tracing of photograph from Pashley, 1966). Cliff in dark shadow at left is that portrayed in Figure 13.

foliation. The foliation in turn controlled the formation of late-stage kink folds. Brittle structures in the plates of unmetamorphosed but deformed cover rocks mostly postdate development of the fabric of the augen gneiss core and metamorphic carapace. Folds and faults formed in the cover rocks during gravity-induced gliding and listric normal faulting.

Ductility contrast influenced the structural deformation as viewed at the outcrop scale (Fig. 16). Likewise, ductility contrast was the major influence in predetermining the nature, relative position, and configuration of the major structural units within the complexes (Fig. 2), that is, core, carapace, decollement zone, and cover.

The array and distribution of structures in the metamorphic core complexes have proved difficult to interpret. Structures in the cover rocks have been traditionally interpreted as products of regional thrusting, but recent work suggests that the structures actually evolved through denudation in middle to late Tertiary time (Davis, 1973, 1975; Coney, 1974; Compton and others, 1977; Davis and Frost, 1976). Structures in the augen gneiss of the core rocks and the metamorphic carapace have been interpreted variously as products of overthrusting (Drewes, 1976a, 1978b; Thorman, 1977), diapiric batholithic emplacement (Creasey and others, 1976), and mantled-gneiss doming (Peterson, 1968; Waag, 1968). The presence of metamorphic core complexes in southeastern Arizona, in a region where the presence of regional low-angle overthrusts has not been proved, weakens the possibility of a dynamic linkage between thrusting and the penetrative structures found in core and carapace rocks. Although metamorphic core complexes bear a superficial resemblance to mantled gneiss domes and igneous batholithic domes, the application of pure granite tectonic and/or diapiric models fails to explain important details of the structural geology of the complexes.

Structures in the augen gneissic core and carapace rocks can be best understood when the cores are pictured as components of megaboudins imposed on heterogeneous crustal rocks (Davis, 1977a; Davis

Figure 16. Tracing of photograph of boudined and folded schist and aplite that illustrates the influence of ductility contrast on deformational style.

and Coney, 1979). By way of explanation, it is instructive to consider Figure 17, a tracing of a photograph from Ramberg's (1955) paper on boudinage. It is useful to consider the scale of this figure not as 0 to 5 cm, but as 0 to 5 km; to consider the enveloping mica schist as a small-scale analog of the metamorphic carapace; and to consider the upper part of the pegmatite boudin as an analog of the augen gneissic core. The interface between the contrasting rock types should be viewed in this discussion as the surface of great unconformity. If this image and Figures 18 and 19 are used as guides, the mechanical interpretation of core-complex evolution may be more easily understood.

Specifically, the megaboudinage concept may provide a basis for understanding the macroscopic and mesoscopic structures as defined here: (1) *augen gneissic core*—upper part of crystalline basement rocks and pretectonic plutons extended in a way to accommodate syntectonic intrusions; (2) *low-dipping foliation*—formed as a result of moderately brittle to moderately ductile response to extensional deformation and flattening in a zone of finite thickness at outer margin of megaboudined "basement"; (3) *penetrative lineation*—response to stretching of moderately brittle to moderately ductile augen gneissic core; (4) *ptygmatic folds*—expression of flattening of augen gneissic core during extension; (5) *mylonitic schist*—produced by strain concentration localized by shearing of the lithologic contact between moderately brittle crystalline basement and moderately ductile to ductile syntectonic instrusion; (6) *metamorphic carapace*—ductile Phanerozoic layered rocks that were stretched, attenuated, and flattened during the passive-flow accommodation to ever-increasing surface area of contact with underlying augen gneissic core; (7) *intrafolial folds*—perturbations in flow regime at surfaces of interface of layers of contrasting ductilities; (8) *lineation, including stretched pebbles*—reflection of profound plastic extension of highly ductile materials; (9) *gash fractures*—late-stage brittle deformational response to extension; (10) *cascades of recumbent folds*—infilling of necked zones by plastic enveloping matrix; (11) *decollement zone*—granulation and rotation of augen gneissic and metamorphic carapace rocks; viewed as "necking" of the moderately brittle megaboudin; or cataclastic degradation of corner of basement block that before mid-Tertiary time, owing to early Mesozoic and Laramide tectonic movements, was in high-angle contact with layered strata; (12) *cover rocks*—Phanerozoic strata or Precambrian crystalline rocks that, for the most part, were above the level of thermal metamorphism but were nonetheless denuded (Davis and Coney, 1979), both during and after the formation of lineated tectonite.

Although still in development, the megaboudinage concept may provide a new way for assessing relations that have proved to be impossible to explain by more traditional structural mechanisms. For example, decollement zones commonly occur only on one or two flanks of specific complexes. In the Rincon Mountains, the decollement zone occurs on the south and west and caps penetratively foliated and lineated augen gneissic core rocks. On the east side of the Rincon Mountains in Happy Valley, the crystalline igneous rocks are overlain directly by quartzite, calc-silicate rocks, and marble of the metamorphic core complex, *without* an intervening strongly developed decollement zone. Furthermore, the crystalline rocks are unlineated and moderately foliated at best. One interpretation lies in

Figure 17. Tracing of photograph in Ramberg's (1955) paper on boudinage. See text for explanation.

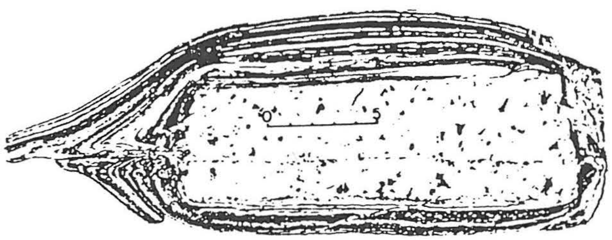
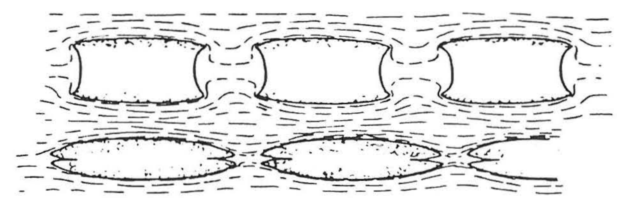


Figure 18. Boudinage characteristics (from Ramsay, 1967). Stippling added to denote regions of likely cataclastic deformation. Decollement zones would preferentially develop at corners of blocks in upper diagram and within necked ends of lensoids in lower diagram.



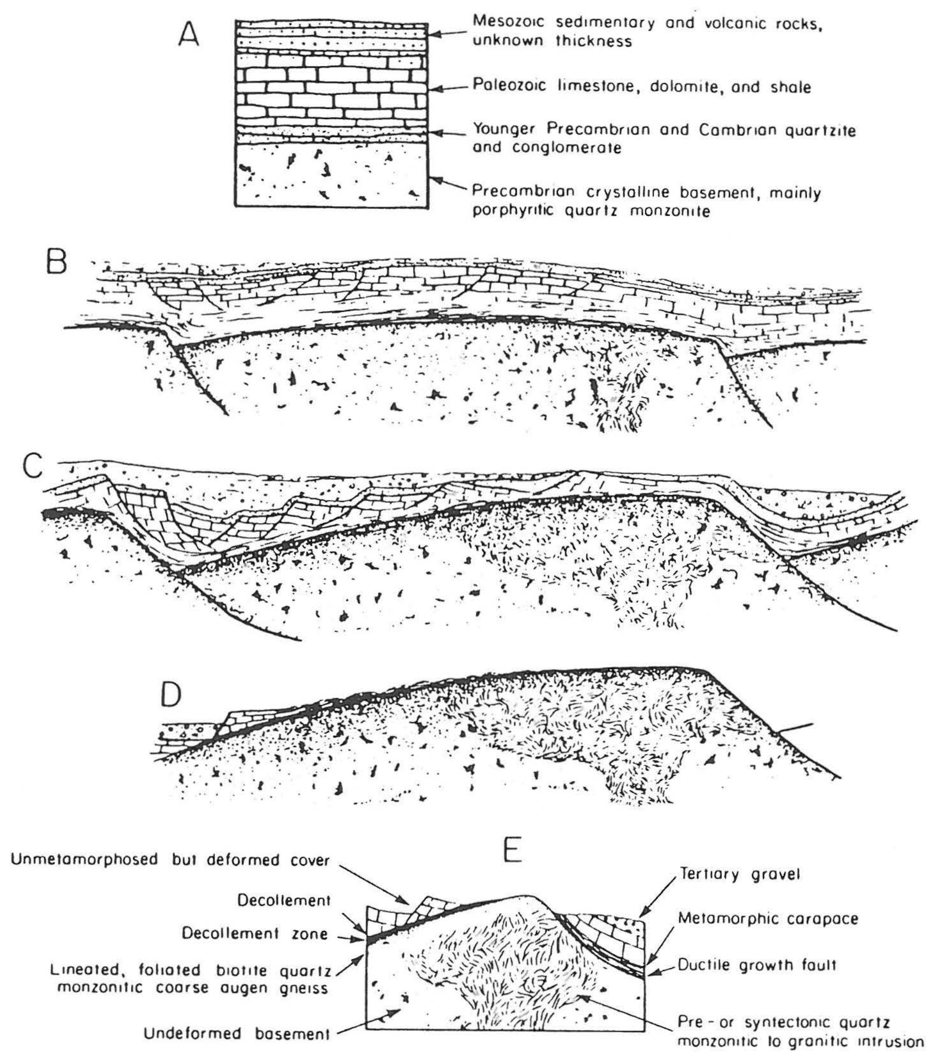


Figure 19. Generalized tectonic reconstruction of the geologic evolution of metamorphic core complexes.

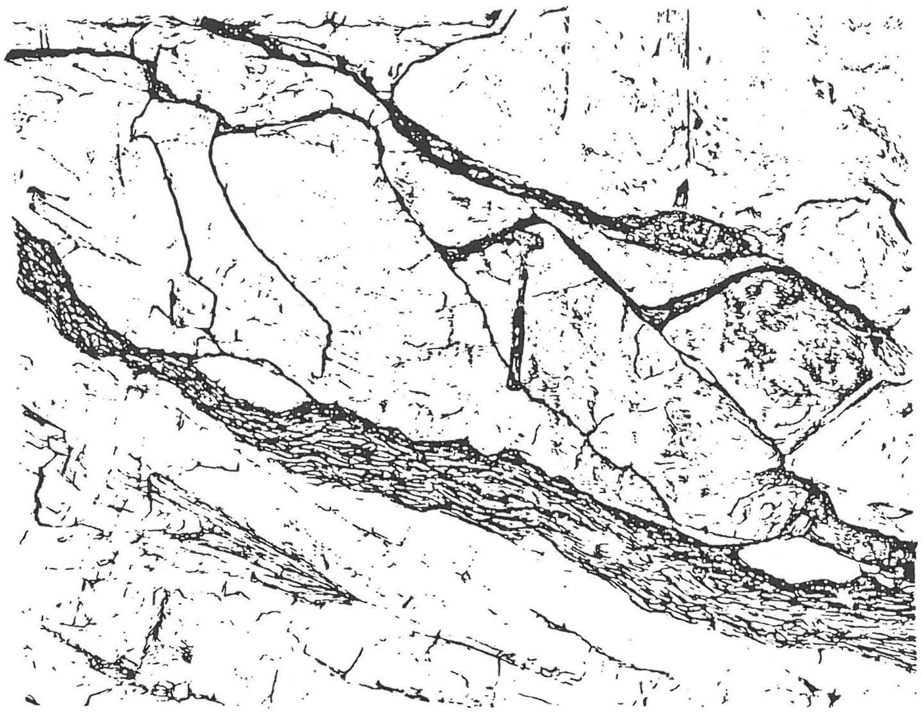


Figure 20. Outcrop-scale example of the ductile plating of a relatively thin lithological layer (white, pinched-and-swelled unit above hammer) onto the footwall of underlying, normal faulted "basement" layer. Quartzofeldspathic gneiss and schist in zone of lineated augen gneiss in the Pinaleno Mountains (see Fig. 41).

considering the so-called Rincon dome as akin to part of a lozenge-shaped megaboudin (Rast, 1956). When considered in this way, an important distinction comes to light: the decollement may coincide with the great unconformity; on the other hand, rocks of the metamorphic carapace (in this example) may *never* have been in depositional contact with the underlying crystalline rocks, but progressively flowed into contact by spreading over the surface of developing normal faults that define the eastern edge of the lozenge-shaped megaboudin (Figs. 19C, 19D, 19E). The unlineated, moderately to poorly foliated crystalline rocks reflect a deeper (and hotter) structural level than those augen gneisses beneath the decollement zone. The normal faulting portrayed in Figure 19 constitutes a new concept in large-scale faulting, one that is termed herein "ductile growth faulting." The relationship shown in Figure 20 is a wonderful outcrop example of this type of "disharmonic" faulting. The term "growth" implies that the area of the surface on which metamorphic carapace materials are stretched and welded to the crystalline basement continually increases during the life of the fault. The faulting reflects a mode by which the surface area of the crystalline basement can be increased during extension. Such faults and associated folds were recognized at the outcrop scale by Davis and others (1975) and termed "ductile normal faults."

This emphasis on megaboudinage is not intended to suggest that each metamorphic core complex is an individual boudin. Rather, the core complexes are partial exposures of a regional system of extensional deformation where properties might be best understood in the context of megaboudinage. Countless styles of boudins have been reported, both by field geologists and experimentalists. The

properties of boudin systems depend on many factors, most important of which are ductility contrast and percent strain. The challenge in further investigations of core complexes will be (1) documenting the great variety of macroscopic strain styles that have evolved diachronously in early to mid-Tertiary time, (2) attempting to establish relations between structure relief and ancient or present topography (Max Crittenden, Jr., 1979, personal commun.), and (3) assessing the superimposition of brittle, mid-Miocene deformation on the pre-existing, ductile extensional systems.

Stretching of basement in a manner akin to megaboudinage took place after Laramide and before basin and range tectonism. Early to middle Tertiary granitic plutons which were emplaced during the crustal stretching were affected by movements that produced penetrative foliation and lineation. Topographic and structural basins created by regional pinch-and-swell were synchronously filled by early to mid-Tertiary continental sediments. Brittle faulting of necked megaboudins prompted the outpouring of mid-Tertiary ignimbrites through rifted basement. High pore-fluid pressure concentrated in the vicinity of decollement zones during the dynamo-thermal event favored near-wholesale denudation of the metamorphic carapace and cover from the culminations and flanks of the megaboudin. Late-stage mid-Miocene listric normal faulting further denuded the cover and heightened the physical distinctiveness of the decollement zones. The stretching and thinning of lithosphere that resulted in metamorphic core complexes led to conditions within the western Cordillera which favored the *collapse* that formed the basin and range structure.

ACKNOWLEDGMENTS

Foremost among those that I would like to thank are present and former students, especially Eric Frost, Brett Liming, John Schloderer, Monte Swan, Terry Budden, Stan Keith, Ken Brook, Chuck Kiven, Bob Varga, Steve Reynolds, Gene Suemnicht, Chuck Kluth, and Stevey Lingrey. I would acknowledge the helpfulness of provocative discussions with geologists who have worked in the Rincon-Catalina-Tortolita terrane, especially Norm Banks, Paul Damon, Harald Drewes, Evans Mayo, and Doug Shakel. I will be forever grateful for the endless hours of discussion with my friends and colleagues Tom Anderson, Peter Coney, and Greg Davis.

Structural investigations in the augen gneissic core zones and metamorphic carapace were supported through National Science Foundation Grant EAR 76-84167. Also, Continental Oil Company, through the help and cooperation of William Rehrig, provided financial aid to support Steve Lingrey as a field associate during the second summer of the project. The Department of Geosciences at the University of Arizona has provided continued financial and logistical support of my structural investigations over the years. I am grateful to the University of Arizona's Bureau of Audiovisual Services for drafting of maps and figures and to David O'Day for his artistic rendering of the illustrations. I extend my appreciation to Cliff Wright and Jim Hardy who served as willing field assistants.

Early drafts of the manuscripts were read and edited by Tom Anderson, Bob Compton, Peter Coney, and Max Crittenden, Jr. To them I am very grateful.

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MANUSCRIPT RECEIVED BY THE SOCIETY JUNE 1, 1979

MANUSCRIPT ACCEPTED AUGUST 7, 1979

Arizona Geological Society Field Trip 3

Road Log and Trip Guide to the Geology of the Northern Flank of the Pinaleno Mountains, Southeastern Arizona

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INTRODUCTION

This segment of the trip on Mylonitic Tectonites and Detachment Faults, southeastern Arizona, is planned primarily to examine the cataclastic rocks on the north side of the Pinaleno Mountains. In order to better understand these rocks one should examine their protoliths, the parent rocks from which the protomylonitic and mylonitic rocks were derived. If weather permits, we will examine several of these rock units, one of which underlies most of the range. To this end we will drive up Swift Trail (Arizona 366), a paved road into Jacobson Canyon at the east end of the range. Because of unknown weather conditions, including possible snow-covered switchbacks, several optional stops are included. Distances given in the road log are in miles between observation points.

The rock units and general geology of the range are described in the accompanying preprint of a paper by me on the Pinaleno Mountains that will be published this spring in volume 13 of the Arizona Geological Society Digest. Therefore, only brief descriptions of rock units are repeated below.

Proceed south from Safford on U.S. 666; turn right at Swift Trail Junction onto Arizona 366; Federal Prison on southeast side of highway. You are crossing the Jacobson-Marijilda Canyons pediment that is cut on Pliocene-Pleistocene fluvial-lacustrine valley fill. The partially preserved benches and dissected, gently sloping mesas northwest and southeast of the highway are degradational features that reflect multiple downcutting phases which have occurred in the Safford Valley during recent time. These surfaces are pediments capped by thin gravel deposits which can be traced into the range to elevations several thousands of feet above the Safford Valley floor. Gravels cap Precambrian crystalline rocks within the range, whereas the Pliocene-Pleistocene valley fill deposits underlie the pediment surfaces away from the mountain front. This transition can be seen at the entrance to Jacobson Canyon.

ROAD LOG AND TRIP GUIDE

(Morning, first day--March 17, 1981)

Stop 1 - Take left turn onto dirt road at first right-hand bend in road at mountain front. Park and walk up Jacobson Creek to Johns Dam, the type locality for the gneiss of Johns Dam.

The gneiss of Johns Dam comprises two rock types that are interlayered on a mesoscopic to macroscopic scale, a porphyritic granitic rock and quartzofeldspathic gneisses. The gneisses (the gneiss of the Pinaleno Mountains) are host rocks into which the granitic material was intruded lit-par-lit on a grand scale. Generally the granitic rock is weakly foliated with coarse-grained K-feldspar phenocrysts imparting an augen structure to the rock. This is an important feature to recognize because nearby this rock is cataclastically deformed into protomylonite. At first glance, the protomylonite appears to have undergone considerable deformation owing to its pronounced augen and fluxion structures when, in fact, the protolith was a rock with augen structure.

1.8 Proceed up highway and take dirt road to right. Roadcuts to this point expose alluvial debris capping pediment cut on gneiss. Park at sharp bend in road with view to northwest and southeast.

Stop 2 - Gneiss of Johns Dam is well exposed along this road. Cataclasis generally is better developed here than at Stop 1. The interlayering of the dark intrusive phase with the lighter colored gneiss is well displayed to the north in Marijilda Canyon. Also, observe the thin gravel cappings on the multiple pediment surfaces cut on both the Precambrian bedrock and Pliocene-Pleistocene valley fill. Several northwest-trending range-front faults can be seen at the mouth of Marijilda Canyon.

0.55 Return to pavement and continue up Jacobson Canyon. Contact between granite of Veach Ridge and gneiss of Johns Dam.

1.0 Angle Orchard road to left - Do not take; continue straight ahead.

0.1 Contact between granite of Veach Ridge and gneiss of Pinaleno Mountains. Exposures for approximately next 5 miles are in steeply dipping gneiss of Pinaleno Mountains. Gravel-capped pediment on right.

Stop 3 - (optional)

0.1 Park cars on north side of highway and walk up road to see folds in gneiss of Pinaleno Mountains. These earlier folds occur throughout the range and are unrelated to the cataclastic event.

Stop 4 - (optional)

2.0 Pull into parking area at hairpin turn - excellent drinking water. Same geological setting as at Stop 3.

3.0 Contact between gneiss of Pinaleno Mountains and granite of Ladybug Saddle. This pluton is well exposed in roadcuts up to the range crest. Small xenoliths are common locally and generally are alined parallel to the foliation of pluton. The granite is believed to be a post-tectonic to anorogenic pluton.

Stop 5 - (optional)

1.5 Take dirt road to right within hairpin left turn (just past houses in Turkey Flat). This abandoned road was originally planned to connect with abandoned road of Stop 2. Approximately first 0.8 mile is in granite of Ladybug Saddle and the remainder in gneiss of Pinaleno Mountains.

Preprint--to be published in Arizona Geological Society Digest, volume 13, 1981.

GEOLOGY OF THE PINALENO MOUNTAINS, ARIZONA: A PRELIMINARY REPORT

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Abstract

The Pinaleno Mountains, a northwest-trending range in southeastern Arizona, are underlain by precambrian rocks ranging in age from 1,700 to 1,100 m.y. old; including gneisses probably of sedimentary and volcanic origin, synorogenic to anorogenic granite plutons, metaquartz latite dikes, and diabase bodies. An Oligocene granite stock intruded Precambrian granite in the Stockton Pass vicinity, on the south side of the range. Farther south, at Greasewood Mountain, a 27 to 23 m.y. old complex dome-flow eruptive center of felsic to intermediate composition rests on Precambrian granite.

The Stockton Pass fault zone, trending west-northwest along the southern flank of the range, records several episodes of movement, beginning with left slip during the Precambrian. Right slip along several of the faults in the zone occurred after emplacement of the Oligocene pluton. The Oak Draw fault, a south-dipping normal fault south of Stockton Pass, juxtaposes the Miocene eruptive center against the underlying Oligocene pluton; the fault cuts north-northeast-trending 23 m.y. old quartz latite dikes.

Mylonitic gneiss derived from Precambrian metamorphic and igneous rocks occur as a partially preserved, gently dipping, sheathlike cover along the northeastern flank of the range. The rocks are weakly to intensely deformed and have a pronounced northeast-trending streaky lineation. The age and tectonic setting of the cataclasis is not clear at this time.

Introduction

The Pinaleno Mountains, better known locally as the Graham Mountains, constitute one of the highest and most rugged ranges in southeastern Arizona (Fig. 1). They lie mainly within the Silver City $1^{o}x2^{o}$ quadrangle, which is being mapped by the U.S. Geological Survey under the Conterminous United States Mineral Appraisal Program (CUSMAP). Large-scale geologic maps of the Pinaleno Mountains within the Silver City quadrangle will be published as the field work is completed. I am doing a study of the thermal history of the Precambrian rocks using fission-track ages of apatite and zircon, which have cooling temperatures of approximately 110^{o} to $140^{o}C$ and 175^{o} to $225^{o}C$, respectively. Zircon studies of several of the Precambrian units are being pursued by Leon T. Silver, using U-Pb isotope systems.

This is a preliminary report on the Precambrian metamorphic and plutonic terrain that underlies the eastern and highest parts of the Pinaleno Mountains and on the Tertiary volcanic and plutonic rocks that underlie the southern part of the range. This report deals with the description (Fig. 2) and distribution (Fig. 3) of rock units of Precambrian and mid-Tertiary age and the geologic history they record. Also, the relationships of several structures to the metamorphic core complex concept are discussed. Pliocene-Pleistocene valley-fill deposits and younger pediment surfaces and deposits that flank and locally occur in valleys of the range are not considered. Higgins' (1971) classification of cataclastic rocks is used in this report.

Previous geologic investigations within the study area consist of a reconnaissance map to the south by Cooper (1960) and a detailed map and structural synthesis of the Stockton Pass fault zone area by Swan (1975, 1976). Geologic maps are available for the northwestern end of the Pinaleno Mountains adjacent to the Silver City quadrangle (Blacet and Miller, 1978;

Bergquist, 1979), and farther west, in the Santa Teresa Mountains, the Klondyke quadrangle has been mapped by Simons (1964).

Rock Units

Rocks in the study area include two distinct terraines separated by a major unconformity. The older terrain comprises Precambrian rocks ranging in age from 1,700 to 1,100 m.y. old, including gneisses probably of sedimentary and volcanic origin, granitic plutons, metaquartz latite dikes, and diabase bodies. The younger terrain includes a mid-Tertiary granitic stock and a complex dome-flow eruptive center of felsic to intermediate composition that includes flows and flow breccias, volcaniclastics, air-fall and ash-flow tuffs, and dike swarms. Lithologic descriptions of the various rock units and their chronologic positions are given in Figure 2. Supplemental descriptions of the gneisses are presented below in reference to their possible protoliths. Names assigned to units are informal; formal names may be applied upon completion of this study.

Precambrian History

The Precambrian history recorded in the Pinaleno Mountains correlates with that outlined for southeastern Arizona by Cooper and Silver (1964), Shride (1967), and Silver (1978). Three major episodes include: (1) accumulation of the protoliths of the gneiss of the Pinaleno Mountains and probably of the metamorphic rocks of Bar X Canyon, prior to about 1,650 m.y. ago; (2) the regional metamorphism and deformation concomitant with emplacement of multiple granitic plutons about 1,400 to 1,650 m.y. ago; and (3) the emplacement of diabase and lamprophyre bodies tentatively correlated with similar activity in southern and central Arizona about 1,100 m.y. ago.

Protoliths of the gneiss of the Pinaleno Mountains and of the metamorphic rocks of Bar X Canyon are presently considered to have been primarily surfaceaccumulated layered rocks such as immature siliceous clastic rocks and (or) felsic volcanic flows and tuffs; Swan (1976) was of the same opinion for the gneiss of the Pinaleno Mountains. Both suites of metamorphic rocks are strongly layered and have pronounced compositional changes from unit to unit. Quartz-rich rocks, containing as much as 70 percent quartz, in the gneiss of the Pinaleno Mountains, and thinly layered, schistose, medium-grained, muscovite-bearing quartzites and a unit of quartzite metaconglomerate intercalated with quartzite and quartz-feldspar-muscovite gneiss in the metamorphic rocks of Bar X Canyon strongly suggest a sedimentary origin for at least some of the rocks. No relict features indicative of volcanic origin have been observed in any of these rocks.

The Bar X Canyon rocks are believed to be coeval with the gneiss of the Pinaleno Mountains. Xenoliths of metaquartzite that resemble metaquartzites of the Bar X Canyon suite occur in the granite of Stockton Pass east of the fault that juxtaposes the Bar X Canyon suite and the granite.

A widespread and major orogenic event about 1,400 to 1,650 m.y. ago accounted for the regional synkinematic metamorphism of the gneiss of the Pinaleno Mountains and the metamorphic rocks of Bar X Canyon. Plutonic activity ranged from synorogenic to postorogenic or anorogenic, based on structural and field relationships. Metamorphism of the exposed rocks appears to have occurred uniformly in the amphibolite facies. All rocks appear to have been completely recrystallized, as no relict minerals or textures have been recognized. Pegmatites are few, being more abundant toward the west, especially along the crest of the range near Webb Peak. Swan (1976) interpreted the initial metamorphism of the gneisses to have been of low rank and considered the subsequent plutonic activity to have been responsible for upgrading the gneisses to the amphibolite facies.

The gneiss of Johns Dam consists of augen-structured granitic material injected lit-par-lit fashion into the gneiss of the Pinaleno Mountains along the northeastern flank of the range. This injection is considered to be synorogenic to late orogenic emplacement, presently the earliest evidence of igneous activity. Foliations in the granitic rock and host gneiss rock are concordant, except for local crosscutting relationships indicating the plutonic nature of the granitic rock. A pronounced layering characterizes the Johns Dam unit.

The granite of Veach Ridge (the white gneissic granite of Swan, 1976) was dated by Swan (1976) as 1,363 ± 14 m.y. old (Rb-Sr whole-rock analysis). Its geometry, internal flow foliation, and xenoliths and schlieren oriented parallel to structural trends in the country rock suggest that the pluton was emplaced during late orogenic time.

The granite of Ladybug Saddle (the granite of Pinaleno of Swan, 1976), was dated by Swan (1976) as 1,384 ± 39 m.y. old (Rb-Sr whole-rock analysis). A biotite K-Ar date of 875 ± 30 m.y. (Shafiqullah and others, 1980) from this unit near the range crest probably indicates a younger thermal event that partially reset the biotite K-Ar clock. Zircon from this locality is presently being analyzed by Leon T. Silver to determine its U-Pb age, and I am determining apatite and zircon fission-track ages of the rock. Where foliation, schlieren, and xenoliths do occur in the granite of Ladybug Saddle, they parallel that of the country rock. I consider this pluton to be late orogenic but younger than the granite of Veach Ridge on the basis of its more discordant contacts and more massive internal fabric.

The granite of Treasure Park, an undated pluton, is a discordant, weakly foliated to massive body. It is weakly altered, unlike the other plutons, which are relatively fresh. This unit may be late orogenic to postorogenic.

The granite of Stockton Pass was dated as 1,405 ± 65 m.y. old (Rb-Sr whole-rock analysis) by Swan (1976). A complex pluton, the granite includes finer and coarser grained phases, pegmatites, and a quartz porphyry variant (found in Bar X Canyon, but not shown on the map). It is typically a massive rock. Its internal structure and complex nature suggest that it may be a postorogenic or anorogenic body.

The granite of Slick Rock, smallest of the granitic plutons, is undated. It is an elongate, discordant, almost massive body nearly devoid of xenoliths. It is probably also a postorogenic or anorogenic body. Although considered part of the Precambrian rocks in this report, it could be as young as Cenozoic.

Metaquartz latite dikes trend northwest and cut units as young as the granite of Veach Ridge. The dikes occur mainly in the northeast part of the range. They differ from the mid-Tertiary quartz latite dikes in being more variable in dip, commonly dipping less than 60 degrees, and in having a holocrystalline, micaceous, foliated groundmass.

Lamprophyre and diabase dikes and irregular bodies were probably intruded along the major faults of the Stockton Pass fault zone. Regional considerations indicate that the widespread diabase event in Arizona about 1,100 m.y. ago (Shride, 1967; Silver, 1978) may have been responsible for the emplacement of the mafic intrusions in Stockton Pass.

Tertiary History

The Tertiary record includes the emplacement of an Oligocene granite pluton and the formation of a Miocene eruptive center, both in the southern part of the Pinaleno Mountains. The granite of Gillespie Mountain (the Gillespie quartz monzonite of Swan, 1976) has been dated by Swan (1976) as 35.6 ± 0.7 m.y. old (biotite, K-Ar analysis). This pluton intrudes the gneiss of the Pinaleno Mountains and the granite of Stockton Pass and is in fault contact with the younger Tertiary volcanic rocks. The granite has been found only in fault contact with the volcanic units.

The conglomerate of Willow Spring underlies the Miocene volcanic rocks and consists of Precambrian detritus preserved in channels up to 0.5 km wide anda 200 to 300 m deep cut in the granite of Stockton Pass. These channel-fill deposits occur east of Greasewood Mountain on the northeast side of an inlier in the Miocene eruptive center but are not shown on the map. The conglomerates are assigned a questionable Miocene age.

The Miocene volcanic rocks, which include six map units, are considered remnants of a complex dome-flow eruptive center ranging from 23 to 27 m.y. old. Although no major vents have been observed, it is postulated that the principal eruptive center was buried beneath its own material in the general vicinity of Greasewood Mountain. Early products of the center were the basal volcanics of Frog Spring, a series of andesite flows and breccias and volcaniclastic beds, and the overlying andesite of Hog Spring, consisting of andesite porphyry flows. A zircon fission-track date of 26.8 ± 1.2 m.y. was obtained from a flow in the lower portion of the volcanics of Frog Spring; these rocks were referred to as Tertiary-Cretaceous by Cooper (1960).

Considerable erosion occurred and local relief developed between the eruption of the andesite of Hog Spring and the volcanics of Frog Spring. Hills underlain by Precambrian rocks, primarily the granite of Stockton Pass, were standing nearby during much of the accumulation of the Hog Spring rocks as indicated by numerous channel deposits of Precambrian granitic detritus interspersed throughout the section; no clasts of the Tertiary granite of Gillespie Mountain were observed. Several small andesite necks occur in the granite of Stockton Pass on the east side of the range just south of the Oak

Draw fault.

The diorite of Little Cottonwood Canyon, an irregular-shaped, sill-like body, was intruded at the base of the volcanic rocks between the underlying granite and overlying Frog Spring and Hog Spring units. Texturally and mineralogically the diorite resembles the andesite of Hog Spring and tentatively is considered to be coeval with the latest phase of Hog Spring eruptive activity.

The volcanics of Greasewood Mountain and of Gillman Canyon, comprising dominantly felsic rocks, are the later phases of the eruptive center. These volcanics contrast sharply with the andesitic rocks that record the initiation of the center. The Greasewood Mountain suite includes dacitic porphyry flows and flow breccias and occasional silicic tuffs. Zircon fission-track dates of 25.4 ± 1.3 m.y. and 26.1 ± 1.6 m.y. were obtained from two flows near the summit of Greasewood Mountain. The overlying volcanic rocks of Gillman Canyon, which are mainly air-fall tuffs and intercalated volcaniclastic beds and local rhyolite vitrophyres, mark a change in composition and eruptive style. Both rock suites dip west, south, and east from Greasewood Mountain, with the lighter colored volcanics of Gillman Canyon ringing the dacitic volcanics of Greasewood Mountain.

Quartz latite and latite dike systems invade the area from Greasewood Mountain to north of Gillespie Mountain. Dominant trends are east-northeast, northeast, northwest, and west-northwest. All rock units in the area are cut by the dikes. Two multiple quartz latite dike systems, southeast- and southwest-trending, nearly converge southeast of Greasewood Mountain and may reflect the presence of a pluton at a relatively shallow depth. Two dikes near the head of Oak Draw have yielded zircon fission-track dates of 22.8 ± 1.0 and 22.9 ± 1.4 m.y. These dikes are considered to be genetically related to the volcanic activity centered about Greasewood Mountain because of their close compositional, temporal, and spatial relationships.

Structure

Three structural elements in the Pinaleno Mountains are of considerable interest and have regional as well as local implications. One, the Stockton Pass fault zone, has been reported on before and is only briefly mentioned below. The other two, the Oak Draw fault and the mylonitic rocks along the northeast flank of the range, are discussed in more detail.

Swan (1975, 1976) described the west-northwest-trending Stockton Pass fault zone as being part of a major lineament and as having a long and involved history. He further indicated that initial movement of a left-slip nature appears to have begun during the Precambrian regional metamorphism of the gneiss of the Pinaleno Mountains. Diabase dikes were intruded along some of these faults in late Precambrian time. The youngest documented movement was right-slip faulting in the middle Tertiary, after emplacement of the Oligocene granite of Gillespie Mountain. Mylonitic fabric was generated during at least some of the Precambrian faulting, whereas crushing and the formation of gouge zones typified the Tertiary deformation. The extent of deformation, if any, that occurred along the fault zone between late Precambrian and middle Tertiary time is unknown.

The Oak Draw fault, a south-dipping normal fault, is the dominant structure in the southeastern part of the range. The fault separates the Tertiary volcanic sequence of the hanging wall from the Oligocene granite of Gillespie Mountain; the Precambrian granite of Stockton Pass occurs in both walls. Miocene quartz latite dikes intrude rocks of both walls. Along its eastern trace the fault dips 30 to 50 degrees south. In the center of the range, near where the strike changes from east-southeast to southwest, the fault steepens to 60 to 70 degrees. Southwest of this change in strike, the fault appears to be vertical. Rocks along the fault are crushed and locally form a gouge sheet; the zone of cataclasis is 0.5 to 10 m thick. Faults with similar crush and gouge zones within the Oligocene granite of Gillespie Mountain parallel the Oak Draw fault.

Movement on the Oak Draw fault was mainly dip slip but includes a right-slip component. The Precambrian granite of Stockton Pass probably enveloped the granite of Gillespie Mountain. Thus, at the west end of the Oak Draw fault, where the fault plane is steep, the Precambrian granite in the south wall has been dropped against the Oligocene granite. Consequently, because the dip of the fault flattens to the east, movement was also that of a normal fault; the hanging wall moved south. As the hanging wall moved south it also moved eastward as reflected by steeply dipping dikes that have right separation of 50 to 100 m across the fault. This separation may be due to a small downward rotation of the eastern end of the hanging wall block, or the entire block may have moved relatively eastward, indicating a southeast dip of the western segment of the fault. The magnitude of dip slip is inferred from stratigraphic evidence in the hanging wall rocks. Clasts in the basal Tertiary unit, the conglomerate of Willow Spring, and in the andesite of Hog Spring include only Precambrian detritus, primarily the granite of Stockton Pass; no clasts of the granite of Gillespie Mountain were observed. Movement was therefore not more than about 1 to 2 km, for the Stockton Pass pluton does not occur north of the Stockton Pass fault zone. The absence of Gillespie Mountain clasts could be explained by having that rock covered either by its enveloping Stockton Pass host rock or by the Tertiary rocks themselves, before they moved off part of the stock. Finally, the absence of potential source vents for the volcanic rocks in the Gillespie Mountain pluton favors limited movement of the volcanics. Movement on the Oak Draw fault, and on parallel faults in the Gillespie Mountain pluton, occurred less than 23 m.y. ago, the age of the quartz latite dikes, but how much less is not known.

Protomylonite and mylonite gneisses derived primarily from the gneiss of the Pinaleno Mountains and the gneiss of Johns Dam occur as a partially preserved, gently dipping, sheath-like covering along the northeastern flank of the range. Shear foliation in the mylonitic rocks generally dips gently north to northeast off the flank of the range, cutting across the foliation of the protoliths. Mylonitization generally dies out rapidly into the range. The typical rock is a lineated medium-grained tectonite containing porphyroclasts of felsic minerals or mineral aggregates set in a finer grained schistose assemblage of quartz, feldspar, and mica. A northeast-trending streaky lineation generally is developed on shear surfaces. All degrees of mylonitization are present, ranging from incipient deformation to ultramylonitization. Ultramylonitized rocks are rare, but where present occur as thin sheets interleaved with protomylonitic rocks. The only readily detected indication of retrogressive metamorphism associated with the cataclasis is locally weakly chloritized biotite.

The presence of mylonitic rocks along the foot of the range was first shown by Swan (1976). In a subsequent study, Davis (1980) referred to the core of the range as being dominantly fine- to medium-grained quartz monzonitic mylonitic augen gneiss, yet his diagrammatic map resembles that of Swan, showing mylonitic rocks occurring only along the foot of the range. As outlined in the description of rock units (Fig. 2) and as shown on the geologic map (Fig. 3), the dominant rock is the gneiss of the Pinaleno Mountains, a nonmylonitic rock unit made up of a succession of layered gneisses, varying in bulk composition, that are medium grained and foliated but lack augen structure. One of the problems in understanding the mylonites is in identifying the protoliths. As a specific example, the porphyritic granite phase of the gneiss of Johns Dam along State Highway 366 unquestionably has been transformed into an augen mylonite gneiss having streaky lineations on shear surfaces. However, the amount of differential shear, penetrative deformation, recrystallization, and comminution needed to form this augen mylonite gneiss was much less than it would appear because the protolith was an augen orthogneiss. Davis (1980) gave the impression that the mylonitic rocks owe their augen or flaser structure wholly to the cataclasis. He called a rock along the northwest end of the range "a thin selvage of mylonitic breccia" along a "dècollement" that juxtaposes Tertiary volcanic and sedimentary rocks against Precambrian igneous and possibly metamorphic rocks (Blacet and Miller, 1978; Bergquist, 1979). This fault cuts northeast-trending rhyolitic dikes 24.65 ± 0.6 m.y. old (Shafiqullah and others, 1980) in the footwall. I have observed the fault contact at many localities and believe that the "mylonitic breccia" is iron-stained shear surfaces in microbreccia, breccia, and fault gouge. This is an important distinction to make because the implication of mylonitization is crucial to the tectonic interpretation of the range.

The age of mylonitization is unknown. Protoliths in every instance in the Pinaleno Mountains are Precambrian rocks. A K-Ar whole-rock date of 28.3 ± 0.7 m.y. was obtained from a quartzofeldspathic ultramylonite lens in Ash Canyon (Shafiqullah and others, 1980). This date is not necessarily the mylonitization, for the significance of such a date on a mylonitic rock is subject to various interpretations. A wide range of thermal events subsequent to mylonitization could produce such an apparent age.

Regional Structural Considerations

The Pinaleno Mountains is the easternmost of the metamorphic-core-complex ranges in the southwest (Davis, 1980; Davis and Coney, 1979; Rehrig and Reynolds, 1980). Only part of the metamorphic-core-complex story, as seen in many of the ranges to the west, is preserved in the Pinaleno Mountains. The range has the typical metamorphic and igneous core with a partially preserved sheath of gently dipping, lineated, mylonitic rocks; in other ranges the sheaths cover much of the core rocks. The plutonic events of the Pinaleno Mountains, which range from about 1,650 to 23 m.y. old, appear to be less complex than those in ranges such as the Rincon-Catalina complex (Drewes, 1977; Drewes and Thorman, unpub. data, 1980; Creasey, Banks, Ashley, and Theodore, 1977), so perhaps a clearer understanding may be obtained of the metamorphic rocks from the Pinaleno Mountains core complex.

Tertiary faults dipping gently off the core rocks are another structural element that the Pinaleno Mountains have in common with other metamorphic core complexes. Such faults and overlying allochthonous rocks occur not only on the south side of the range near Greasewood Mountain within the study area, but also on the southwest side of the saddle between the Pinaleno and Santa Teresa Mountains (Blacet and Miller, 1978; Bergquist, 1979). Several features of these allochthonous Tertiary blocks, such as their young age (as young as 23 m.y.), weak internal deformation, and the presence of only moderately brecciated fault zones separating them from core rocks, are similar to those in other ranges.

Metamorphic-core-complex elements in other ranges not seen in the Pinaleno Mountains include a moderately to strongly deformed and metamorphosed succession of sedimentary and (or) volcanic rocks of late Precambrian through Mesozoic age that occur as tectonic slivers above the mylonitic sheath, and weakly to strongly deformed unmetamorphosed sedimentary and volcanic rocks of late Precambrian through possibly earliest Tertiary age that occur as tectonic slivers above the metamorphosed allochthonous rocks. These two elements plus the undeformed allochthonous Tertiary rocks make up the three-fold allochthonous carapace that is typical of many metamorphic core complexes elsewhere. Differences in style of deformation within and between the three allochthonous elements indicate shallow deformation of mid-Tertiary age and much more complex and intense deep-seated deformation of latest Mesozoic to earliest Tertiary age. By analogy the mylonitic gneiss was formed during the earlier period of intense deep-seated deformation, and the low-angle normal faults formed during the later period of shallow-seated deformation.

Conditions compatible with the earlier period of intense deformation in the Pinaleno Mountains may have been found at depths of at least 3 to 4 km. The mylonitic rocks appear from preliminary study to be of the amphibolite facies, or possibly the highest greenschist facies. Only a few rocks show a weak retrograde metamorphism in the form of chloritization of biotite, suggesting that deformation occurred under P-T conditions where biotite was partially unstable. Such conditions suggest depths of 3 to 4 km or more. The available cover of pre-Tertiary rocks is probably 3 to 4 km thick, an adequate depth for the metamorphic grade. For example, approximately 3.5 km of Paleozoic and Mesozoic sedimentary and volcanic rocks occur in the eastern Santa Teresa Mountains (Simons, 1964). A similar or greater thickness of Paleozoic and Mesozoic rocks is present in the northern Chiricahua and Dos Cabezas Mountains (Drewes, oral commun., 1980). The projection of a cover of Paleozoic and Mesozoic rocks this thick across the Pinaleno Mountains is therefore reasonable. Their absence from the Pinaleno Mountains probably reflects erosion during early Tertiary time.

Conditions compatible with the later period of shallow deformation were found at a depth of a kilometer or so. This depth is equal to the thickness of the Tertiary rocks at Greasewood Mountain and at the western end of the range. As pointed out above, this deformation is younger than the 23- to 25-m.y.-old dikes that are cut by the faults. The Tertiary and Precambrian rocks that constitute the fault block above the Oak Draw fault appear to have moved only a short distance. In contrast, the Tertiary rocks in the fault block at the western end of the range more closely resemble rocks in the Galiuro Mountains, 15 to 20 km to the southwest, which may be the direction in which their source lies.

The triggering mechanism that caused the shallow-seated normal faulting is considered to be rapid differential uplift. Gravity tectonics resulted, and large masses moved along low-angle normal faults towards structurally lower areas. Fault activity along the Stockton Pass fault zone and extending south to the Oak Draw fault may have been more or less contemporaneous with the post-23-m.y.-old low-angle gravity tectonics. In fact, it may have been part of the triggering action that caused the gravity tectonics. Crushed rock and fault gouge mark some faults in the Stockton Pass fault zone and look identical to those along the Oak Draw fault, suggesting a comparable shallow depth for the faulting in both instances. In fact, several of the Stockton Pass fault-zone faults that offset the intrusive contact of the mid-Tertiary granite of Gillespie Mountain have right separation, exactly as occurs along the Oak Draw fault and related structures in the Tertiary granite. Thus strike-slip and dip-slip faulting may have been active concurrently.

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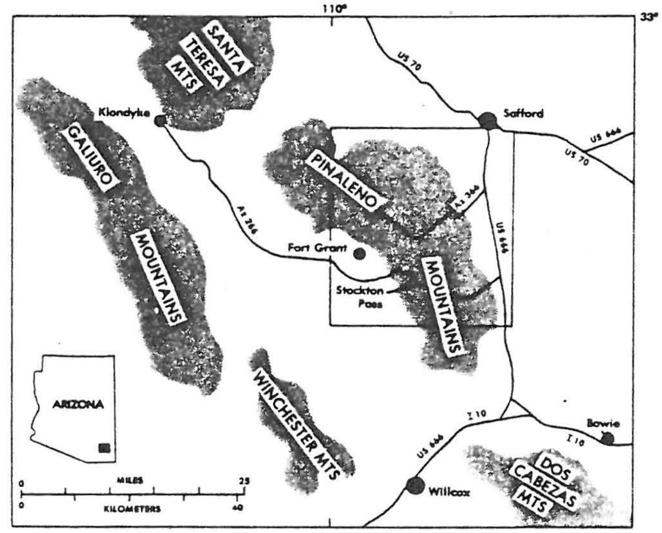

Figure 1. Index map showing location of study area.

AGE		ROCK UNIT	LITHOLOGIC DESCRIPTION	MAP SYMBOL
CENOZOIC	MIOCENE	DIKE ROCKS	Single, multiple, and composite dike systems including quartz latite, latite, and diabase. Quartz latites have a very fine grained quartz-feldspar-mica matrix with fine- to medium-grained quartz and feldspar phenocrysts. Two dikes have zircon fission-track ages of 22.8 ± 1.4 and 22.9 ± 1.0 m.y. Latite dikes, in part porphyritic, are predominantly plagioclase-biotite-hornblende rocks. Quartz latite dikes are dominant; they form the thick multiple dike systems east and north of Greasewood Mountain. Diabase dikes occur in the Stockton Pass area, but are not shown on the map.	Tql
CENOZOIC	MIOCENE	VOLCANICS OF GILLMAN CANYON	Interbedded rhyolite flows and air-fall tuffs and volcaniclastic beds. Flow rocks are light-colored quartz-sanidine-biotite vitrophyres as much as 8-10 m thick. Air-fall tuffs are light colored, well bedded, and contain quartz, sanidine, and biotite crystals, pumice, and lithic fragments from underlying units; they are irregular in thickness and distribution. Volcaniclastic beds occur as channel deposits or thin layers between flows and tuffs.	Tgc
CENOZOIC	MIOCENE	VOLCANICS OF GREASEWOOD MOUNTAIN	Dacitic porphyry flows and flow breccias as much as 10 to 15 m thick. Rocks are plagioclase-biotite-hornblende porphyries with a pilotaxitic matrix of plagioclase, biotite, hornblende, and opaque minerals. A few rocks contain minor amounts of quartz, clinopyroxene, and hypersthene. Zircon fission-track ages of 25.4 ± 1.3 m.y. and 26.1 ± 1.6 m.y. were obtained from two flow units.	Tgm
CENOZOIC	MIOCENE	DIORITE OF LITTLE COTTONWOOD CANYON	Clinopyroxene-biotite diorite to quartz diorite. Rock is medium grained, hypidiomorphic to panidiomorphic granular, and has myrmekitic and graphic interstitial intergrowths.	Tdl
CENOZOIC	MIOCENE	ANDESITE OF HOG SPRING	Plagioclase-pyroxene andesite porphyry flows as much as 8-10 m thick. Rocks contain medium- to coarse-grained plagioclase (labradorite) and clinopyroxene in a pilotaxitic to felty matrix. Rocks are weakly to moderately chloritized and epidotized. Rocks resemble turkey-track andesite of Cooper (1961). Local thin channel deposits of granitic clasts derived from underlying Precambrian terrain.	Tah
CENOZOIC	MIOCENE	VOLCANICS OF FROG SPRING	Interbedded andesite flows and breccias and volcaniclastic beds. Flow rocks have felty to pilotaxitic texture with fine- to medium-grained phenocrysts of saussuritized plagioclase and altered pyroxene and hornblende. Volcaniclastic beds are intraformational sandstones, conglomerates, and breccias. Zircon fission-track age of 26.8 ± 1.2 m.y. from flow rock.	Tfs
CENOZOIC	MIOCENE?	CONGLOMERATE OF WILLOW SPRING	Cobble to boulder conglomerate derived from underlying Precambrian terrain. Conglomerate occurs as local channel deposits between volcanics of Frog Spring and granite of Stockton Pass in Willow Spring Canyon.	not shown on map
CENOZOIC	OLIGOCENE	GRANITE OF GILLESPIE MOUNTAIN	Porphyritic biotite granite. Medium- to coarse-grained feldspar phenocrysts in a medium-grained rock containing quartz, plagioclase (An_{5-15}), microcline, biotite, sphene, muscovite, epidote, apatite, zircon, and opaque minerals. Rock ranges from quartz syenite to quartz monzonite to granite, being mainly in granite field (IUGS classification of igneous rocks). Schachbrett (checkerboard) intergrowth of plagioclase and microcline present in most rocks. K-Ar biotite age of 35.6 ± 0.7 m.y. (Swan, 1976).	Tgg

Figure 2. Lithologic descriptions of rock units in the Pinaleno Mountains. Symbols used for geologic map (Fig. 3) shown in right-hand column.

AGE		ROCK UNIT	LITHOLOGIC DESCRIPTION	MAP SYMBOL
PROTEROZOIC	Y	METAQUARTZ LATITE DIKES	Schistose quartz latite dikes. Rocks have fine- to medium-grained quartz, plagioclase, and alkali feldspar phenocrysts in very fine to fine-grained schistose to granoblastic matrix of quartz, plagioclase, alkali feldspar, muscovite, biotite, and epidote.	
		DIABASE	Diabase dikes and irregular-shaped bodies. Rocks comprise fine- to coarse-grained plagioclase (labradorite), clinopyroxene, and opaque minerals; clinopyroxene altered to chlorite, biotite, and hornblende.	Yd
		LAMPROPHYRE	Dikes composed mainly of fine- to medium-grained, massive to weakly foliated biotite and chloritized biotite. Bodies are small and have close affinity spatially to diabase bodies in Stockton Pass.	not shown of map
		GRANITE OF SLICK ROCK	Two-mica granite. Medium-grained rock with hypidiomorphic-granular texture, typically containing quartz (31%), microcline (27%), plagioclase An_{30} (35%), biotite (2%), muscovite (2%), epidote (2-3%), sphene (<0.5%), and trace amounts of zircon, apatite, and opaque minerals. Mafic minerals occur chiefly in clots.	Ysr
		GRANITE OF LADYBUG SADDLE	Porphyritic biotite granite. Massive to weakly foliated, medium- to coarse-grained rock with hypidiomorphic-granular texture; microcline phenocrysts. Typical rock includes quartz (27-35%), microcline (24-26%), plagioclase An_{30-34} (35%), biotite (4-12%), sphene (1-1.5%), epidote (1-2%), and trace amounts of zircon, apatite, allanite, and opaque minerals. Mafic minerals occur primarily in clots. Xenoliths and schlieren present mainly along margin, where they are oriented parallel to trends in country rock.	Yl
		GRANITE OF STOCKTON PASS	Porphyritic two-mica granite. Medium- to coarse-grained, hypidiomorphic-granular rock, with microcline phenocrysts, having distinct finer and coarser phases in a complex intrusive relationship. Typical rock includes quartz, microcline, plagioclase, biotite, muscovite, and sphene and trace amounts of zircon, apatite, and opaque minerals. Mafic minerals are both disseminated and in clots. Graphic and myrmekitic intergrowths are common; pegmatitic phases common near contact with granite of Gillespie Mountain and have megascopic graphic intergrowths. Mafic minerals commonly absent on weathered surface giving leucocratic appearance.	Ysp
		GRANITE OF TREASURE PARK	Porphyritic two-mica granite. Massive to weakly foliated, medium-grained porphyritic rock with hypidiomorphic-granular texture. Typical mineral assemblage includes quartz, microcline, saussuritized plagioclase, chloritized biotite, muscovite, and epidote with trace amounts of zircon, sphene, apatite, and opaque minerals. Rock is typically moderately to deeply weathered in contrast to fresh country rock.	Yt
		GRANITE OF VEACH RIDGE	Biotite granite. Massive to weakly foliated, medium- to coarse-grained rock with hypidiomorphic-granular texture. Typical rock includes quartz, microcline, plagioclase, biotite (1-15%), and epidote with trace amounts of sphene, allanite, zircon, apatite, and opaque minerals. Xenoliths, 0.5 to 100 m across by 1 to 300 m long, of gneiss of Pinaleno Mountains are scattered throughout pluton and are aligned parallel to regional trend.	Yvr
	X&Y	GNEISS OF JOHNS DAM	Intercalated granitic gneiss and gneiss of Pinaleno Mountains. Granitic gneiss is medium-grained biotite-rich rock with coarse- to very coarse grained microcline phenocrysts, having augen texture where rock is foliated. Granitic gneiss intrudes finer grained gneiss of Pinaleno Mountains; individual units are as much as 50 m thick, giving pronounced banded appearance on aerial photographs. Typical granitic gneiss mineral assemblage includes quartz, microcline, oligoclase-andesine plagioclase, biotite, epidote, and sphene.	YXjd
	X	GNEISS OF PINALENO MOUNTAIN	Granitic to granodioritic gneisses. Fine- to medium-grained, moderately to strongly foliated rocks. Foliation includes aligned disseminated mafic minerals, compositional banding or layering, and elongation of normally equant or stubby minerals. Rock types include quartz-two feldspar-biotite, quartz-plagioclase-biotite, quartz-plagioclase-microcline, quartz-two feldspar-two mica, and muscovite-bearing quartzite rocks. Amphibolites are locally present. Probably metamorphosed sandstone or felsic volcanic rocks.	Xpm
		METAMORPHIC ROCKS OF BAR X CANYON	Intercalated gneiss and quartzite. Fine- to medium-grained, moderately foliated rocks. Rock units include muscovite-bearing quartzite, quartz-two mica-plagioclase gneiss, and a few beds of quartzitic metaconglomerate. Rocks locally strongly weathered to purplish red and brown hues. Probably metamorphosed sandstones and (or) felsic volcanic rocks.	Xm

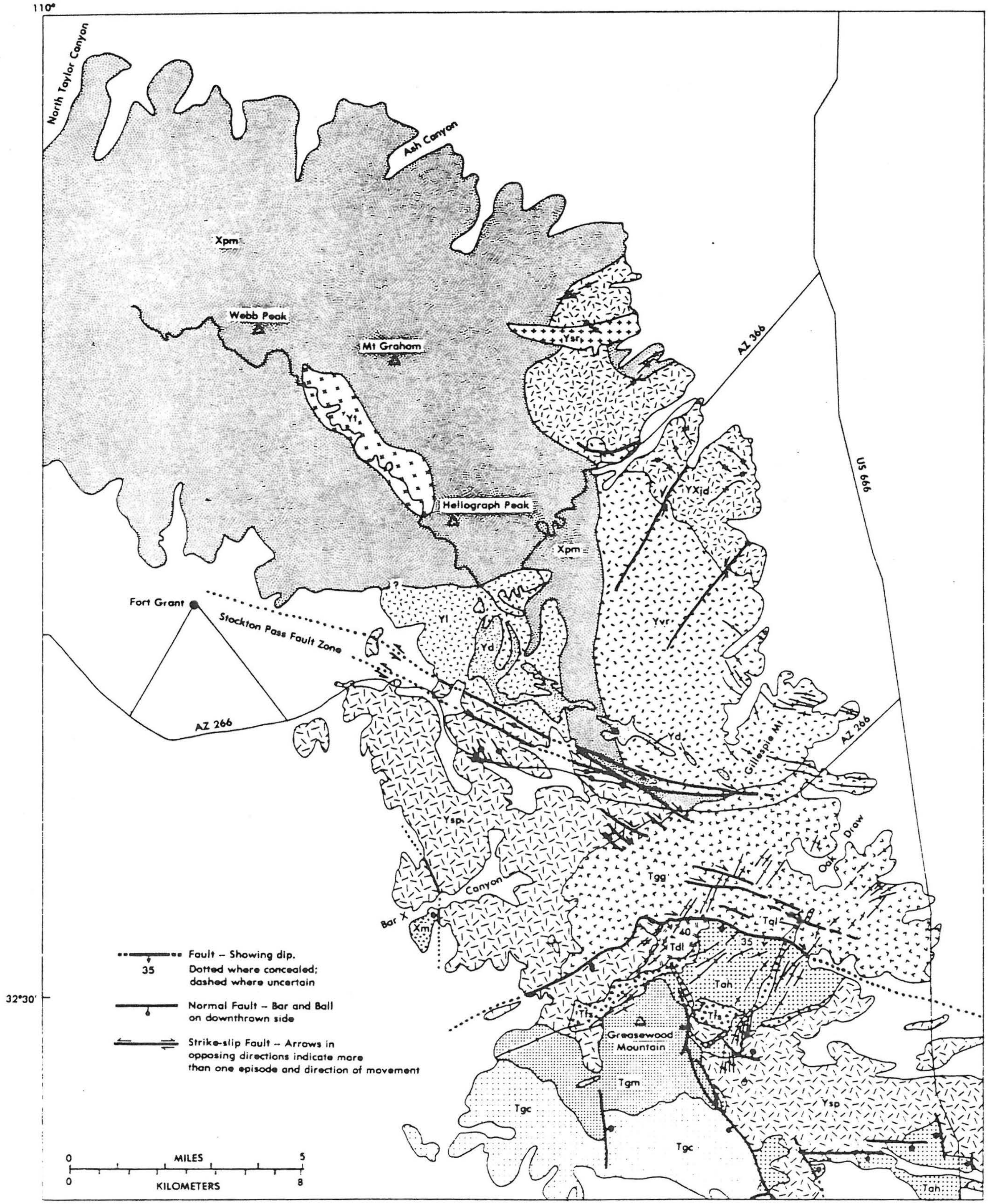

Figure 3. Geologic map of the Pinaleno Mountains. Symbols for rock units explained on Figure 2.

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THE EAGLE PASS DETACHMENT, SOUTHEASTERN ARIZONA: PRODUCT OF MID-MIOCENE LISTRIC (?) NORMAL FAULTING IN THE SOUTHERN BASIN AND RANGE

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ABSTRACT

Mid-Miocene low-angle normal faulting is an integral part of the regional tectonic strain of the southern Basin and Range. The distinguishing characteristic is the presence of "detachments" -- allochthons of rotated hanging-wall rocks resting in very low-angle fault contact on structurally deeper, generally older, footwall rocks. Rotation of detachment strata was accomplished by imbricate listric (?) normal faulting and coordinated tear faulting. Regional structural domains, each thousands of kilometers in area, contain detachments marked by uniform sense of detachment-strata rotation.

The Eagle Pass detachment, 70 km northeast of Tucson, Arizona, reveals informative structural details of parts of the system. Steeply SW-dipping mid-Miocene volcanic and sedimentary rocks, several thousands of meters in thickness, rest in very low-angle ($10^{\circ}\pm$) fault contact on Precambrian granitic basement. Detachment strata were faulted into position by NE-directed translation accompanied by rotation of strata to steep SW dips. The detachment fault which underlies the 25 km^2 detachment has the form of a large NE-trending mullion-trough with relief of at least 500 m. Along the keel of the detachment, Precambrian quartz monzonite was transformed into a hydrothermally altered selvage of microbreccia. The steep ($80^{\circ}\pm$) northwest margin of the mullion-trough served as a tear-fault along which quartz monzonite was crushed, volcanics were dragged, and fanglomerate was locally sheared into flattened-pebble conglomerate.

Detachments like that at Eagle Pass resemble slump blocks on a grand scale. They are elements of a regional system of extensional deformation which coincides with the belt of Cordilleran metamorphic core complexes. Mid-Miocene detachment faulting post-dates the formation of mylonitic tectonite of core complex affinity, but the direction of translation of the detachments is parallel to penetrative lineation in nearby and/or underlying tectonite. Moreover, in southeastern Arizona examples, the sense of translation of detachments matches the sense of simple-shear movements disclosed by tectonite fabrics. We speculate that large crustal displacements achieved during extension-induced formation of tectonite produced a geometric and/or dynamic condition which strongly influenced subsequent detachment faulting. Consequently, with renewed (?) or continued (?) extension in mid-Miocene time, deep upper-crustal slumping was triggered and detachment rocks were translated along listric (?) normal faults in the direction and sense of the earlier displacments which had accommodated the formation of tectonite.

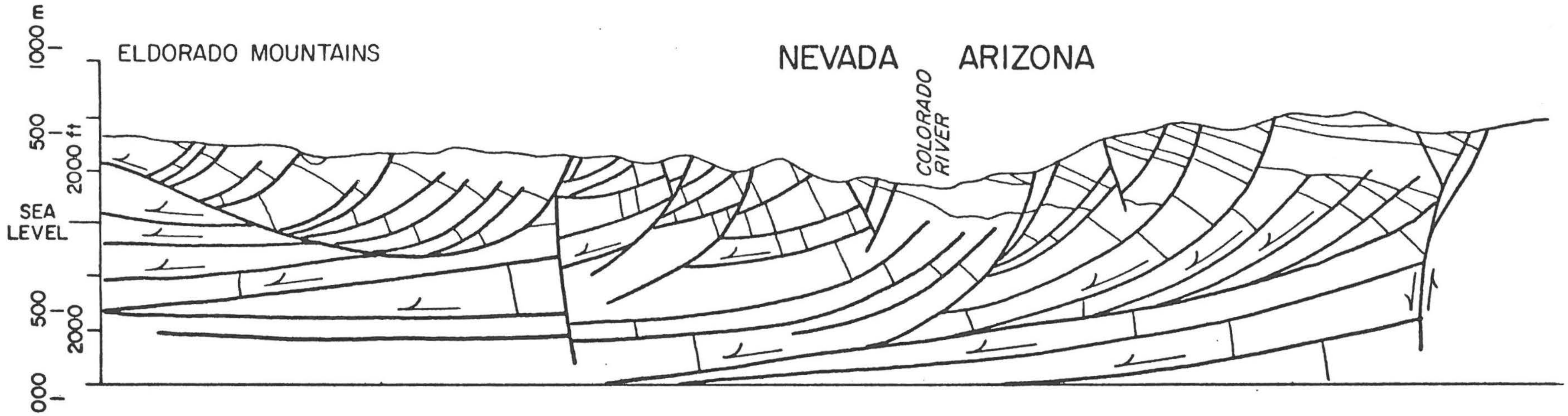
INTRODUCTION

Perceptions of the structural evolution of the Basin and Range tectonic province are changing in accord with new facts derived from careful studies of the mid- to late-Tertiary record. Armstrong's (1972) success in discriminating between Cenozoic and pre-Cenozoic low-angle faults in the Basin and Range of western Utah and eastern Nevada provided a glimpse of extensive Tertiary 'denudational' faulting which had largely gone unnoticed. Anderson's (1971, 1978) mapping and synthesis of structures in the region south of Lake Mead in Arizona and Nevada yielded geological details regarding one class of denudational faults, namely a listric normal faulting which he regarded as having accommodated thin-skin distension of the crust in an east-west direction. The fundamental relationship which he observed repeatedly is faulted and rotated Miocene strata separated from underlying Precambrian granitic basement by listric normal-slip faults (Fig. 1). Anderson demonstrated that the listric normal faulting pre-dated the high-angle faulting which blocked out the present basins and ranges.

The fault system which Anderson (1971) described is an integral and representative component of regional tectonic strain in the southern Basin and Range. G.A. Davis, E.G. Frost, E.J. Shackelford, and J.L. Anderson have identified "detachment faulting" in southeastern California and westernmost Arizona, and S.C. Creasey, M.H. Krieger, H.W. Peirce, R.B. Scarborough, W.A. Rehrig, S.R. Reynolds, P.E. Damon, and M. Shafiqullah have mapped the "shallow, rotational faults" in west central and southeastern Arizona (references follow). The mid-Miocene timing for the faulting can be proved or convincingly argued in the structural systems described by these workers. Eberly and Stanley (1978) helped to pioneer the awareness that mid-Miocene faulting is a distinctive tectonic episode which preceded late Miocene block faulting. Papers by Scarborough and Peirce (1978) and Shafiqullah and others (1978) likewise have emphasized that this major extensional faulting affected the southern Basin and Range region during the Miocene, before the classical Basin and Range disturbance. Wright and Troxel (1973) recognized and grasped the significance of this "shallow" faulting early on, and their descriptions and insights remain fundamental.

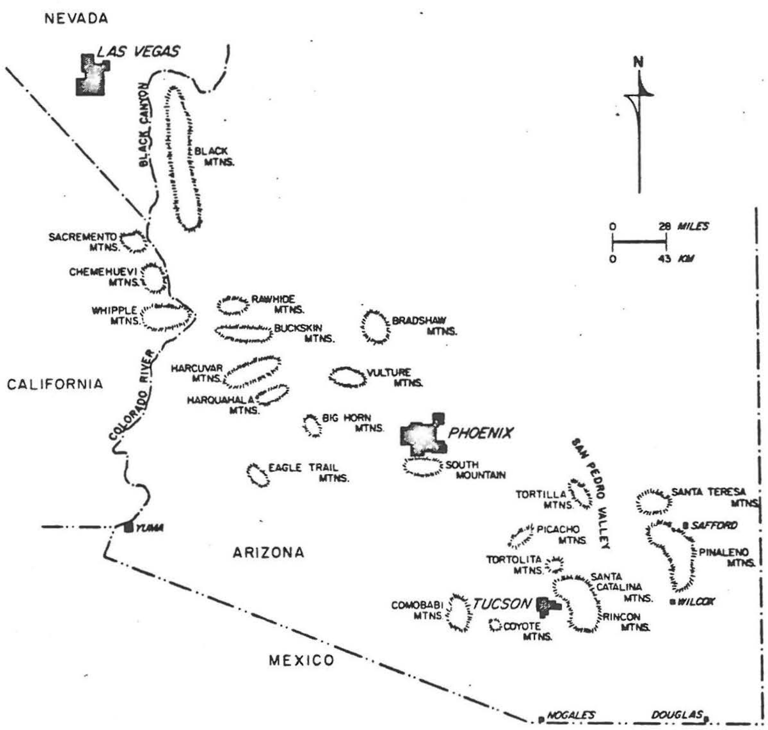
BACKGROUND

The geologic characteristics of the faulted terranes in southeastern California and in Arizona are summarized in Table 1. Geographical reference is provided in Figure 2. The distinguishing characteristic of these terranes is the presence of "detachments" -- i.e., allochthons of rotated handing-wall rocks resting in very low-angle fault contact on structurally deeper, generally older, footwall rocks. The term "detachment" is extracted from the expression "detachment fault", introduced by G.A. Davis and others (1979, 1980) to describe low-angle faults which mark the base of mid-Miocene allochthons in the Whipple-Buckskin-Rawhide Mountains terrane in southeastern California



ELDORADO MOUNTAINS
NEVADA
ARIZONA
COLORADO RIVER
1000 m
500
2000 ft
SEA LEVEL
500
2000
1000

REGION	WORKERS	SIZE OF TERRANE	CHIEF FOOTWALL ROCKS	CHIEF DETACHMENT ROCKS	DIRECTION OF DIP OF DETACHMENT STRATA	SENSE OF TRANSPORT	TREND OF LINEATION IN CLOSEST MYLONITIC TECTONITE
NW Arizona, SE Nevada: Black Mtns.	Anderson (1971, 1977, 1978).	1000$\pm$Km^2	Precambrian granite and granitic gneiss.	Oligocene (?)-Miocene volcanic and sedimentary. rocks	East-northeast at moderate to steep angles.	S70°W$\pm$10°	N60°E$\pm$10°
SE California: Sacremento, Chemehuevi, and Whipple Mtns.	Davis and others (1977, 1979, 1980), Anderson and others (1979), Frost (1979).	3000$\pm$Km^2	Mylonitic gneiss, Precambrian quartzo-feldspathic gneiss, Mesozoic crystalline rock.	Oligocene-Miocene volcanic and sedimentary rocks, Mesozoic granitic rocks, Precambrian quartzo-feldspathic banded gneiss.	S50°W$\pm$10° at moderate to steep angles.	N50°E$\pm$10°	N50°E$\pm$10°
WC Arizona: Harcuvar, Harquahala, Vulture, Big Horn, Eagle Tail, and South Mtn(s).	Rehrig and Reynolds (1977, 1980), Reynolds and Rehrig (1980), Shackelford, (1977, 1980), Rehrig and others (1980), Davis and others (1980).	3000$\pm$Km^2	Precambrian quartzo-feldspathic gneisses, Precambrian granite, metamorphosed Pale-ozoic rocks, Mesozoic crystalline rocks, and Tertiary granitic rocks.	Oligocene-Miocene sedi-mentary and volocanic rocks, Paleozoic sedi-mentary rocks, Precam-brian basement rocks.	S55°W$\pm$10° at moderate to steep angles, except between Big Horn and Vulture Mtns. where strata dip N60°E$\pm$10°.	N55°E$\pm$10°	N55°E$\pm$10°
SE Arizona: Picacho, Catalina-Rincon, and Pinaleño Mtns. and San Pedro Valley.	Creasey (1965), Pashley (1966), Krieger (1974a, b,c), Cornwall and Krieger (1975), Drewes (1975, 1977), Blacet and Miller (1978), Scarborough and Pierce (1978), Davis (1980).	4000$\pm$Km^2	Precambrian granite and schist, mylonitic tec-tonite, Paleozoic sedi-mentary rocks.	Oligocene-Miocene volcanic and sedimentary rocks, Mesozoic and Paleozoic sedimentary rocks, Precambrian crystal-line rocks.	S60°W$\pm$10° at moderate to steep angles, except near Pinaleno Mtns. where strata dip S50°W$\pm$10°.	S60°W$\pm$10° except near Pinalenos (N50°E$\pm$10°)	N60°E$\pm$10° except near Pinalenos (N40°E$\pm$10°)


NEVADA
LAS VEGAS
BLACK CANYON
BLACK MTNS.
N
0 28 MILES
0 43 KM
SACREMENTO MTNS.
CHEMEHUEVI MTNS.
WHIPPLE MTNS.
RAWHIDE MTNS.
BUCKSKIN MTNS.
BRADSHAW MTNS.
CALIFORNIA
COLORADO RIVER
HARCUVAR MTNS.
HARQUAHALA MTNS.
VULTURE MTNS.
BIG HORN MTNS.
PHOENIX
SAN PEDRO VALLEY
EAGLE TRAIL MTNS.
SOUTH MOUNTAIN
TORTILLA MTNS.
SANTA TERESA MTNS.
SAFFORD
PICACHO MTNS.
PINALENO MTNS.
YUMA
ARIZONA
TORTOLITA MTNS.
SANTA CATALINA MTNS.
WILCOX
COMOBABI MTNS.
TUCSON
COYOTE MTNS.
RINCON MTNS.
MEXICO
NOGALES
DOUGLAS

and western Arizona. The use of the term "detachment" follows the descriptive emphasis of G.A. Davis and others (1979, 1980) and underscores the allochthonous nature of these rocks without imposing constraints on size, shape, or dynamics of origin.

Rocks in the detachments range in age from Precambrian to middle Miocene. The allochthonous rocks are always composed of rocks from the local geologic setting. Some structures within the detachments evolved prior to the faulting, but much of the conspicuously brittle deformation appears to have attended the faulting itself. Rotation of detachment strata was accomplished by some form of normal faulting and coordinated tear faulting. R.E. Anderson (1979, pers. comm.) has observed that 'marginal shear zones' appear to have coordinated the internal movements of structural lobes within larger masses. The strike of bedding of hanging-wall strata is typically oriented at right angles to the direction of greatest extension in the detachments.

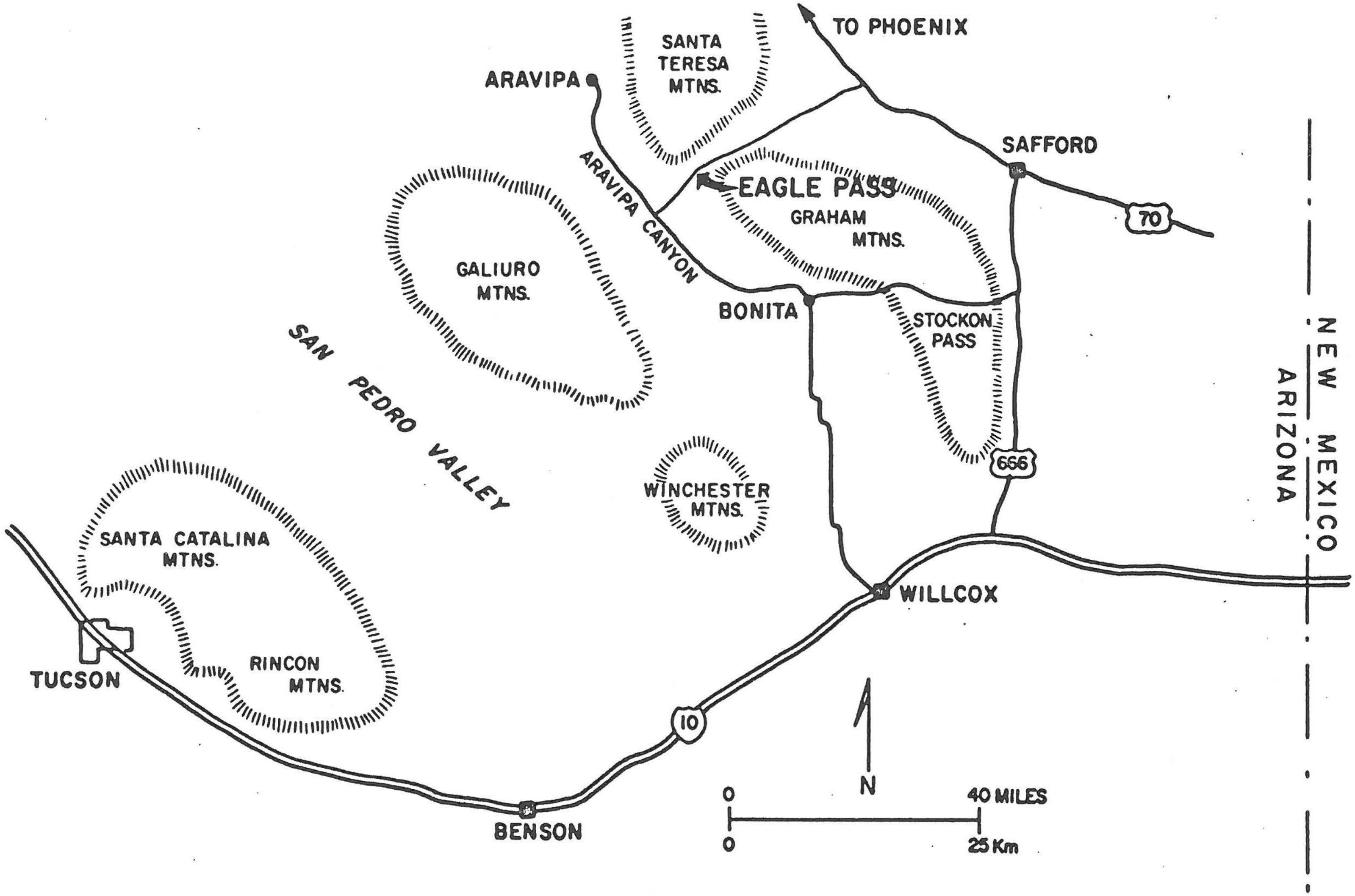
Regionally, there are broad structural domains of homoclinal detachment strata (Table 1), each of which is characterized by (1) uniform dip-direction of hanging-wall strata and thus (2) uniform sense of bedding rotation. Rehrig and Heidrick (1976) first called attention to the regularity of dip direction of Oligocene/Miocene volcanic and sedimentary rocks in the Basin and Range of Arizona. They concluded that the regional pattern was produced by "thin-skinned, rotational faulting and tilting" (Rehrig and Heidrick, 1976, p. 217). Stewart (1980) has observed that homoclinal-dip domains are characteristic of the entire Basin and Range province.

The footwall rocks on which the detachments rest are typically composed of Precambrian basement, although any relatively old, relatively deep component of the local geological column might compose the footwall. The footwall rocks generally appear to be autochthonous, but it is probable that the footwall terranes may themselves be underlain and rotated by yet deeper faults, perhaps in the manner shown in Figure 1. Two types of footwall terranes may be distinguished: normal country rock, and country rock which has been converted to mylonitic tectonites through shear-induced flow. The latter include portions of metamorphic core complexes (see Crittenden and others, 1980). In footwall terranes of normal country rock, only the uppermost several meters of the footwall rocks show conspicious signs of fault-induced deformation: intense fracturing, microbrecciation, and chloritic and ferruginous hydrothermal alteration. In metamorphic core complexes, footwall rocks directly beneath the detachments may be extensively microbrecciated, fractured, and altered, locally to depths of 100 m or more. Furthermore the microbreccia capping is typically underlain by hundreds of meters of lineated mylonitic tectonite. The fracturing and microbrecciation is physically superimposed on the tectonite fabric. Final emplacement of detachments clearly post-dated the development of lineated tectonite in such settings.

THE EAGLE PASS DETACHMENT, SOUTHEASTERN ARIZONA

INTRODUCTION

The Eagle Pass detachment northeast of Tucson near Safford, Arizona, (Fig. 3) displays a structural style which both conforms and adds to our understanding of the mid-Miocene denudational systems. The Eagle Pass structure is the easternmost known detachment in southeastern Arizona. This structure and its surrounding setting was previously mapped by P.M. Blacet and S.T. Miller (1978) and Berquist (1979). J.J. Hardy, Jr. mapped the structure again during the course of independent undergraduate research at The University of Arizona, emphasizing structural details exposed along the sole of the detachment. The geological location of the Eagle Pass detachment is especially instructive for its footwall is non-tectonite Precambrian quartz monzonite, even though lineated tectonite derived from this same rock crops out only several kilometers to the east (Swan, 1976; Davis, 1977, 1980).

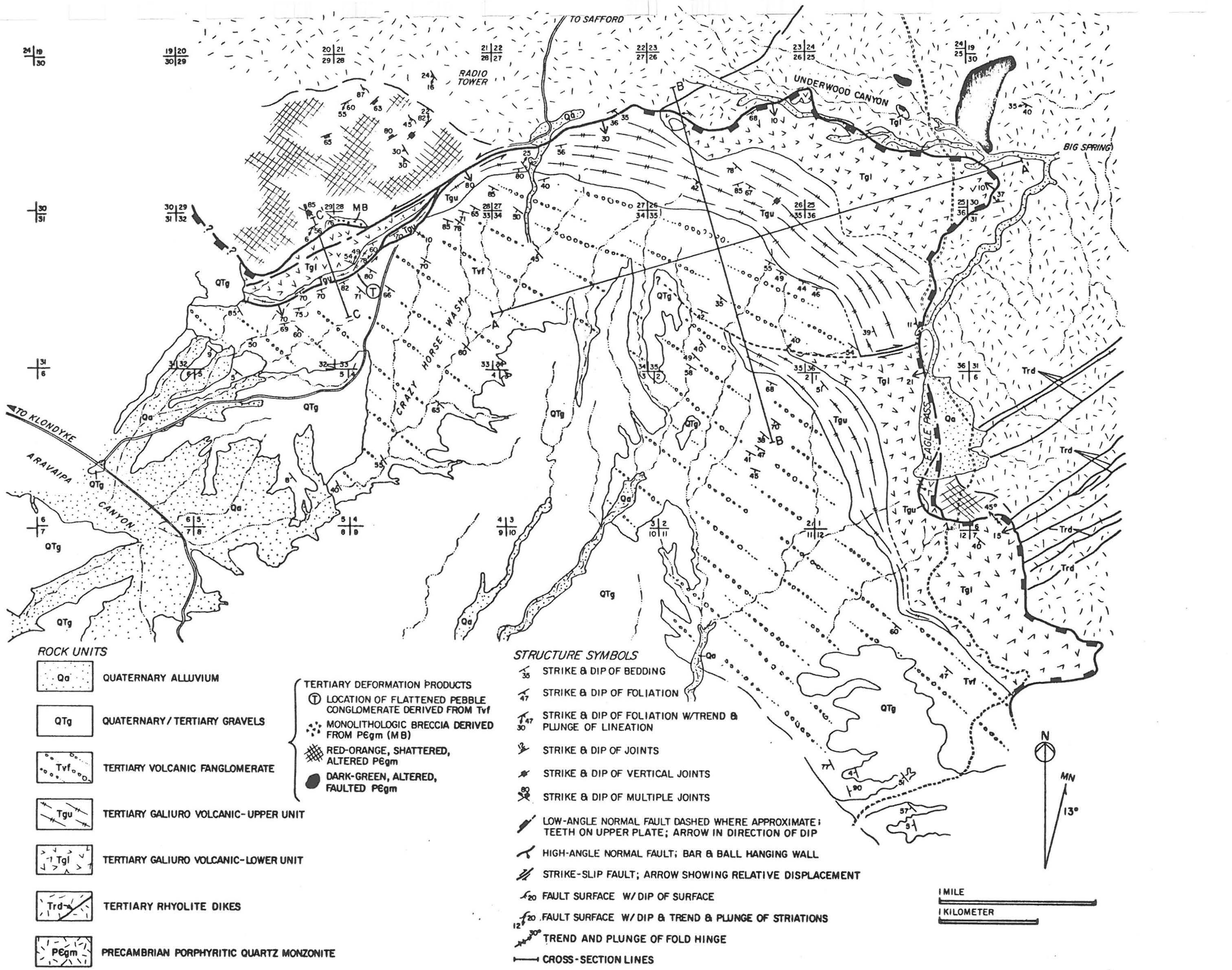

TO PHOENIX
SANTA TERESA MTNS.
ARAVIPA
SAFFORD
EAGLE PASS
GRAHAM MTNS.
70
ARAVIPA CANYON
GALIURO MTNS.
BONITA
STOCKON PASS
NEW MEXICO
ARIZONA
SAN PEDRO VALLEY
666
WINCHESTER MTNS.
SANTA CATALINA MTNS.
WILLCOX
TUCSON
RINCON MTNS.
10
N
0
40 MILES
0
25 Km
BENSON

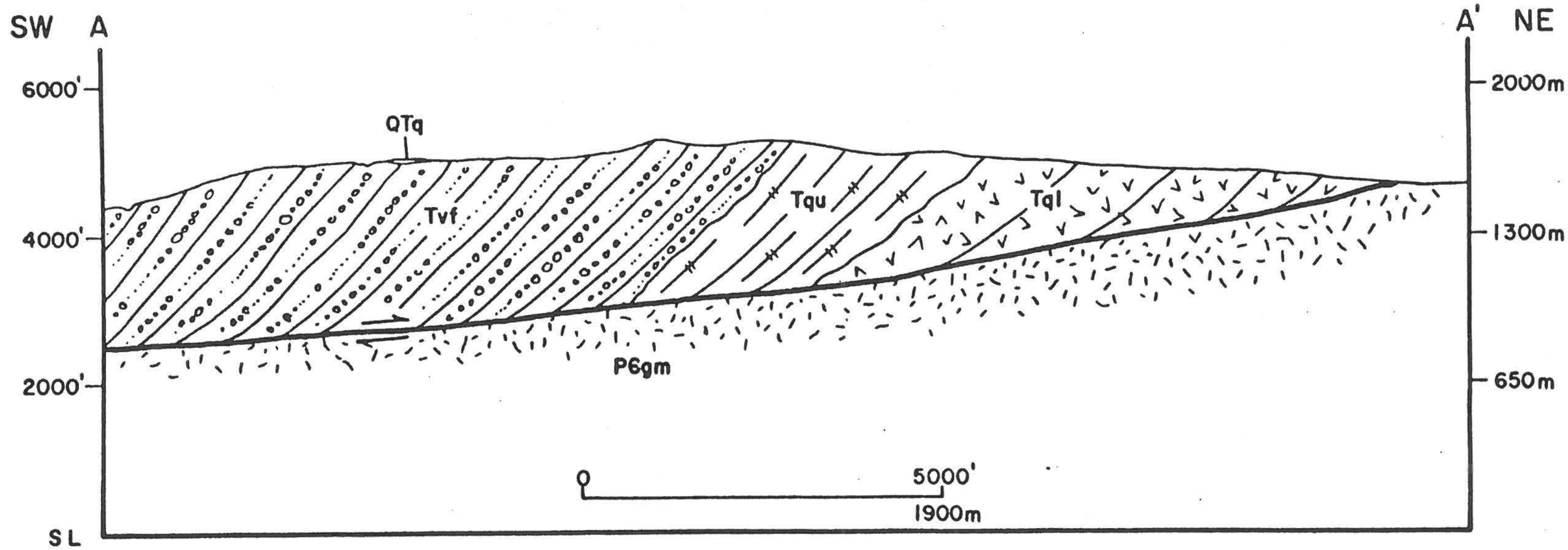
The Eagle Pass detachment occupies a broad, high pediment-saddle between the imposing Pinaleño Mountains to the southeast and the Santa Teresa Mountains to the north (Fig. 3). These mountains are a typical 'range' of the southern Basin and Range province. The pre-Basin and Range origin of the detachment is partly indicated by its presence within the range. Precambrian quartz monzonite crops out over most of the saddle between the Pinaleño and Santa Teresa Mountains. Toward the western edge of the saddle a low semi-circular ridge expresses lower - and middle-Miocene strata of the detachment which rest in fault contact on the quartz monzonite. Still further west, major westward-draining canyons carve into the faulted rock system and afford special views of its anatomy.

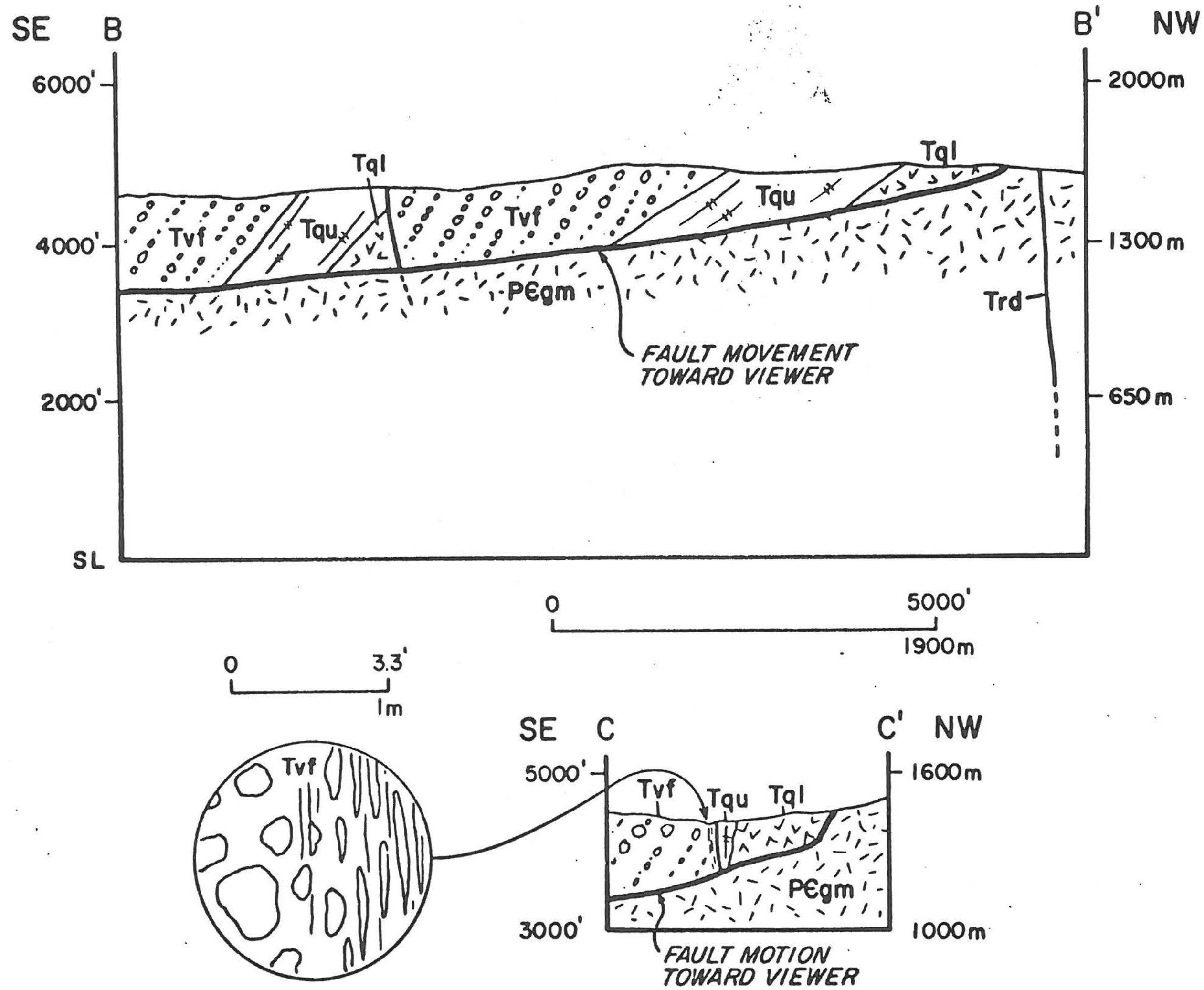
General Geology

The Eagle Pass detachment is underlain by a fault surface which displays a curved, convex-northeast trace approximately 11 km in length (Fig. 4). The footwall is everywhere Precambrian basement rocks, predominantly composed of the medium - to coarse-grained porphyritic quartz monzonite. This rock is probably part of the regional 1.42 $\pm$ b.y. anorogenic quartz monzonite suite (Silver and others, 1977). For the most part the quartz monzonite is light tan to off-white in color. Small patches of older Precambrian Pinal Schist occur as pendants in the quartz monzonite. A swarm of northeast-striking, vertical dikes of rhyolite and rhyolite porphyry intrudes the Precambrian quartz monzonitic footwall rocks, but the dikes are truncated at the fault surfaces and do not cut the upper-plate strata. Rehrig and Reynolds (1980) report a 25 m.y. K-Ar age based on analysis of biotite from these dikes.

The detachment occupies 25 km^2 and consists of Miocene volcanic and sedimentary rocks (Fig. 4). Three major units were identified in the detachment by Blacet and Miller (1978). The oldest is composed primarily of


TO SAFFORD
RADIO TOWER
UNDERWOOD CANYON
BIG SPRING
CRAZY HORSE WASH
EAGLE PASS
TO KLONDYKE
ARAVAIPA CANYON
MB
Qa
QTg
Tvf
Tgu
Tgl
Trd
ROCK UNITS
Qa QUATERNARY ALLUVIUM
QTg QUATERNARY / TERTIARY GRAVELS
Tvf TERTIARY VOLCANIC FANGLOMERATE
Tgu TERTIARY GALIURO VOLCANIC-UPPER UNIT
Tgl TERTIARY GALIURO VOLCANIC-LOWER UNIT
Trd TERTIARY RHYOLITE DIKES
P€gm PRECAMBRIAN PORPHYRITIC QUARTZ MONZONITE
TERTIARY DEFORMATION PRODUCTS
LOCATION OF FLATTENED PEBBLE CONGLOMERATE DERIVED FROM Tvf
MONOLITHOLOGIC BRECCIA DERIVED FROM P€gm (MB)
RED-ORANGE, SHATTERED, ALTERED P€gm
DARK-GREEN, ALTERED, FAULTED P€gm
STRUCTURE SYMBOLS
STRIKE & DIP OF BEDDING
STRIKE & DIP OF FOLIATION
STRIKE & DIP OF FOLIATION W/TREND & PLUNGE OF LINEATION
STRIKE & DIP OF JOINTS
STRIKE & DIP OF VERTICAL JOINTS
STRIKE & DIP OF MULTIPLE JOINTS
LOW-ANGLE NORMAL FAULT DASHED WHERE APPROXIMATE; TEETH ON UPPER PLATE; ARROW IN DIRECTION OF DIP
HIGH-ANGLE NORMAL FAULT; BAR & BALL HANGING WALL
STRIKE-SLIP FAULT; ARROW SHOWING RELATIVE DISPLACEMENT
FAULT SURFACE W/ DIP OF SURFACE
FAULT SURFACE W/ DIP & TREND & PLUNGE OF STRIATIONS
TREND AND PLUNGE OF FOLD HINGE
CROSS-SECTION LINES
N
MN
13°
1 MILE
1 KILOMETER


SW
A
A'
NE
6000'
4000'
2000'
SL
2000m
1300m
650m
QTq
Tvf
Tqu
Tql
P6gm
0
5000'
1900m


SE
B
B'
NW
6000'
4000'
2000'
SL
2000m
1300m
650m
Tql
Tqu
Tvf
PЄgm
Trd
FAULT MOVEMENT TOWARD VIEWER
0
5000'
1900m
3.3'
1m
SE
C
C'
NW
5000'
3000'
1600m
1000m
FAULT MOTION TOWARD VIEWER

numerous reddish-brown porphyritic andesite flows containing randomly oriented to radiating plagioclase phenocryst laths. Its maximum thickness is approximately 945 m. These andesite flows are known to workers in southern Arizona as 'turkey track porphyry' (Cooper 1961), and are thought to be indicative of uppermost Oligocene/lower Miocene volcanism. Overlying the andesite is a unit which consists mainly of a rhyolitic welded ash-flow tuff with thin rhyolite and rhyodacite flows. It also contains intercalated rhyolite breccias and coarse sedimentary breccia. Its thickness is approximately 550 m. The andesite and rhyolite, taken together, are undoubtedly part of the Galiuro Volcanics (Cooper and Silver, 1964; Simons, 1964; Creasey and Krieger, 1978), the type area for which is in the next range west (Fig. 3). The Galiuro Volcanics are nearly 1900 m thick in the central Galiuro Mountains. Creasey and Krieger (1978) report 11 K-Ar age determinations for the formation, ranging from 22 to 28 m.y. The youngest unit in the detachment is a very thick (4485 m!), red-brown, well indurated, poorly stratified fanglomerate and mud-flow breccia composed of heterogeneous boulders and cobbles of angular to subrounded andesite and rhyolite. Clasts range in diameter from several centimeters to more than 5 m. This coarse clastic sequence may be correlative with the Hell Hole Conglomerate (Simons, 1964), the type locality of which is on the east flank of the northern Galiuro Mountains (Fig. 3). Its age is uncertain, although in the Galiuro Mountains it depositionally overlies the Galiuro Volcanics. All of the strata within the detachment dip moderately to steeply to the southwest (Fig. 4). They are overlain unconformably by very gently dipping, moderately well indurated conglomerate of Plio-Pleistocene (?) age.

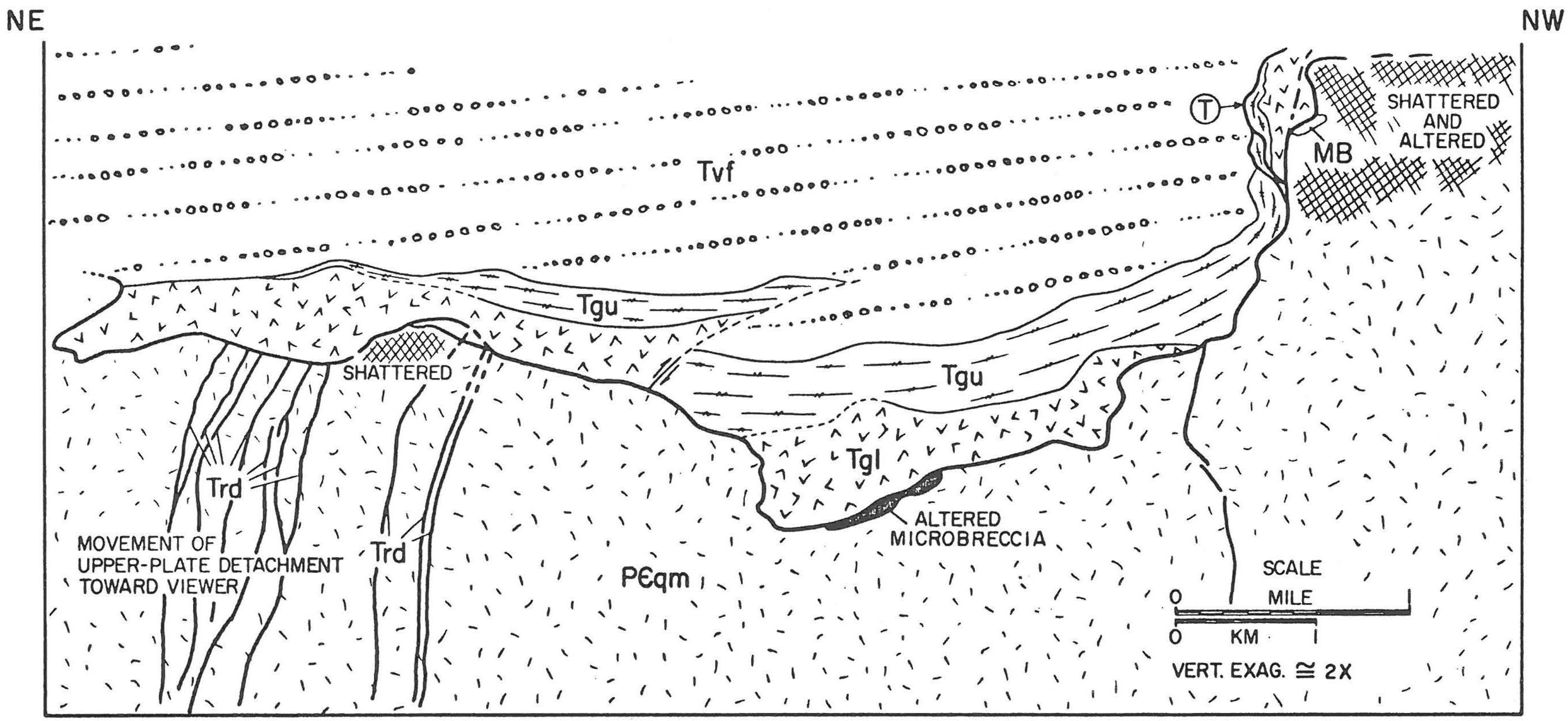
Structural Geology

The Eagle Pass fault trace is markedly sinuous (Fig. 4). Although the fault zone is continuous and curviplanar, it is convenient to think of it as composed of a number of segments which are distinctive in attitude. Along the northwest margin of the Eagle Pass detachment, the fault zone strikes N 50°E and dips steeply southeast. In the vicinity of Underwood Canyon, the fault swings in strike from northeast through southeast to south-southwest and assumes a very low angle of dip (less than 15°). From the vicinity of Eagle Pass to the south end of the study area, the fault maintains a low dip and strikes generally to the north, but there are short segments where the fault strikes roughly east-west. In effect the fault has a trough-like form, the axis of which plunges about 10° S50°W.

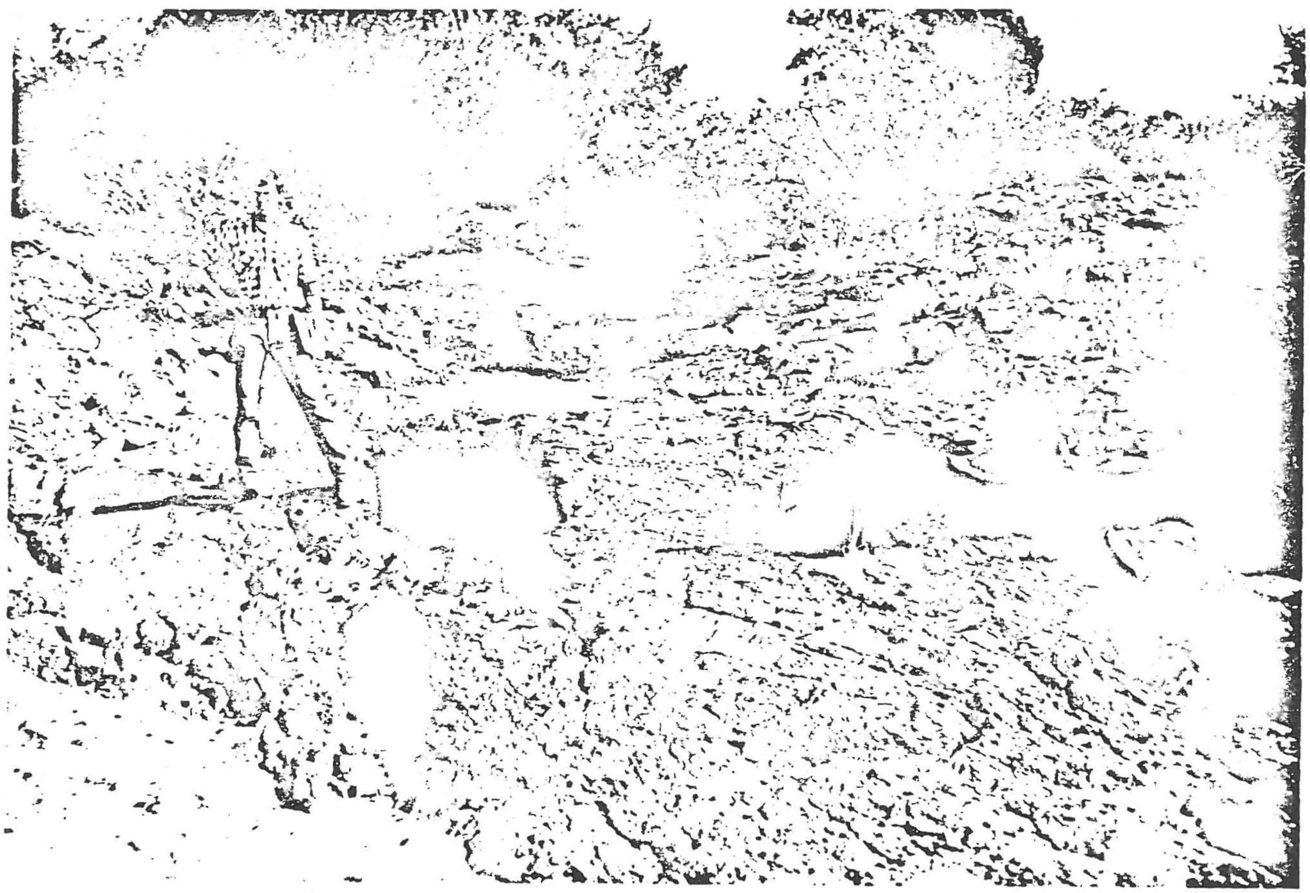
The trough-like form of the Eagle Pass fault has markedly influenced the orientation and style of deformation of both detachment and footwall rocks. Bedding within the detachment strikes consistently N50°-55°W and dips moderately to steeply southwest, except along the northwest margin. There, the volcanic rocks swing into parallelism with the fault, conforming both in strike and dip to the attitude of the fault zone. This shift in attitude close to the fault resulted in intense strain as expressed by an unusually high degree of fracturing, abundant striae reflecting adjustments along bedding and faults, and the mechanical breakdown of the volcanic units into discontinuous lenses (Fig. 4). Along the northwestern and northeastern parts of the fault the rhyolitic unit is locally juxtaposed directly against basement granite, without intervening andesite.

The fanglomerate is not as sensitive as the volcanic rocks to changes in fault attitude. The fanglomerate strikes N50°-55°W and dips steeply southwest throughout the area, except within tens of meters of the northwest margin of the detachment. There, the fanglomerate is locally dragged and faulted into parallelism with the northeast-striking, steeply southeast-dipping volcanic rocks, and with the fault zone itself.

Applying J. Hoover Mackin's (1950) down-structure method for viewing geologic maps proves to be a powerful approach in grasping the structural significance of the fault/bedding relationships in the Eagle Pass area. When viewed southwest down the gently dipping slope of the fault plane, the geologic map is transformed into a revealing structure section (Fig. 5). The irregularity of the surface of the detachment fault is seen more vividly. It appears mullion-like in terms of its macroscopic form (Wright and others, 1974). The northwest wall of the trough is especially steep and has served to localize subparallel fault zones in the detached strata. The steep wall may originally have been controlled by part of the fracture system along which the 25 m.y. dikes intruded. Along the northwest wall, detached strata are steeply dipping and attenuated. Locally, rhyolite abuts directly against the quartz monzonite. The footwall quartz monzonite is severely fractured along that margin for a distance of several hundred meters, and ferruginous alteration in the zone of intense fracturing has converted the terrane into one of distinctive red-orange hues. At the base of the detachment there is a keel below which the footwall quartz monzonite has been fractured, microbrecciated, and altered within an interval of 10±m. Alteration is both chloritic and ferruginous. Planar surfaces in the fault zone are locally metallic-gray/black in color and marked by tectonic polish and striations (Fig. 6). Just southeast of the keel, the andesite unit is attenuated and faulted in such a way that rhyolite again rests directly on granite. A few meters to tens of meters below the fault, the Precambrian quartz monzonite is absolutely undeformed and unaltered.



NE
NW
Tvf
Tgu
Tgu
Tgl
T
MB
SHATTERED AND ALTERED
SHATTERED
Trd
Trd
ALTERED MICROBRECCIA
MOVEMENT OF UPPER-PLATE DETACHMENT TOWARD VIEWER
PЄqm
SCALE
0 MILE 1
0 KM 1
VERT. EXAG. ≅ 2X

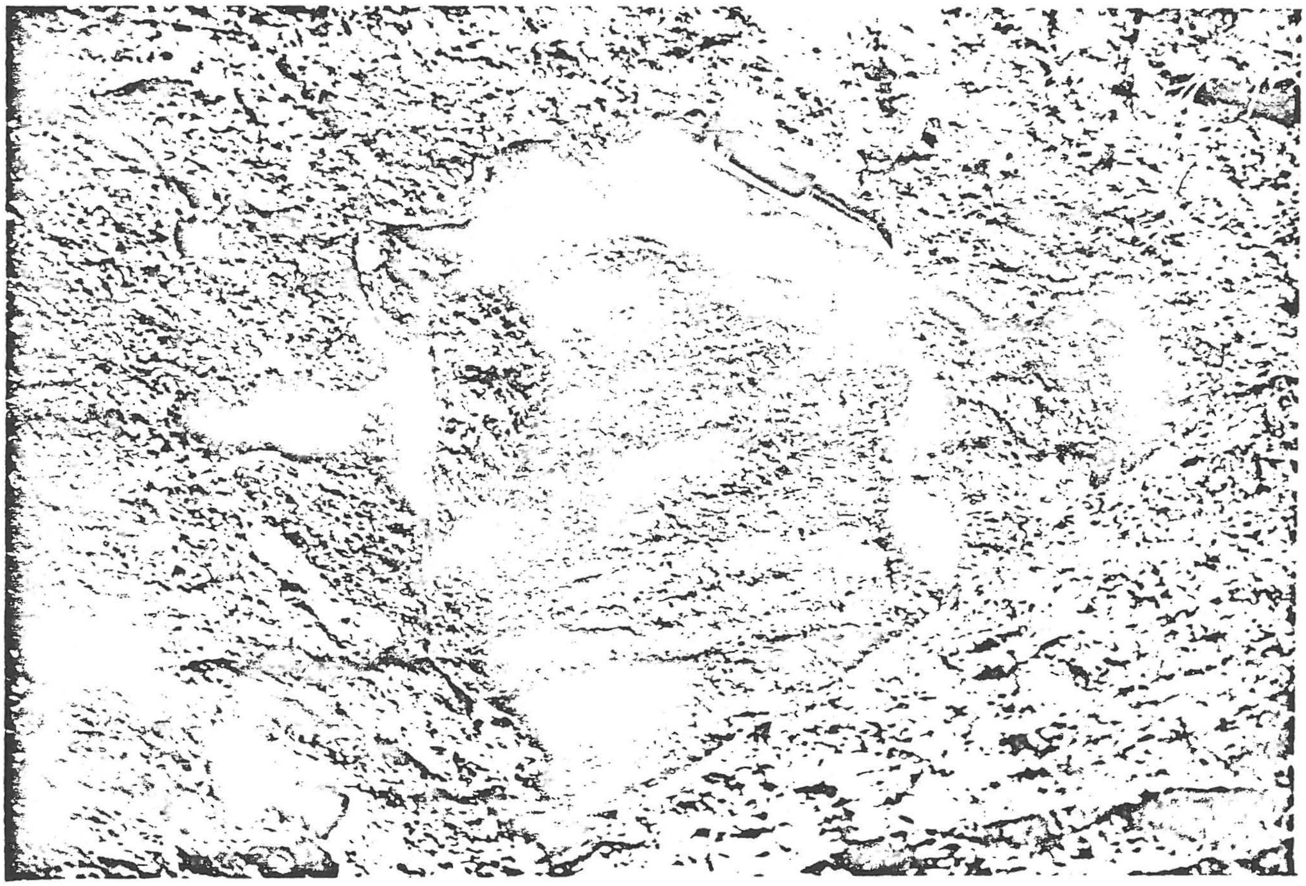


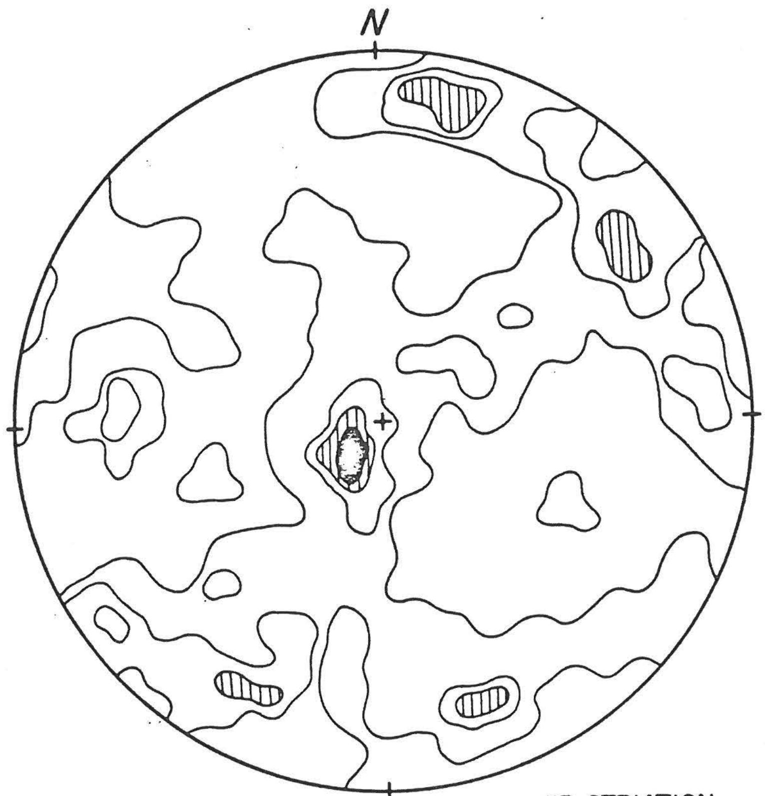
Movement Plan and Deformational Characteristics

A number of lines of evidence indicate that the upper-plate detachment strata were faulted into position by northeast-directed translation. The translation was accompanied by rotation of strata to steep southwest dips (Fig. 4B).

During faulting the upper part of the footwall quartz monzonite became highly fractured. Along the keel of the detachment, the footwall rocks were converted into a thin rind of microbreccia. Although there are data which suggest that the upper-plate strata were extended NE/SW during faulting, there is no evidence in the footwall rocks for such elongation. Rather, the footwall rocks appear to have been rigid when they were overridden by the detachment strata. Loci of the most intense alteration in the footwall quartz monzonite are positioned along the keel of the detachment and along the steep northwest trough-wall (Fig. 4). These locations were vulnerable to migration of hydrothermal solutions along the detachment strata/fault interface and (then) into fractured and brecciated quartz monzonite. This observation has been noted and emphasized by Heidrick and Wilkens (1980) on a regional basis.

Data which suggest that the line of fault movement was N40^{o}-50^{o}E/ S40^{o}-50^{o}W include (1) the orientation of the axis of the trough-like form of the fault, (2) the uniform N50^{o}W orientation of bedding strike in the tilted detachment strata, and (3) grooves and striae in deformed quartz monzonite, especially below the keel of the detachment. It is inferred that detachment strata were rotated and thus tilted during faulting. The axis of the fault trough and the average strike of bedding are essentially orthogonal. Grooves and striae measured mainly at the Big Spring locality (Fig. 7) are northeast-southwest in azimuth, with a great-circle distribution oriented approximately N50^{o}E (Fig. 8). There is scatter in the trend of





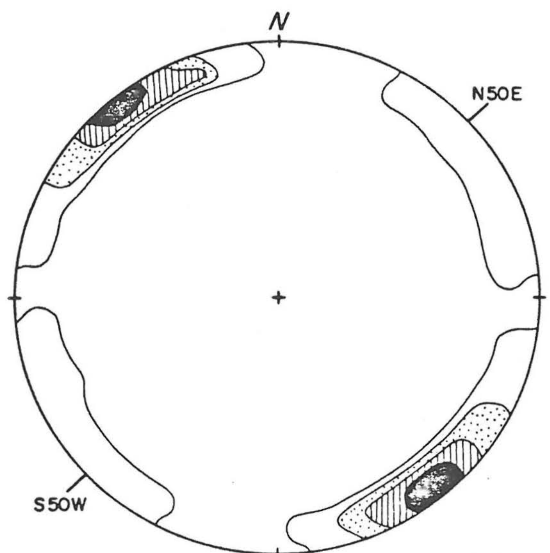
0-2-4-6% TOTAL DATA POINTS PER 1% AREA OF NET

113 STRIATION

striae, revealing a complex internal movement plan. Indeed, one gains the impression that faulting was accompanied by converging/diverging path movements which were highly dependent on local boundary conditions. In the quartz monzonite footwall, large mullion-like structures form elongate ridges, 20 to 40 m long, 10 to 20 m in wavelength, with polished, micro-brecciated, altered, striated surfaces. These are well developed near Big Spring and are oriented northeast-southwest. The form of these structures is identical to those mapped by Drewes (1978) along the Catalina fault in the Rincon Mountains. Joints were measured on polished and striated outcrops in the fault zone at Big Spring, and although the joint fabric cannot be used to evaluate movement plan, it may be significant that the stereographically plotted patterns are symmetrical with respect to a N40°E/S40°W line of movement (Fig. 9).

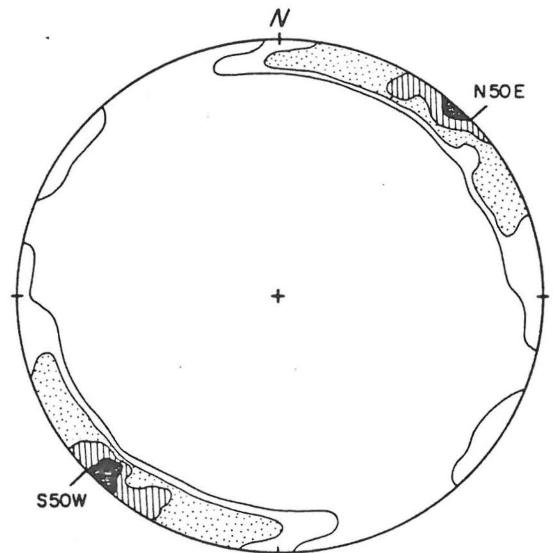
Striated surfaces in the detachment strata are not particularly abundant, except along the northwest faulted boundary of the detachment. Thus, it is difficult to corroborate the northeast/southwest inferred movement plan on the basis of minor structures in the detachment strata. A notable exception is a series of keystone fault blocks in rhyolite in Section 27 in the north central part of the area. Normal-slip faults mark the borders of several rotated blocks of rhyolite. These structures lie 20 m above the Eagle Pass fault. Their presence discloses that some northeast-southwest extension accompanied faulting. Faults in the andesite and rhyolite are difficult to find because of the poor quality of many of the hill-side exposures. Certainly these rocks are highly fractured. The fanglomerate is well exposed, but faults are absent except along the northwest border of the detachment. Jointing in the fanglomerate is mild to almost non-existent.

The main evidence that the <u>sense</u> of translation on the Eagle Pass fault was northeast-directed is an extraordinary zone of tear-faulting on the north-



0-5-10-15% TOTAL CUMULATIVE FRACTURE LENGTH PER 1% AREA OF NET

597cm CUMULATIVE FRACTURE LENGTH FRACTURE DENSITY EQUALS .12cm^{-1}



0-5-10-15% TOTAL CUMULATIVE FRACTURE LENGTH PER 1% AREA OF NET

1130cm CUMULATIVE FRACTURE LENGTH FRACTURE DENSITY EQUALS .06cm^{-1}

west boundary of the detachment. Evidence for strike-slip faulting is conspicuously displayed by horizontal striae on polished fault, fracture, and bedding surfaces in rhyolite (Fig. 10). The rhyolite forms a wall of steeply dipping strata which were dragged into parallelism with the quartz monzonite along this steep-dipping segment of the detachment fault (see Figs. 4, 5, and 11). The andesite, a much weaker rock than the rhyolite, was severely attenuated within this zone and is generally absent. Where it does crop out, exposures are poor and striae are not generally evident. The more competent rhyolite is pervaded by small conjugate strike-slip faults which disclose flattening perpendicular to the northeast line of strike.

Nowhere along the northwest tear-fault boundary of the detachment do the volcanic units display the N50^{o}W strike typical of the majority of the detached rocks. The volcanic rocks are dragged and smeared-out against the footwall quartz monzonite along the fault. In contrast, the fanglomerate maintains a N50^{o}W strike until within tens of meters of the tear-fault boundary. At that point, the layers are dragged abruptly westward, and/or the rock is transformed to flattened-pebble conglomerate by shear movements along very closely spaced faults, including microfaults (Fig. 12). The drag effects clearly reveal left-slip of the detachment along the fault interface. The development of stretched-pebble conglomerate at the the expense of 20 m.y. $\pm$ fanglomerate containing competent rhyolite clasts is surprising. The zones of simple shear are up to ten meters wide and are separated by intervening zones of approximately the same width where the fanglomerate is highly faulted and fractured, but not penetratively deformed. The "tectonite fabric" must have formed under dry, cool, relatively shallow conditions of deformation. The strain characteristics of the simple-shear zones are identical to those discussed by Ramsay and Graham (1976). Movement of the

fanglomerate detachment strata northeastward along the footwall of quartz monzonite must have been met by substantial frictional resistance to movement.

Another slip-line indicator is a single asymmetrical overturned fold, 2 m in amplitude, in highly deformed andesite just a meter above the Eagle Pass fault in the southeast corner of the map area. The fold trends N40^{o}W, and is clearly overturned northeastward.

One of the most peculiar structural relationships along the tear-fault boundary is the presence of a stratified monolithologic breccia which is sandwiched between the highly fractured/shattered quartz monzonite footwall and steeply dipping, northeast-striking andesite (see Fig. 4, location 'MB', Section 28). Layering within the breccia is concordant with that of the andesite. The exposure is approximately 200 m in trace length, and the thickness of the breccia is about 30 m, maximum. Virtually all of the clasts and matrix material of the monolithologic breccia is derived from the adjacent quartz monzonite. Clasts are as large as 1/2 m in size, and the layering in the breccia is primarily due to size-sorting of fragments and matrix. One interpretation for the origin of this rock is that it has been derived from comminution of the footwall. There is no rock like this anywhere in the detachment stratigraphy outside of the tear-fault zone. Its position within the tear-fault zone is suspicious in that it lies at an abrupt bend in the major fault (Fig. 4). The form of the bend, combined with the left-slip nature of movement on the fault, indicates that the location of this breccia was a site of significant compressive stress during fault movement. Secondly, the andesite and rhyolite just southeast of the monolithologic breccia are extremely highly fractured and faulted, yet most of the breccia is absolutely devoid of fractures, even joints. Only in a few places are single, through-going fractures evident, and these tend to off-set breccia clast by small displacements. Additionally, the clasts of quartz monzonite are so internally shattered that it is unlikely they could have been physically transported to their present locale in such condition. Conceivably penetrative comminution, granulation, and dry flow superceded regular fracturing and shattering of the quartz monzonite.

Although the formation of the breccia by structural comminution is attractive in some ways, the fact remains that the rock in question looks exactly like a water-lain conglomerate breccia. Insight into the origin of the breccia comes from Dokka (1979, 1980). His structural studies in the Newberry Mountains in the Central Mojave Desert have disclosed that the Miocene listric faulting there is accompanied by synkinematic formation of thick wedge-like deposits of monolithologic basement-derived breccias and conglomerates. Dokka has emphasized that such breccias are bound along the perimeter of the extended terrane, including sites of extensional grabens along tear-fault boundaries. This seems to be the most likely explanation for the monolithologic breccia at Eagle Pass, although the history of deposition and subsequent tilting of this breccia has not been worked out in detail.

Structural Interpretation

Rehrig and Reynolds (1980) examined in reconnaissance the Eagle Pass fault zone and concluded that deformation was achieved by listric normal faulting which produced southwest-to-northeast transport of upper plate detachment strata. Listric faulting along a northeast-dipping fault zone best explains the moderate to steep southwest dip of the upper-plate detachment strata. We agree with Rehrig and Reynolds' interpretation and regard the data and relationships presented herein as compelling support. We believe that the Eagle Pass fault is the sole of a large, regional, listric (?) normal fault zone whose upper reaches have been removed by erosion. That the fault zone was curved is an inference born from documented examples and published models concerning fault-rotated strata. We assume that the fault was originally horizontal or gently east-dipping, and attained its westerly dip, its present attitude, by rotation along yet deeper listric faults and/or by arching of the Pinaleño Mountains and adjacent terranes.

Movement indicators support the interpretation of northeastward transport. The marginal tear-fault boundary is perceived as a zone of accommodation to differential relative movement within the larger mass. The volcanic rocks and fanglomerate in the detachment are identical to Galiuro and Hell Hole stratigraphy in the Galiuro Mountains, and this too strongly supports a westward provenance for the detachment.

Progressive faulting resulted in rotating and lowering of the sedimentary and volcanic rocks onto Precambrian quartz monzonite footwall. By the time the andesite and rhyolite reached the level of basement, the top of the footwall of the fault was already marked by significant curvature and irregularity, including the large, mullion-like trough. The volcanic rocks were smeared concordantly against the southeast flank of the mullion as the detachment moved glacier-like in left-slip fashion past the rigid quartz monzonite footwall. Dragging of the volcanic rocks along the gently dipping parts of the detachment fault may have been responsible for the decrease of dip of the volcanic layering (Fig. 4). The fanglomerate exposed along the northwest margin of the Eagle Pass detachment is believed to express a stage of the faulting in which that rock first felt the effect of interference with the basement obstruction. As the fanglomerate was lowered, only the closest fanglomerate layers responded to the frictional resistance to steady northeastward movement. The northeast transport must have been on the order of kilometers, and thus it is not surprising that extreme granulation and brecciation of quartz monzonite were locally achieved.

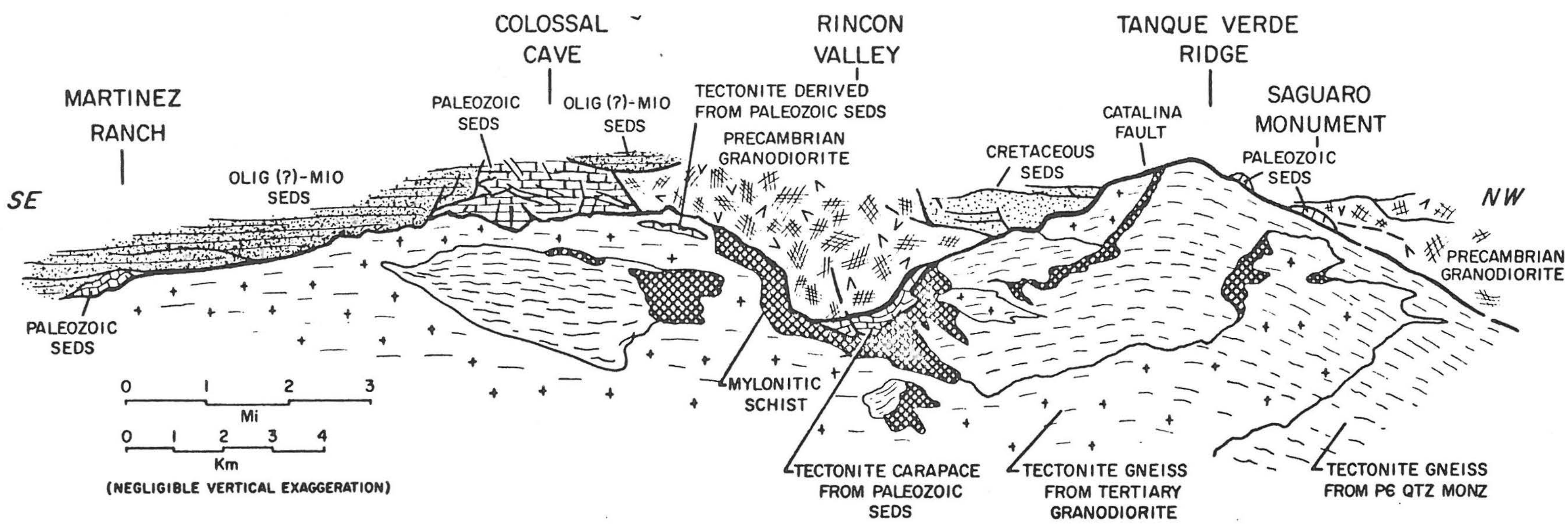
SPECULATIONS

Evaluating the nature and origin of mid-Miocene detachment faulting is almost inseparable from trying to unravel the tectonic evolution of metamorphic core complexes (Crittenden and others, 1980). The presence of detachments contributes to the distinctiveness of metamorphic core complex terranes even though the detachments are by no means restricted to such complexes. The paradox which confronts workers who are interested in evaluating the relationship, if any, between mid-Miocene faulting and the evolution of metamorphic core complexes is reconciling the following facts: (1) mid-Miocene faulting post-dates the formation of lineated mylonitic tectonite, but (2) almost everywhere in the southern Basin and Range the slip-line direction which describes transport of detachments is virtually parallel to penetrative lineation in the closest tectonite terrane of core complex affinity (see Table 1).

The Eagle Pass detachment setting illustrates these points. The fault lies just 5 km west of exposures of lineated tectonite. The tectonite forms a northeast-dipping lensoidal to tabular zone, several hundred meters thick, which marks the northeast margin of the Pinaleño Mountains (Fig. 3) (Swan, 1976, Davis, 1977, 1980). The lineated tectonite gives way upward and westward at an elevation and location which coincides closely with the Eagle Pass datum. The orientation of mineral lineation (N40^{o}-50^{o}E) in the tectonite is parallel to the direction of slip determined for the Eagle Pass detachment and the northeast sense of simple-shear in the mylonitic tectonite is the same as the sense of movement on the Eagle Pass detachment fault. This latter point is especially significant. Lineated tectonite in most metamorphic core complex terranes, particularly where the protolith is equigranular or porphyritic quartz monzonite, usually lacks fold structures which would otherwise disclose slip-line paths. However, the tectonite along the flank of the Pinaleño Mountains is one of the few exceptions. M. M. Swan in 1977 first pointed out the northeast vergence of folds to G. H. Davis. More recently C. H. Thorman (1980, pers. comm.) has firmly documented the northeast sense of simple-shear disclosed by the tectonite.

Although these relationships might be fortuitous, unique to the Pinaleño Mountain region, the same coordination characterizes the detachment/ tectonite terrane on the west side of the Galiuro Mountains (Fig. 3). But there the sense of movement is southwestward, not northeastward. Detachments of upper-plate strata all along the San Pedro Valley (Cornwall and Krieger, 1975; Krieger, 1974b,c), including the eastern-most Rincon Mountains (Drewes, 1975; Lingren, 1981) (Fig. 3), dip northeast and reflect northeast-to-southwest translation and rotation. The detachments in the Tanque Verde/Rincon Mountains have been reinterpreted as having undergone southwestward translation in mid-Miocene time (Davis, 1975; Davis, Gardulski, and Anderson, 1981), and as in the Eagle Pass area these detachments appear to have been translated on and against large mullion-like arches and troughs in underlying footwall (Fig. 13). The direction of translation of detachments in the San Pedro Valley and vicinity matches the N60$^{o}\pm$E trend of penetrative mineral lineation in the Rincon-Santa Catalina-Tortolita complex (Davis, 1980). Fold data in tectonite gneiss and carapace rocks in that complex are interpreted to reveal a southwesterly sense of simple shear (Davis, 1981).

The regional symmetry of these relationships is very striking. A northeast/southwest structural profile of the Rincon-Galiuro-Pinaleño region would reveal a 100 km broad distended crustal block whose outer

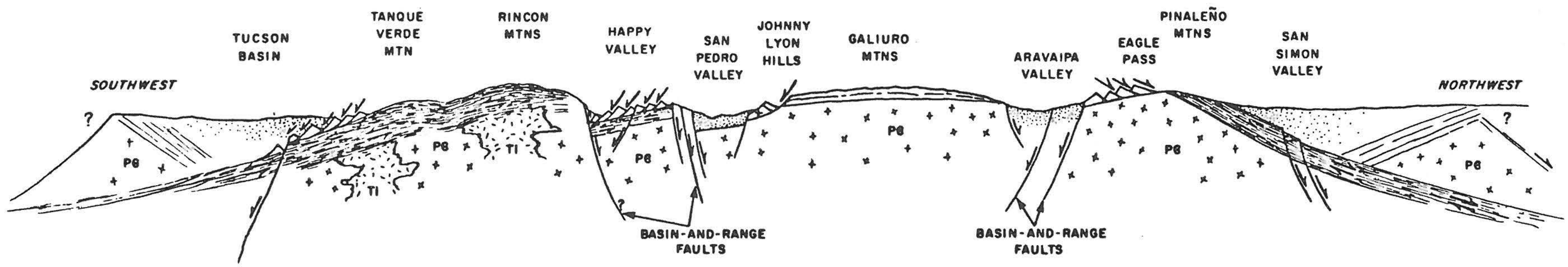


SE
MARTINEZ RANCH
OLIG (?)-MIO SEDS
PALEOZOIC SEDS
COLOSSAL CAVE
PALEOZOIC SEDS
OLIG (?)-MIO SEDS
RINCON VALLEY
TECTONITE DERIVED FROM PALEOZOIC SEDS
PRECAMBRIAN GRANODIORITE
TANQUE VERDE RIDGE
CATALINA FAULT
CRETACEOUS SEDS
SAGUARO MONUMENT
PALEOZOIC SEDS
NW
PRECAMBRIAN GRANODIORITE
MYLONITIC SCHIST
TECTONITE CARAPACE FROM PALEOZOIC SEDS
TECTONITE GNEISS FROM TERTIARY GRANODIORITE
TECTONITE GNEISS FROM P€ QTZ MONZ
0 1 2 3
Mi
0 1 2 3 4
Km
(NEGLIGIBLE VERTICAL EXAGGERATION)

margins are marked by outward-dipping tectonite shear zones (Fig. 14). Detachments of faulted, rotated strata occur along and just inward from these margins, dipping inward toward the medial part of the faulted upper crustal block. Sense of translation of the detachments appears to match the sense of simple shear revealed in minor structures within the mylonitic tectonite.

Davis (1977, 1980; Davis and Coney, 1979) proposed that mechanical interpretation of the formation of the tectonite can be understood in the context of mega-boudinage. Applied to the setting of interest here, the Rincon-Galiuro-Pinaleño block can be envisioned as an upper-crustal mega-boudin bounded by ductile normal shear (tectonite) zones on the southwest and northeast. The ductile shear zones accommodated regional extension in early to mid-Tertiary time. Cataclasis and ductile flow to form mylonitic tectonite was a response to hot plastic extension (and necking) of the crust. The effect of the stretching was to produce regional basins bounded by growth-faults and filled by Oligocene/early Miocene redbeds and volcanic rocks, like those studied at Eagle Pass. Some detachments now resting on tectonite may have begun to descend to their final positions in early to mid-Tertiary time, and while lineation was developing. In contrast, young detachments like that at Eagle Pass were emplaced after the formation of the tectonite fabric.

The physical form and geometry of the mid-Miocene detachments suggest that they are the exposed remains of a system of gigantic upper-crustal slump blocks which moved on a family of curved (?) normal faults. The faulting probably reflects regional extension. But the detachments were differentially translated over non-extending, rigid footwall rocks at the present levels of exposure. We believe that the nature of the detachment faulting was influenced by earlier tectonic movements which produced the


SOUTHWEST
TUCSON BASIN
TANQUE VERDE MTN
RINCON MTNS
HAPPY VALLEY
SAN PEDRO VALLEY
JOHNNY LYON HILLS
GALIURO MTNS
ARAVAIPA VALLEY
EAGLE PASS
PINALEÑO MTNS
SAN SIMON VALLEY
NORTHWEST
P6
Ti
BASIN-AND-RANGE FAULTS
BASIN-AND-RANGE FAULTS

mylonitic tectonite. We speculate that large crustal displacements achieved during extension-induced formation of tectonite produced a geometric and/or dynamic condition which strongly influenced subsequent detachment faulting. Detachment rocks were translated in the direction and sense of the earlier displacements which had accommodated the formation of tectonite.

For the Eagle Pass detachment in particular, the volcanic and sedimentary rocks may have "slumped" from a thick horizontal sequence of Tertiary rocks resting nonconformably on Precambrian granite. An undisturbed nonconformable relationship between Precambrian and Tertiary rocks occurs in the central part of the Galiuro Mountains (Fig. 15). A listric normal fault zone could have cut through the Tertiary volcanic and sedimentary rocks, cutting into the nonconformity and pealing off only the Tertiary rocks (Fig. 15). Alternatively, the fault could have cut down through the unconformity, thus capturing Precambrian granite in the detachment block (Fig. 15). After significant field effort we have not found Precambrian rocks in the detachment rocks at Eagle Pass. However, in the eastern Rincon Mountains (Drewes, 1975; Lingrey, 1981) the Rincon Valley granite of Precambrian age may be an example of basement caught up in mid-Miocene detachment faulting (Fig. 13).

Detailed kinematic reconstructions in detachment settings will be difficult because the degree of rotation of footwall rock beneath detachments is hard to evaluate. Both the rocks <u>and</u> the faults have rotated during deformation. Although the footwall rocks at Eagle Pass appear rigid and stationery, they may have been rotated significantly along yet deeper faults, in a way which is consistent with Anderson's interpretation of the nature of deep portions of listric fault systems (Fig. 1). In the end, understanding of these terranes may largely depend on facts

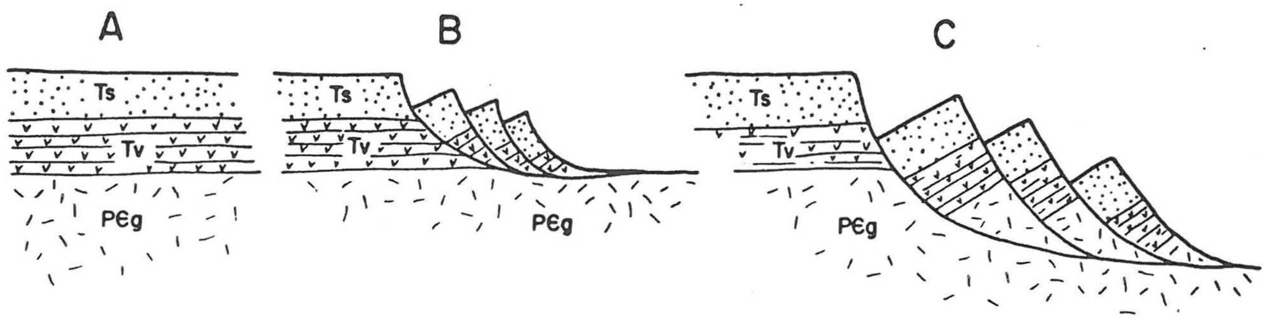


Fig 15

related to the various stacked levels of decoupling.

ACKNOWLEDGEMENTS

We are indebted to a number of people, especially P.M. Blacet and S.T. Miller who mapped the quandrangle within which the Eagle Pass detachment lies, and W.A. Rehrig and S.J. Reynolds who introduced one of the authors (Davis) to the area. R.E. Anderson, in 1979, led a tour of the Black Canyon region to structure-tectonics students and faculty of The University of Arizona, and what was learned on that trip aided us immeasureably in understanding the Eagle Pass geology. We are grateful to G.A. Davis and E.G. Frost for sharing with us their understanding of dislocational terranes in southeastern California. Finally, we acknowledge the financial support for the project provided by NSF grants #EAR 76-84167 and EAR 78-23404. Max Crittenden, Jr., Eric Frost, and Chuck Thorman critically reviewed an earlier draft of this manuscript and offered very useful suggestions for its improvement. We thank them for their help.

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CAPTIONS FOR ILLUSTRATIONS

1. Cross-section showing style of listric normal faults in the Black Canyon region, after Anderson (1978). Tertiary sedimentary and volcanic strata rest directly on Precambrian basement.

2. Map showing geography of region of mid-Miocene detachments.

3. Map showing location of Eagle Pass area.

4. A. Geologic map of Eagle Pass area, by J. J. Hardy, Jr. B. Structure sections showing geologic relationships.

5. Down-structure view of geologic map of Eagle Pass area. Direction of view is southwestward.

6. Photograph of microbrecciated and altered granite on Eagle Pass fault at Big Spring locality. View is northward. Note northeast-trending striations on rock surface in foreground.

7. Slickenside striae on surface of microbrecciated granite at the Big Spring locality.

8. Lower-hemisphere pole-density diagram of slickenside striae on Eagle Pass fault at Big Spring locality and vicinity.

9. Lower-hemisphere pole-density diagram of poles to fractures at two inventory areas in microbrecciated granite at the Big Spring locality of the Eagle Pass fault.

10. Fault surface in rhyolite along the northwest margin of the Eagle Pass detachment. View is westerly.

11. Natural wall of vertically inclined resistant rhyolite along the zone of tear faulting at the northwest margin of the Eagle Pass detachment. View is northwestward. Smooth outcrops in right foreground represent fanglomerate striking directly into the fault-bounded rhyolite wall.

12. Faulted clast of rhyolite in fanglomerate within shear zone along the tear-fault margin of the Eagle Pass detachment.

13. Southwestward-directed down-structure view of Drewes' (1977) geologic map of Tanque Verde/Rincon Mountains. Note mullion-like nature of detachment (décollement) surface, the Catalina fault.

14. Generalized, schematic structure section showing geometric and kinematic symmetry of tectonite zones, detachment surfaces, and detachments in upper-crustal block between the Pinaleño Mountains on the northeast and the Tanque Verde Mountains on the southwest.

15. Interpretive kinematics of detachment faulting. See text.

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STRUCTURAL AND STRUCTURAL-PETROLOGICAL CHARACTERISTICS OF SOME METAMORPHIC CORE COMPLEX TERRANES IN SOUTHERN ARIZONA AND NORTHERN SONORA

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INTRODUCTION

Cordilleran metamorphic core complexes in southern Arizona and northern Sonora owe their character to disharmonic structural relationships. Four distinctly different structural levels produce the disharmony: these are occupied by protolith, tectonite, décollement zone, and detachments (Davis, 1980b) (Fig. 1).

The lowermost level is country-rock protolith. The protolith may consist of any rock which falls within the age-range of Precambrian through (at least) early Tertiary. Structures in rocks of the protolith conform to whatever habit typifies rocks of the region. Upward, the country-rock protolith is progressively or discontinuously transformed into tectonites, commonly mylonitic, characterized by low-dipping foliation and penetrative unidirectional lineation (Fig. 1). Minerals, textures, and structures of the protolith are overprinted thoroughly by the tectonite fabrics. Granites and quartz monzonites are transformed into mylonitic gneisses (Fig. 2A); sedimentary rocks are converted to carapace-like assemblages of tectonite marble, quartzite, calc-silicate rock, and flattened-pebble conglomerate (Fig. 2B); and low-grade metasedimentary and metavolcanic schists are deformed into tectonite schists with superposed, foliate fabrics (Fig. 2C). The tectonite foliation and lineation indiscriminately overprint all rocks and structural components of the protolith assemblage.

The next level upward is distinguished by intensive microbrecciation and chlorite-iron-epidote alteration of mylonitic tectonite within relatively thin décollement zones (Figs. 1 and 2). The décollement zones thus consist of microbrecciated and altered mylonitic tectonite. Commonly the original subhorizontal foliation and lineation is rotated to moderately or steeply dipping attitudes. The top of the décollement zone is usually capped by a massive one to two-meter thick ledge of fine-grained microbreccia, the upper surface of which is the décollement surface (Figs. 1 and 2).

Above the décollement are detachments of allochthonous non-tectonite country rock. As used here the term "detachment" refers to allochthonous, commonly rotated, hanging-wall rocks resting in very low-angle fault contact on

Davis ... (...)

the structurally deeper, generally older, footwall rocks. The term detachment is extracted from the expression "detachment fault", introduced by G.A. Davis and others(1979, 1980) to describe low-angle faults which mark the base of mid-Miocene allochthons in the Whipple-Buckskin-Rawhide Mountains terrane in southeastern California and western Arizona. The detachment rocks clearly have been faulted from their original underpinnings. They may be of any lithology and may range in age from Precambrian to mid-Miocene. Structural deformation in the detachment rocks is diverse, including faulting and rotation and/or shattering and/or folding.

Given the unusual nature of the disharmonic relationships, it should be no surprise that the origin of the Cordilleran metamorphic core complexes is hotly debated. Central issues are (1) time(s) of formation of mylonitic tectonite, (2) dynamic conditions under which the mylonitic tectonite evolved, (3) tectonic significance of décollement, and (4) mode of emplacement of detachments. Pertinent descriptive and interpretive details are available in a number of published articles (Coney, 1974, 1979, 1980; Crittenden and others, 1980; Davis, 1975, 1977, 1980a,b; Davis and Coney, 1979, 1980; Davis and other, 1979, 1980; DeWitt, 1980; Drewes, 1978, 1979; Rehrig and Reynolds, 1980; Reynolds and Rehrig, 1980; and Thorman, 1977).

The purpose of this paper is to try to bring into sharper focus the structural and structural-petrological properties of the disharmonic components of the core complexes. We attempt this through presentation of details of some core-complex study areas in northern Sonora and southern Arizona(Fig.3): Sierra Mazatan, Sierra Magdelena, Pozo Verde Mountain, Coyote Mountains, Tanque Verde Mountains, and the Rincon Mountains. The descriptions are preceded by a section on "Interpretive Remarks" so that the chief components of core complexes (protolith, mylonitic tectonite, décollement zones, and detachments may be viewed in a broader, tectonic context. The interpretive remarks are drawn largely from Davis (1980a).

INTERPRETIVE REMARKS

The effect of major plutonic invasion into the upper crust in the early Tertiary set the stage for the unusual style of penetrative deformation recorded in the metamorphic core complex terranes. There are few localities where metamorphic core complexes lack any sign of known or inferred Tertiary intrusive bodies. The thermal input, in effect, softened the country rocks, more in some places than in others. The rheology of the cooling plutons and the locally heated country rocks made the systems vulnerable to ductile flow. The effect of intrusion was to convert the regional country rock into domains of cool rigid rocks and contrasting domains of hot ductile rocks.

Thermal effects were superimposed on a mechanically anisotropic crust. The upper crustal rocks did not possess laterally continuous litho-tectonic units. Instead, the combination of Jurassic and Laramide faulting in the southern Basin and Range placed Precambrian basement laterally against Paleozoic and/or Mesozoic layered strata; folded, thrusted strata against domains of homoclinal strata; and complete sections of Precambrian-Paleozoic-Mesozoic rocks against section where Cretaceous strata rest nonconformably on Precambrian basement (Davis, 1979). In fact, parts of southern Arizona and northern Sonora even seem to lack Precambrian rock altogether (Anderson and others, 1980). The combined influences of thermal and mechanical anisotropy seem to have predetermined the location of individual core-complex terranes.

Given the thermal and mechanical condition of the upper crustal rocks, what regional tectonic movements gave rise to the penetrative componental movements which produced mylonitic tectonite? The mylonitic tectonite fabric is considered to be a flattening/extension fabric, with direction of extension parallel to penetrative mineral lineation (Davis, 1977, 1980a; Davis and Coney, 1979; Davis and others, 1975; Coney, 1979, 1980. The tectonite fabric is interpreted to have formed by ductile normal shearing within gently dipping curvitabular zones of simple-shear (Davis, 1980b, 1981). These zones reflect regional tectonic movements which stretched the upper-crustal rocks, producing major extension (Davis and Coney, 1979). The cause of the stretching is not known. It could be a "passive" product of regional strike-slip movements. It could be a result of a collapse of what had been a shallow-dipping Benioff zone at the culmination of major Tertiary plutonism (Coney, 1979, 1980; Davis and Coney, 1979). Any interpretive theory must explain the consistent trends of fabric elements across more than 100,000 km^2 in rocks as young as 55 m.y. (Anderson and others, 1980) and maybe younger (Reynolds and Rehrig, 1980).

As a response to stretching, the heterogeneous upper crust was partitioned into an array of rigid blocks with ductile margins. "Megaboudin"-like rigid crustal blocks became marked by shoulders and necks of normal shearing characterized by significant thinning and flattening (Davis and Coney, 1979). At depth, these zones of normal shearing formed mylonitic tectonite at the expense of Precambrian basement. At higher structural levels these zones converted Paleozoic and Mesozoic strata into tectonite. The projection of the shear zones to the surface may have been marked by distributed, imbricate normal faulting, producing growth-fault basins of sediment accumulation (Davis and Coney, 1979). Downward, the zones of tectonite may have been transitional into horizontal regimes of lamellar flow.

As cooling of the system took place, some already-formed mylonitic tectonite was disrupted, rotated, and microbrecciated by closely spaced, mesoscopically penetrative, normal-slip listric (?) faults. This process could have been an upper-level, brittle counterpart of still-active mylonite tectonite formation at greater depth. The kinematic coordination is perfect. While mylonitic tectonite and microbrecciated mylonitic tectonite were being formed at depth, the rocks which now occupy detachments began to move downward and laterally, and thus rocks once far removed from deep normal shear zones were translated ultimately to positions directly above the mylonitic tectonite (Davis, 1977). This tectonic denudation began to fashion the dramatic disharmony between rocks of the mylonitic tectonite and the rocks above.

Abrupt cooling of the mylonitic tectonite rocks is recorded in mid-Tertiary K-Ar and fission-track ages (about 25 m.y.) (see Crittenden and others, 1980). Most workers have ascribed this cooling event to "uplift". The structural implications of such uplift are not clear. Perhaps the 25 m.y. ± cooling ages record development of the décollement and its ledge of microbreccia as a result of major imbricate normal faulting. The normal faults coalesced at or near the uppermost level of mylonitic tectonite. The trigger for this denudational faulting may be been resurgence (or continuation) of faulting, especially near the boundaries of megaboudin-like tectonic units. The faulting accommodated further extensional strain in the upper crust as a whole. Ignimbritic volcanics (25 m.y. ±) exploded out of deeply rifted parts of the crust throughout the entire belt of metamorphic core complexes (Davis and Coney, 1979).

It is important to consider that all of the events which proceeded from about 55 m.y. (?) to 25 m.y. may have accompanied and/or may have been produced by the concurrent formation of mylonitic tectonite (at some depth). However, the final emplacement of detachments in post-25 m.y. time, largely mid-

Miocene, postdated formation of all exposed tectonite. Oddly enough, the slip-line direction for translation of the detachments identically parallels the trend of penetrative lineation in nearby or underlying tectonite (Davis and Hardy, 1981).

The final emplacement of the detachments in mid-Miocene time probably accompanied continued regional extension. The faulting may in part have been influenced by geometric and/or dynamic conditions produced during the earlier inferred normal-slip descent of hanging-wall terranes above zones of mylonitic tectonite. The pure mid-Miocene detachments, i.e., those containing only mid-Miocene strata, were translated perfectly parallel to the direction of the penetrative lineation in the closest mylonitic tectonite, and for southeastern Arizona examples apparently (?) in the same sense of slip as that of simple-shear in the tectonite (Davis, 1981). The structural profiles of the mid-Miocene detachments convey the properties of slump- or toreva-block tectonics. Tempting as it is to explain the fault profiles as the result of stretching of brittle multilayers atop a ductile stretching medium, the geologic relationships disclose that many of the fault-block detachments now rest on footwall rocks which show no sign of concomitant penetrative flow movements: differential translation took place along low-dipping fault contacts between rigid but not necessarily stationery, basement and overlying sedimentary and volcanic rocks. Rocks nearest the margins of boudin-like crustal blocks were translated along curved (?) normal faults in the direction of the earlier stretched and descended terranes. All of these events appear to have postdated Laramide compression-induced shortening and preceded classical Basin and Range high-angle normal faulting.

SIERRA MAZATAN

Introductory Comments.

Sierra Mazatan lies 70 km east of Hermosillo, the capital city of Sonora (Fig. 3). It is a roughly circular range with a diameter of 12 km; the crest of the range is marked by knobs of granite that rise a few tens of metres above a rolling surface of low relief which in places is dissected by shallow stream valleys. As seen on aerial and orbiter photographs, Sierra Mazatan is one of the most distinctive and easily recognized physiographic features in northwestern Mexico. It rises nearly 1,000 m above the surrounding terrane and its circular flanks which outline the crest are emphasized by a cover of dark oak trees above 600 m elevation, in contrast to the lighter shades of brown and yellow of the adjacent desert.

Sierra Mazatan exhibits a strikingly smooth profile that is broken along the western flank by jutting outcrops of carbonate beds of Paleozoic age. These beds are structurally the highest elements of the range. They rest upon a strongly lineated surface developed upon a thin layer of mylonitic gneiss which is transitional downward into somewhat less strongly sheared porphyritic granodiorite gneiss. U-Pb isotopic analyses on zircon from the granodiorite gneiss yield an interpreted age of 58 ± 3 m.y. (Anderson and others, 1980). This Tertiary granite records the southernmost known extent of Tertiary deformation characteristic of more northerly Cordilleran metamorphic core complexes.

Protolith.

Sierra Mazatan porphyritic granodiorite and its wall rocks (?) are the major rock elements most affected by deformation. The extent of the gneissic fabric is restricted largely to the west flank of Sierra Mazatan, whereas the rocks on the top and eastern flank of the range contain scant evidence of the strong Tertiary overprint.

Tertiary granitic protolith for the gneiss is composed of rather uniform porphyritic granodiorite which is locally transitional into a more leucocratic muscovitic phase. Along the west flank of Sierra Mazatan the porphyritic granodiorite becomes gneissic and interfingers with layered quartzo-feldspathic, biotite-rich gneiss and locally with quartzitic gneiss. These heterogeneous units are interpreted to be country rock into which the granodiorite was emplaced. The protolith of these gneisses is unknown. The gneisses do not contain abundant marble, quartzite, or graphitic schist and therefore it seems unlikely that they were derived from the Paleozoic carbonates or the coal-bearing Triassic beds which are found in this part of northern Sonora.

Sierra Mazatan lies near the southern extreme of a region dominantly underlain by granitic rocks. The granitic rocks appear to form a batholithic terrane which underlies a 100+ km x 25+ km region marked by north-trending ranges. The batholith is approximately bisected by Rio Sonora, and the walls of its canyon reveal excellent exposures of the granite. Granitic rocks along the flanks of this batholith record weak to moderate foliation.

Mylonitic Tectonite.

Mylonitic rocks are well exposed along the west flank of Sierra Mazatan particularly where intermittent streams have incised arroyos. The mylonitic granodiorite gneiss is characterized by classic mortar texture in which porphyroclasts of feldspar are set in a fine-grained aggregate of quartz. The porphyroclasts are from 1 mm to a few centimeters in diameter and display undulose extinction, deformation twinning, and abundant fractures. The nature of the porphyroclasts indicates that they are pretectonic and have persisted as relicts from primary igneous texture. Fractures that traverse the larger crystals are filled with granulated feldspar, and foliation defined by quartz aggregates wraps around the large feldspars. Quartz is abundant and its grain size, texture, and degree of strain are commonly a function of proximity to grain boundaries. Biotite, commonly partially altered to chlorite, and rare muscovite occur as (1) fine-grained aggregates that wrap around coarser feldspar crystals; (2) bent and twisted flakes where preserved from the effects of high strain; and (3) lenticular and ribbonlike aggregates of fine-grained, parallel flakes and prisms distributed among the well-oriented, strongly granulated recrystallized quartz that forms the matrix. Less abundant constituents such as sphene, apatite, zircon, and opaque minerals are commonly associated with the biotite. Of these, the opaque minerals in places show distinct granulation, whereas other minor components surrounded by more easily deformed mica seem to have escaped much of the deformation. The quartzo-feldspathic gneisses are generally characterized by low-dipping foliation and penetrative lineation which commonly plunges less than 20°. Impressive swarms of aplitic and pegmatitic veins locally cut the layered country rock, and although they are not transposed into concordance with compositional layering, they are consistently slightly foliated. Locally migmatitic gneisses record extensive flowage (Fig. 4), spectacular boudinage (Fig. 5), and folds and faults. Folded layers of lineated mylonitic rock are not uncommon. Foliation and lineation become less distinct as the top of the range is approached. The rock is cut by striated surfaces, but it is not deformed to any unusual degree.

Figure 4. Migmatitic gneisses in Sierra Mazatan.

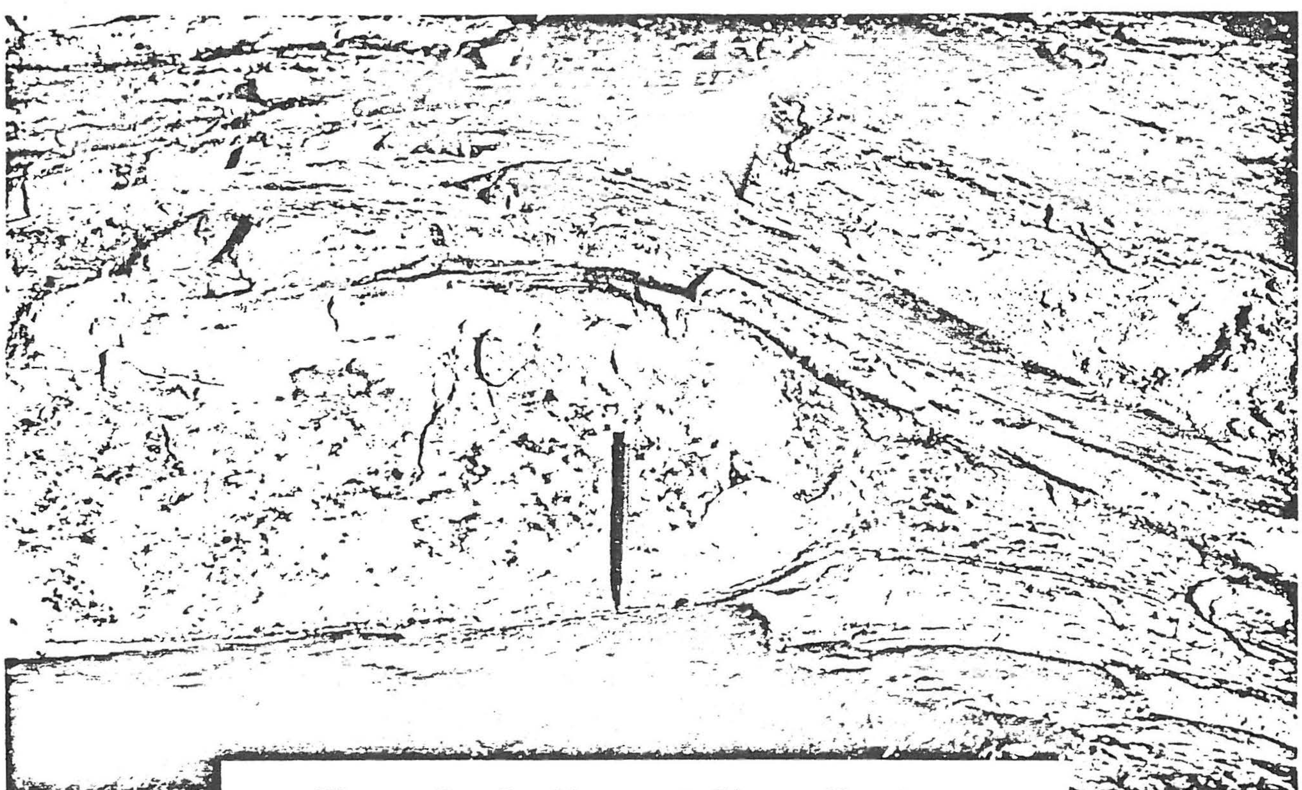

Figure 5. Boudinage at Sierra Mazatan.

Davis, ET AL. (6)

Décollement.

Carbonate beds of Paleozoic age rest upon a distinct, strongly lineated, platy surface of strongly mylonitic granodiorite gneiss and quartzo-feldspathic gneisses. The carbonate detachment rocks are locally folded. The intense microbrecciation and chlorite-epidote alteration so commonly seen in décollement zones have not been observed at Sierra Mazatan although detailed mapping of the décollement zone has not yet been undertaken.

MAGDALENA REGION

Introductory Comments.

In north-central Sonora, strongly deformed tectonites comprised of low- to medium-grade metasedimentary and metavolcanic rocks and associated silicic plutons underlie major north-trending sierras. These metamorphic tectonites are extensively developed along a north-south line in the region between Santa Ana and Imuris (Fig. 3). The tectonites extend from Sierra Madera on the east to at least as far west as the town of Altar. The zone of intensive deformation appears to be bounded by major NW-trending faults.

The town of Magdalena lies within this terrane of metamorphic core complex rocks. West of town, rocks on the eastern flank of Sierra Magdalena record the structural and metamorphic effects of Tertiary deformation.

Protolith.

Initial reconnaissance field mapping of a large area in north-central Sonora by Salas (1968[1970]) resulted in discovery of an extensive suite of greenschist-facies rocks composed of phyllite with subordinate marble, meta-conglomerate, quartzite, and metagraywacke. In places, mineral assemblages characteristic of the almandine-amphibolite facies occur. Salas (1968[1970]) concluded that the layered rocks were composed of two suites, both of which were interpreted to be Precambrian in age. His conclusion was based on the fact that metamorphism recorded in this region is comparable to that reported from areas of Precambrian crystalline rocks that crop out to the west, south of Caborca.

Radiometric data (Anderson and others, 1980) indicate that one of the strongly deformed plutons considered to be Precambrian is actually of latest Cretaceous age. To date, no paleontologic or radiometric studies that bear directly on the age of the layered metamorphic rocks have been conducted. However, lithologically similar beds on regional strike from these metamorphic units yield mid-Jurassic ages.

For rocks on the eastern flank of Sierra Magdalena, metamorphism has obscured most primary features. The rocks are quartzo-feldspathic gneisses and granitoid rocks. The gneisses are considered to be representative of the deformed silicic volcanic rocks of Jurassic age. Similar gneisses seen at other localities grade transitionally into Jurassic quartz porphyry. The strongly porphyritic granitic gneisses may have been derived from bona fide intrusive plutons, but they may have been derived as well from igneous material mobilized in place through anatectic processes.

Mylonitic Tectonite.

The tectonites derived from the Cretaceous plutonic rocks and the

Davis, et al. (7)

Jurassic volcanic and sedimentary rocks share a common Tertiary (?) deformational fabric. Folds inherited from younger, pre-Tertiary (?) deformation can be discerned in the metamorphic rocks, but these are overprinted by a consistently aligned foliation/lineation fabric.

POZO VERDE MOUNTAIN

Introductory Comments.

Pozo Verde Mountain lies just west of the border-town of Sasabe, Arizona (Fig. 3). It occupies the southernmost end of the Baboquivari Mountain complex, which includes in the broadest sense the Pozo Verde, Baboquivari, Quinlan, and Coyote Mountains. Within the Pozo Verde Mountain terrane there is a 100 km^2 area of lineated mylonitic tectonite, formed at the expense of granitic protolith. The only other exposure of mylonitic tectonite within the entire Baboquivari Mountain complex is at the northernmost end, where the lineated tectonites of the Coyote Mountains crop out. Arched foliation in mylonitic tectonite in the Pozo Verde Mountains plunges gently south; that in the Coyote Mountains dips northerly. In between, the 50-km long Baboquivari Mountains consist entirely of non-mylonitic country-rock protolith. It is possible that the mylonitic tectonite may have passed as a continuous tectonite sheet over the entire stretch of terrain now occupied by the Baboquivari Mountains. However, it seems more likely that the northerly and southerly zones of mylonitic tectonite were never regionally continuous. Rather, they represent separate zones of shearing within a common system.

Protolith.

Haxel and others(1980) have discussed the geology of the Baboquivari Mountains in the context of the geology of the southern part of the Papago Indian Reservation. The main rock in the Pozo Verde Mountains was named (informally) by them "granites of the Presumido type", where Presumido refers to Presumido Peak in the northern part of the Pozo Verde Mountains. These granites are muscovite-garnet-biotite granites which generally are fine- to medium-grained, hypidiomorphic in texture. Haxel and others (1980) distinguish an older leucocratic phase which contains little garnet, and a younger, white-weathering leucocratic phase which usually has garnet. Compositionally the rocks are leucocratic monzogranites (Haxel and others, 1980).

The Pan Tak granite in the Coyote Mountains, one of the granites of the Presumido type, was dated at 58 ± 3 m.y., based on Ur-Pb on zircon (Wright and Haxel, 1980). Haxel and others (1980) believe that the granite in the Pozo Verde Mountain is also early Tertiary in age. A K-Ar determination on muscovite from granites of the Pozo Verde Mountain revealed a 24 m.y. cooling age (W.A. Rehrig and S.J. Reynolds, pers. comm., 1978).

Granites of the Presumido type intrude a country rock framework of dominantly Mesozoic rocks (Heindl and Fair, 1965; Haxel and others, 1980), although Mesozoic rocks are not part of the Pozo Verde Mountain. The Mesozoic rocks include metavolcanic and metasedimentary rocks, dominantly rhyolites and interbedded quartzites; unmetamorphosed clastic sedimentary rocks and rhyodacitic to andesitic volcanics; and syenogranitic to granodioritic plutons (Haxel and others, 1980). The plutons range in age from 170 to 140 m.y. and intrude the inferred Triassic (?) to early Jurassic sequence of metamorphic, sedimentary, and volcanic rocks (Haxel and others, 1980). The Mesozoic country-rock framework makes up most of the Baboquivari Mountain complex. For example, the imposing Baboquivari Peak is undergirded by a 150 ± 10 m.y. granite (Haxel and others, 1980).

Mylonitic Tectonite.

Mylonitic tectonite in the Pozo Verde Mountains is wholly fashioned from granites of the Presumido type. Although the rocks are monotonously uniform compositionally, it is interesting to examine the upward increase in intensity of foliation and lineation. Granitic protolith with scarcely any distinguishable foliation and lineation is progressively transformed into penetratively deformed tectonite. The physiographic expression of this transformation is quite striking. Non-mylonitic outcrops of leucocratic rocks are weathered to rounded, exfoliated forms. In contrast, zones of mylonitic tectonite weather to subhorizontal layers and benches of moderately fissile rock.

Where most deformed, rocks of the Pozo Verde Mountain terrain are fine- to medium-grained quartz monzonitic augen gneisses. Aplite and quartz veins are ptygmatically folded into tight to isoclinal recumbent forms. In essence they are transposed into foliation-parallel alignment.

Mapping by Davis (1980b) has disclosed the general arch-like nature of mylonitic tectonite in the Pozo Verde Mountains. Foliation shows broad warping about the gently south-plunging axis (Fig. 6). Dips range in inclination from horizontal to approximately 35°. Foliations in the mylonitic tectonite do not conform to a simple arch, but one that is locally irregular in shape. Smooth curviplanar pinch and swell relationships are apparent both in map and outcrop relationships.

Mineral lineation is physically defined by alignment of unequidimensional chips of feldspar and elongate aggregates of plastically deformed quartz. More brittle counterparts of the lineation are slickenside striae. Lineation trends consistently north-northeast, with plunge values and directions congruent with local orientations of the mylonitic foliation in which the lineation is found. The trend of lineation in the Pozo Verde Mountains is transitional between the dominantly north-south trends which characterize the Papago Reservation domain to the north and west (Davis, 1980b) and the N60°E trends which typify both the Rincon-Catalina complex near Tucson and the complexes in northern Sonora.

The structural level of the upper part of the mylonitic tectonite is very close to the décollement zone, for rocks of décollement-zone affinity are exposed at westwardmost exposures along the base of the western flank of the Pozo Verde Mountains (Fig. 6). There, pegmatites and leucocratic granitic rocks are microbrecciated, shattered, and altered.

COYOTE MOUNTAINS

Introductory Comments.

The Coyote Mountains lie at the north end of the Baboquivari Mountain complex. The mountain is located south of State Highway 86 west of Three Points (Fig. 3). The northern flank of the Coyote Mountains has geologic properties which conform to characteristics of metamorphic core complexes (Davis, 1980b). Details of the structural geology and petrology of the mylonitic tectonite and décollement zone were investigated by A.F. Gardulski (1980). Unless otherwise noted, the descriptions and interpretations which follow are taken from her work.

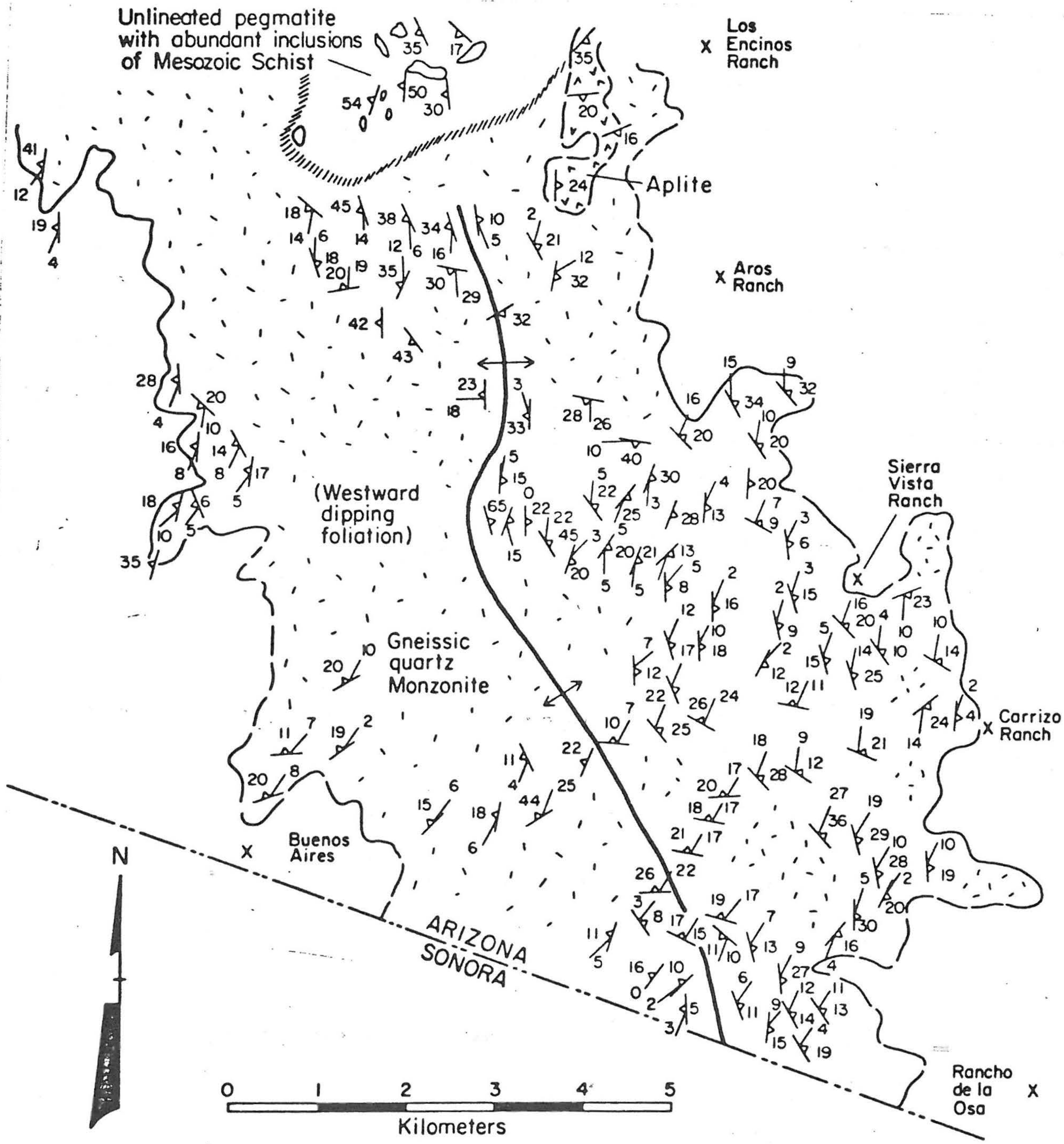

Figure 6. Structure map of Pozo Verde Mountains.

Davis, et al. (9)

The general geology of the area which Gardulski (1980) studied is shown in Figure 7. The map shows a number of units, ranging in age from lower Paleozoic to Tertiary. The rocks which constitute the main focus of this discussion are Pan Tak granite and Bolsa Quartzite. The Pan Tak granite is the granite of Presumido type for which Wright and Haxel (1980) interpreted a U-Pb zircon age of 58 m.y. Both the granite and the Bolsa Quartzite were converted into mylonitic tectonite. Additionally, rocks of the Cambrian Abrigo and Devonian Martin Formations locally occur within the zone of mylonitic tectonite. These rocks crop out as deformed calc-silicate and marble assemblages.

A décollement zone, the Ajo Road décollement, separates mylonitic tectonite from overlying non-tectonite, unmetamorphosed sedimentary and volcanic rocks (Fig. 7). Closest to the décollement are andesitic to latitic volcanic flows, breccias, and derivative sediments of the Roadside Formation (Gardulski, 1980). Heindl (1965) assigned a Cretaceous age to this lithologic assemblage. Overlying the rocks of the Roadside Formation, still within the upper-plate detachment, are red clastic sedimentary rocks of the Cretaceous (?) Sand Wells Formation (G. Haxel, personal communication, 1980).

Protolith.

The Pan Tak granite has four phases -- an older equigranular granite and three younger phases. These younger phases include equigranular, coarse-grained granite; xenomorphic, medium-grained granite; and swarms of pegmatites. The pegmatites comprise a large part of the northern Coyote Mountains. The phase of the Pan Tak granite which yielded the 58 m.y. age date is the younger equigranular coarse-grained granite. The pegmatites are even younger. According to Wright and Haxel (1980), the isotopic systematics of zircons in the Pan Tak granite demonstrate an anatectic origin for the intrusion. A Precambrian parent rock is indicated.

The older granitic phase of the Pan Tak granite does not crop out in the study area which Gardulski mapped (1980). The younger phases crop out but are nearly identical compositionally. They are so intimately mixed that they are generally impossible to map as discrete units at a reasonable scale. The pegmatite and granite are commonly gradational, and there are few examples of clear-cut intrusive relationships between the phases.

Mineralogically, the pegmatite and granite consist of quartz, microcline, albitic plagioclase, biotite, and locally-abundant muscovite and garnet. In some sills, a delicate zoning is developed as compositional banding.

The Bolsa Quartzite is mineralogically homogeneous, quartzose rock containing only very minor feldspar, biotite, muscovite, and heavy minerals. It is white, grey, black, or more rarely, maroon in outcrop. The total exposed thickness of the quartzite is estimated at 30 m. Assigning this quartzite to the Bolsa Quartzite is an interpretation which we think is compatible with general outcrop appearance, mineralogy, and structural position underneath recognizable Abrigo and Martin Formations. No conglomeratic layers or lenses occur in the quartzite.

The overall geometry of rock relations in the northern Coyote Mountains has been greatly influenced by the intrusion of the Pan Tak granite suite. Intrusion apparently caused significant dilation of the Paleozoic section. Sills of the Pan Tak granitic system engulfed the sediments. Quartzite lenses are sandwiched between undulatory surfaces of the pegmatite sills. Sharp contacts are maintained in the Abrigo and Martin Formations, with little encroachment of pegmatite into the interiors of these sedimentary rock pods.

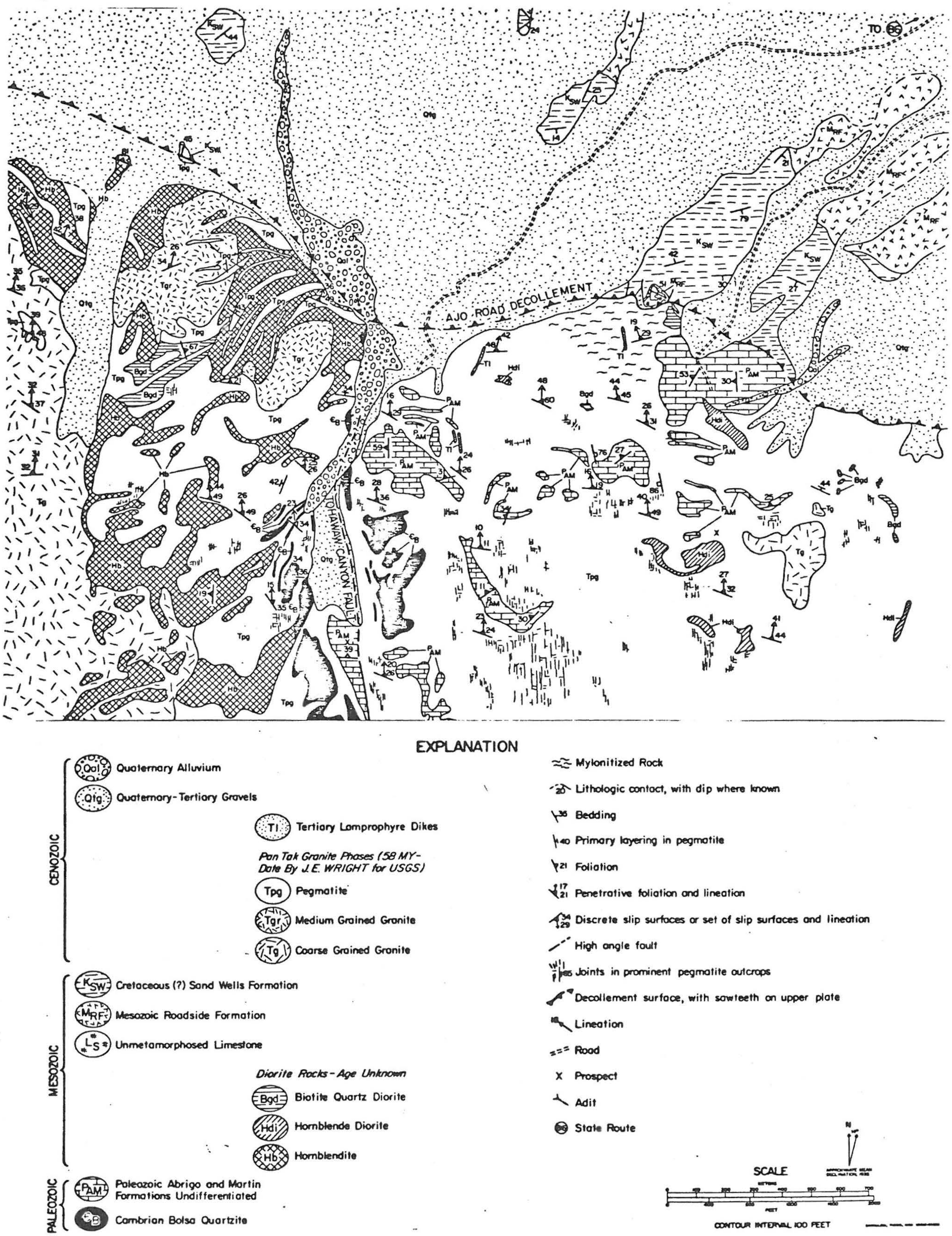

Figure 7. Geologic map of Tohawaw Canyon area, Coyote Mountains (Gardulski, 1980).

Mylonitic Tectonite.

Mylonitic tectonite is exceptionally well exposed in a canyon that Gardulski (1980) informally named Tohawaw Canyon. Tectonite quartzite derived from the Bolsa Quartzite is interlayered with pegmatite sills, the outer margins of which are converted to tectonite (Fig. 8). There are unusually good exposures of the quartzite plated or smeared onto tops of underlying sills (Fig. 9). Abrupt changes in dip of the quartzite, like that shown in Figure 9, reflect radical pinch-and-swell morphology of underlying pegmatite. Some of the sills are like giant augen (Fig. 8).

The exposure at the Tohawaw Canyon study site is mainly a broad expanse of tectonitic pegmatite. Quartzite layers are exposed in the walls, but the undulating floor of the canyon is mostly devoid of quartzite. Much of the floor seems to be a structural surface on the top of a series of pegmatite sills from which the once overlying quartzite has been removed. In effect, the canyon floor is a plan-view rendition of pinch-and-swell morphology of tectonitic pegmatite.

A map of the relationships at the study site is shown in Figure 10. This map was prepared by students in the graduate-level detailed structural analysis class at The University of Arizona (see caption, Fig. 10). The topography and geology of the site was mapped with plane table and alidade at a scale of 1" = 20 feet. The topographic map is an approximation of a structural contour map on the top of one family of pegmatites. Geologic contacts are the same as those shown in Gardulski's map, but in addition the large-scale map shows folded primary compositional layering in pegmatite, mylonitic shear zones which cut the pegmatite, several normal faults, and major through-going joints.

Bolsa Quartzite.

The quartzite is lineated tectonite wherever it crops out. A very strong mineral lineation and foliation penetrate the rock (Fig. 11), and these elements are defined by the orientation of streaked plates of muscovite and biotite as well as by flattened, elongate quartz grains (Fig. 12). Foliation strikes west-northwest and dips gently north-northeast; lineation plunges gently north (Fig. 13).

In thin section, the grains are seen to be undulose and partially to completely recrystallized; flattening has been extreme (Fig. 12). Elongation of quartz parallel to lineation is quite pronounced. Both original and recrystallized grains are deformed. Original grain shape is not preserved. Within individual lenses of quartzite, a great variation in strain is evident. The average long dimension of grains is at least an order of magnitude smaller in the tail of a lens than in the center (e.g., .02 mm vs. .2mm).

Small pinch-and-swell features with wavelengths of 1-10 cm occur locally in the quartzite, and the foliation and lineation have been warped by these small structures (Fig. 14). Where quartzite layers are thin (0.5 cm) and alternating with concordant veins of pegmatite, this pinch-and-swell feature more closely resembles mullion structures whose presence is due to the contrasting ductilities of the two media. They occur where underlying pegmatite has been stepped-down by normal faulting, thus requiring the quartzite to undergo major local stretching.

Small normal faults occur locally in the quartzite, and slip ranges from 1 mm to 20 cm in the direction of plunge of the lineation (Fig. 15).

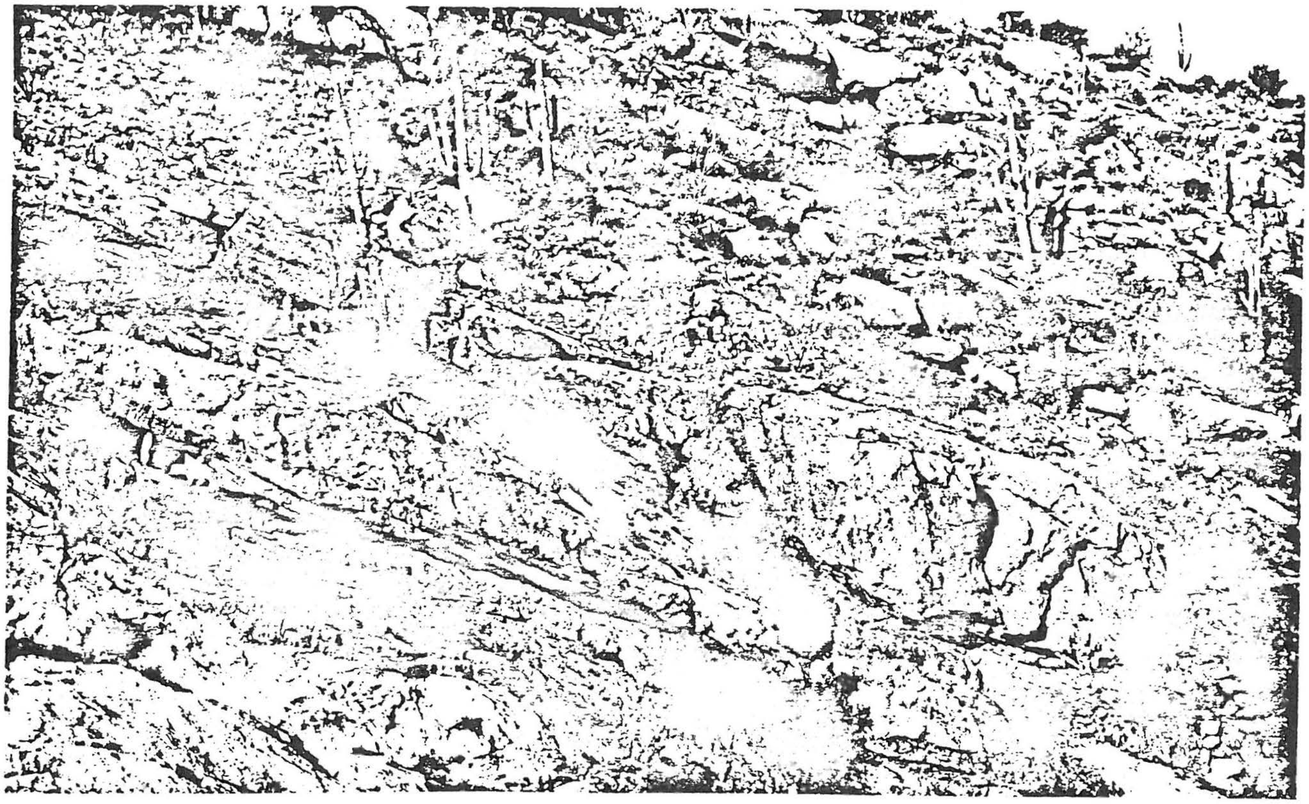

Figure 8. Interlayered quartzite and pegmatite sills, Tohawaw Canyon.

Figure 9. Quartzite plated on top of pegmatite sill, Tohawaw Canyon.

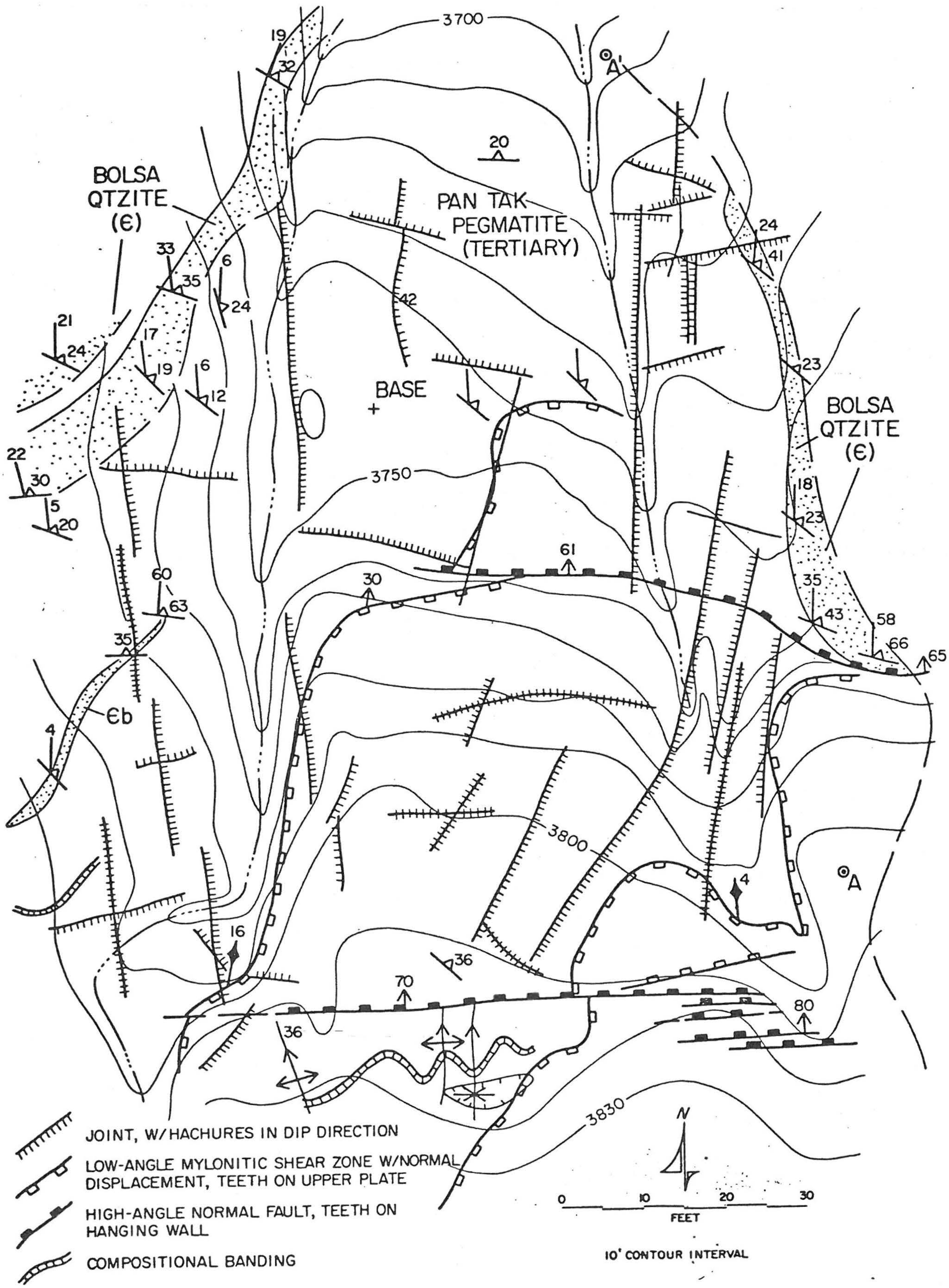

Figure 10. Large-scale structure map of part of Tohawaw Canyon.

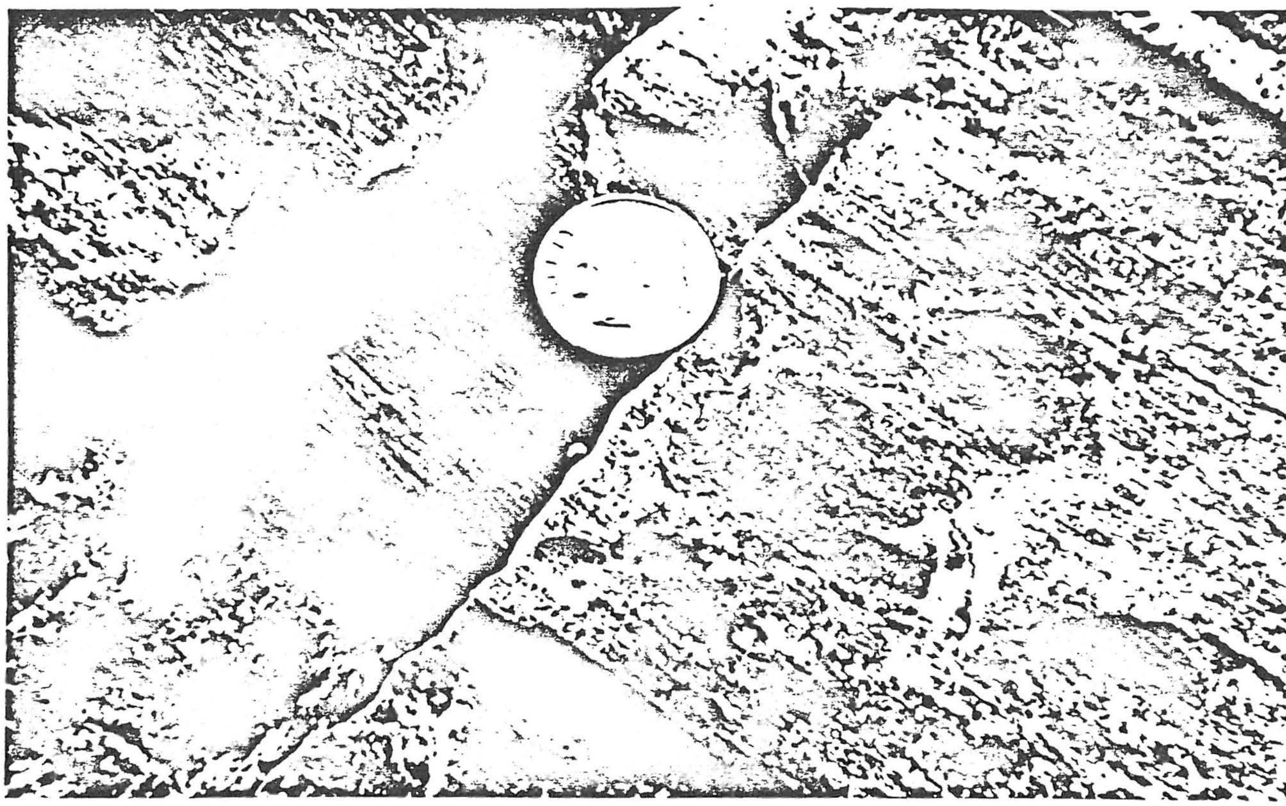

Figure 11. Lineated tectonite quartzite, Tohawaw Canyon.

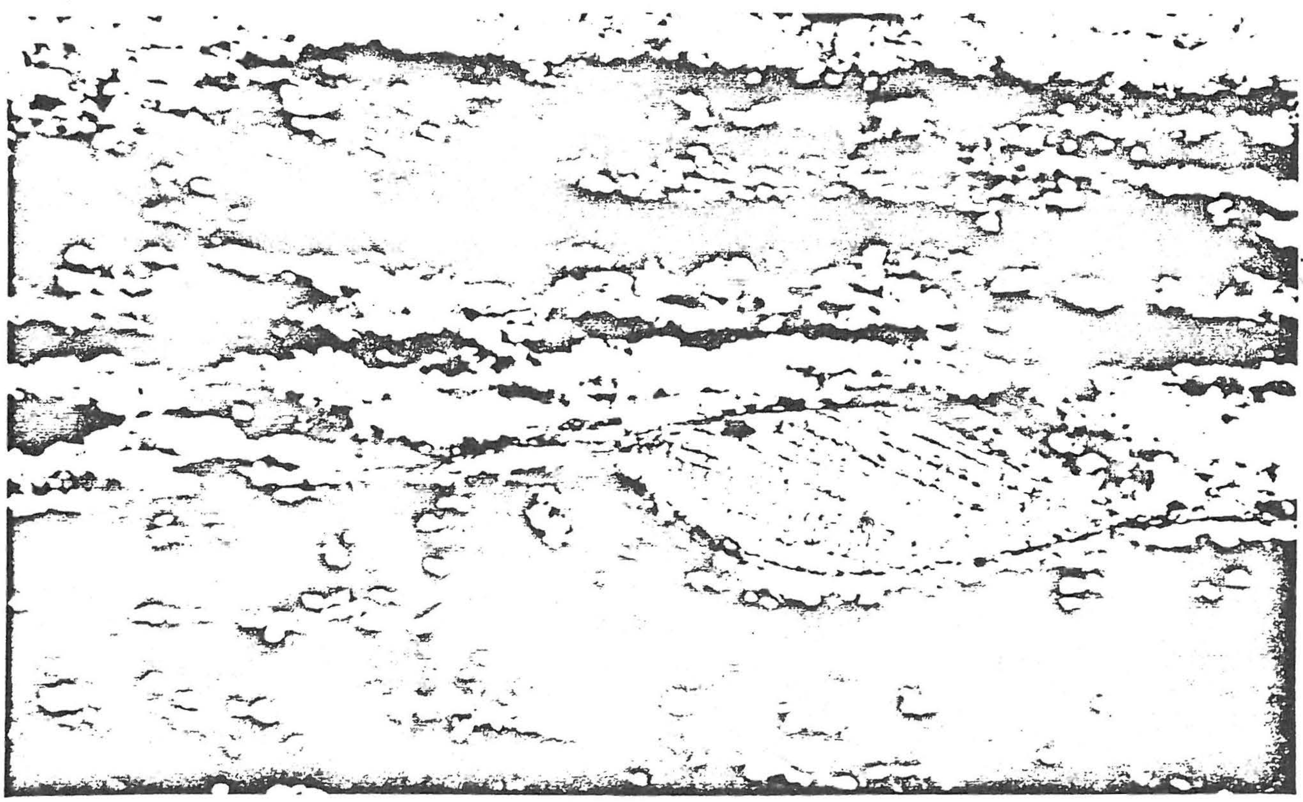

Figure 12. Photomicrograph of tectonite quartzite. 70X.

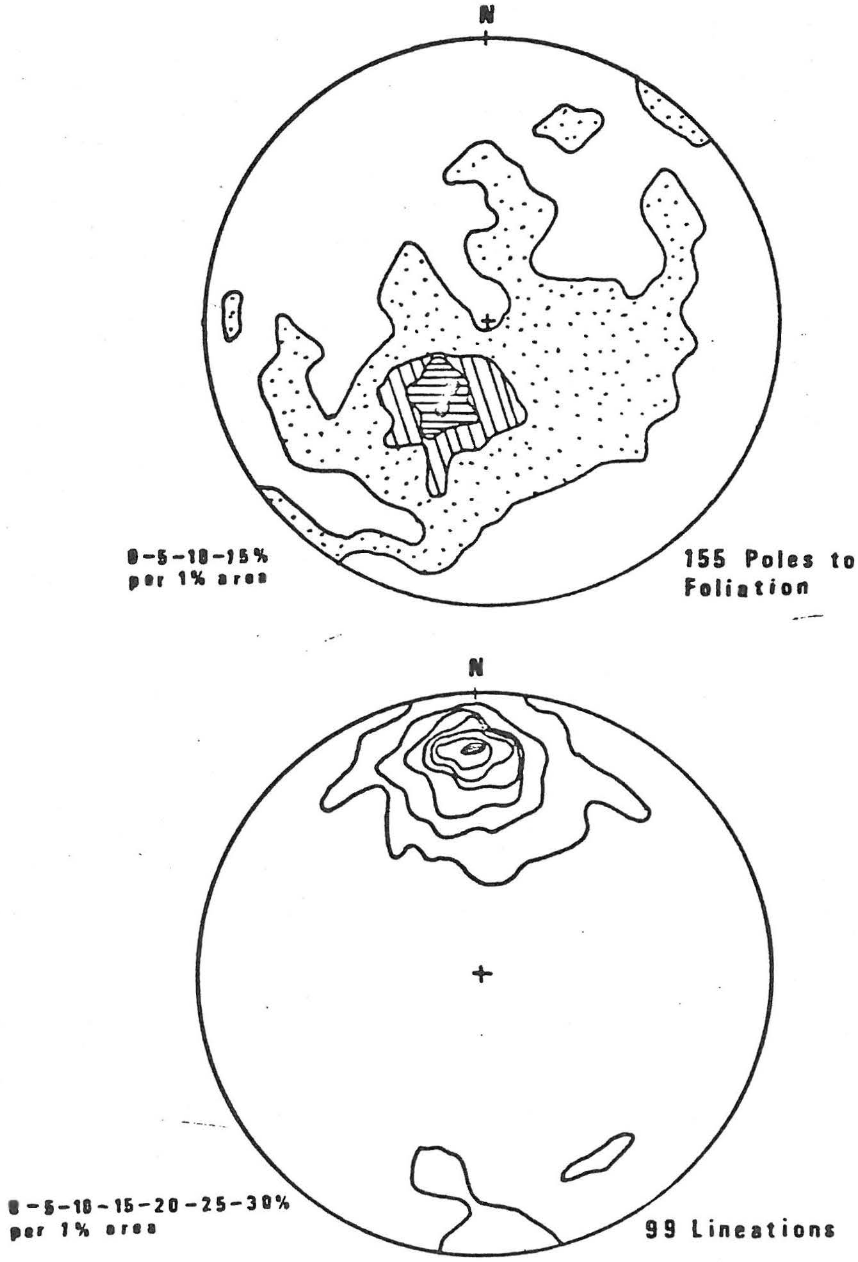

Figure 13. Lower-hemisphere equal-area projection of foliation and lineation, Tohawaw Canyon.

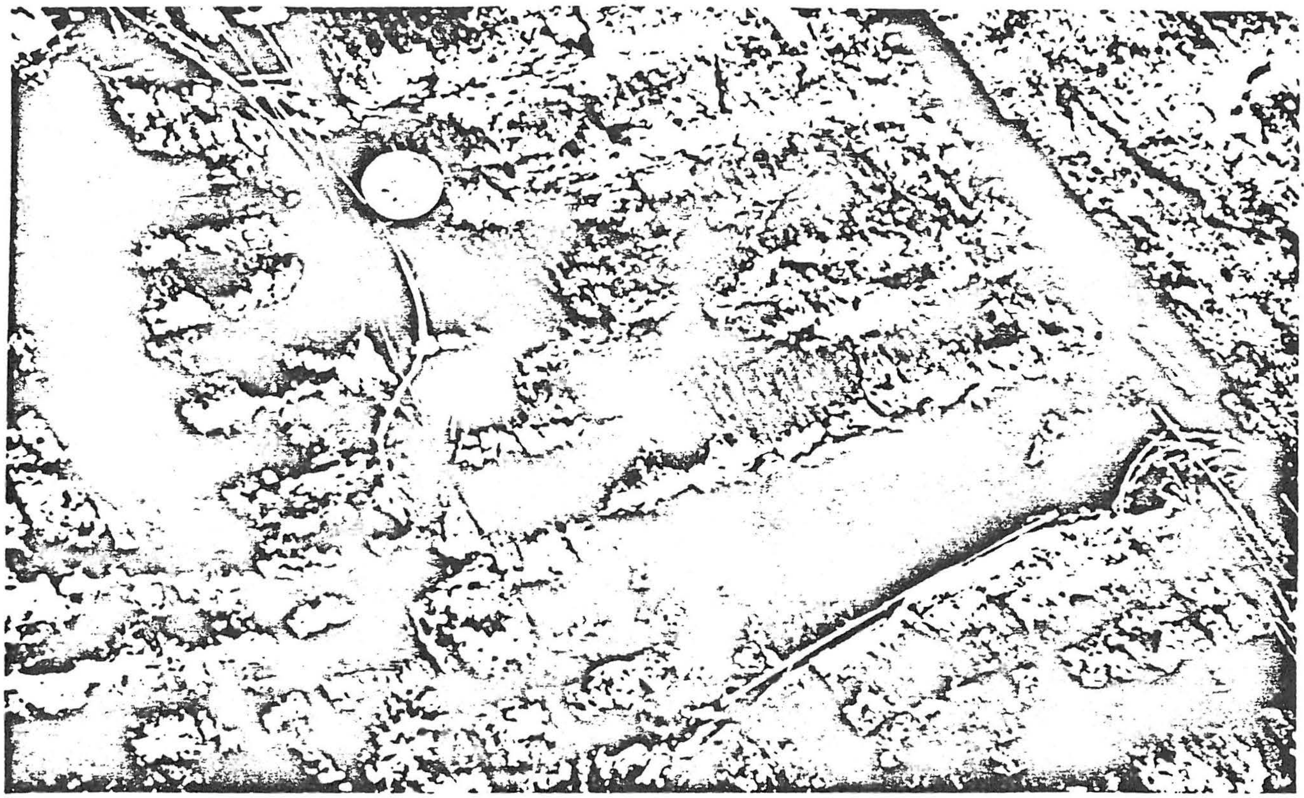

Figure 14. Small crenulations in lineated tectonite quartzite.

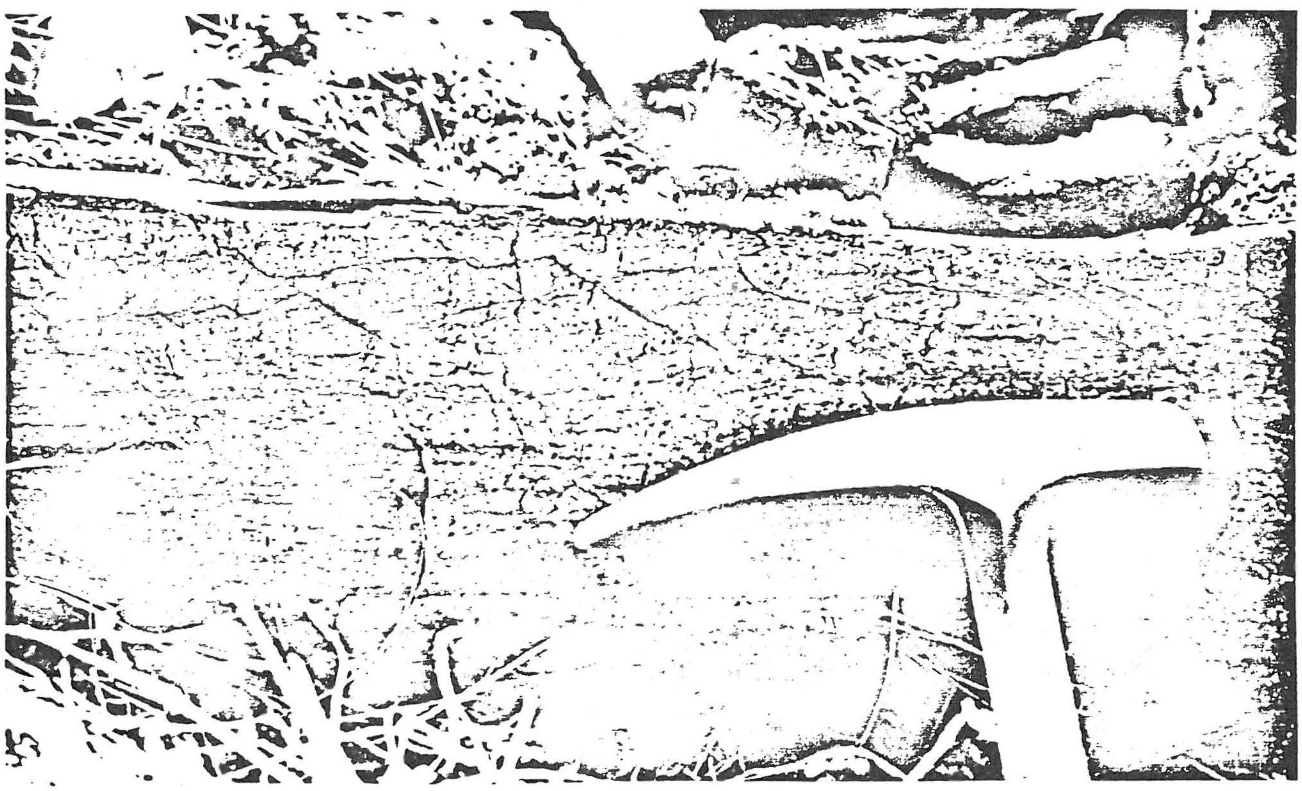

Figure 15. Normal faults in tectonite quartzite.

Faults with very small displacements occur within and parallel to the troughs of the pinch-and-swell features. Normal faults with larger amounts of slip are rare, but they affect both quartzite and pegmatite.

Thin section examination reveals the ductile nature of the small normal faults. In the necked regions of the pinch-and-swell structure, north-dipping steep surfaces of shear have developed, but there is no loss of cohesion. Shear zones cut through the necked regions of quartzite, stepping down, across, and through the layers, but without concurrent development of jointing. Large-scale geologic map relationships (Fig. 10) show comparable but larger shear zones in the tectonitic pegmatite. The zones display abrupt shifts in inclination, from subhorizontal to vertical. The mylonitic fabric of the shear zones shows micro-pinch-and-swell and penetrative micro-fractures at a high angle to foliation.

The light and dark banding visible in quartzite outcrops may represent relict bedding. Foliation is generally subparallel to this layering, but the angle between the two may be as great as 30°. There is evidence that bedding may have been isoclinally folded and transposed through passive flow. The strong overprint of the foliation and lineation in this rock has obscured these structures, and the homogeneous nature of the quartzite makes relict bedding hard to discern.

Structural petrological analysis of the quartzite by Gardulski (1980) revealed that the quartzite samples exhibit a strong maximum or split maximum that is located at high angle (60 to 90 degrees) to the lineation and within or very near the plane of the foliation (Fig. 16). The symmetry of the c-axis plots is nearly orthorhomoic for several of the specimens, but strictly speaking they are all monoclinic to triclinic. An important feature of all the orientation diagrams for quartzite is the pole-free areas perpendicular to foliation and parallel to lineation. Assuming that the normal to foliation represents an axis of shortening and that the lineation represents an axis of lengthening, the c-axes of quartz cluster about the intermediate strain axis. Such distributions have been produced experimentally as discussed by Wilson (1975), and they arise when prismatic slip predominates over slip on the basal plane of quartz. Bouchez (1977) noted a c-axis maximum at the intermediate axis where prismatic slip in a basal direction was operative at relatively high temperatures. Tullis and others (1973) obtained similar results at high temperatures with slow strain rates and high strains. It is apparent that basal slip was essentially not operative by the time the quartz fabric was "frozen" in the rocks.

Pan Tak Granite.

The coarse-grained young phase of the Pan Tak granite has developed a strong foliation and lineation which may be considered penetrative within the constraints of its grain size and mineralogic heterogeneity. The quartz grains show features typical of plastic deformation--undulatory extinction, grain elongation, and recrystallization. A great deal of the strain in the granite is accommodated in the quartz component of the lithology, while the feldspar fraction has been merely fractured and sheared.

This style of deformation is also characteristic of the pegmatite phase, but it is not as penetrative. Discrete slip surfaces exist in the pegmatite, and these are well-developed only at the margins of the pegmatite bodies. They become very closely spaced as contacts with the quartzite are approached.

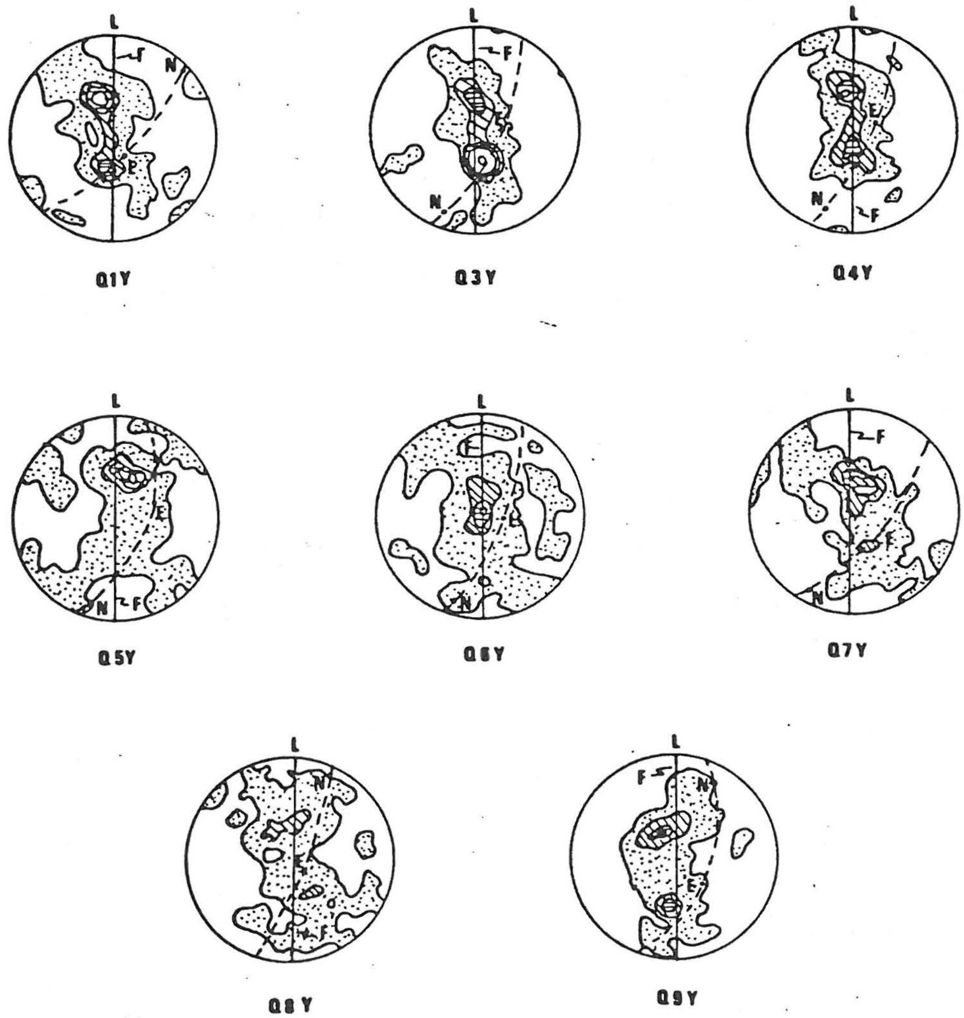

Figure 16. Lower hemisphere pole-density diagrams of c-axis orientations of quartz in tectonite quartzite, Tohawaw Canyon.

Where sedimentary rock packages are insignificant within large volumes of pegmatite, as in the southeastern part of the map area, fairly widely spaced slip surfaces are developed through zones as thick as 100 m. The zones of slip surfaces are separated by undeformed rock, and such alternating bands of deformed and undeformed rock may be traced east-west for hundreds of meters.

Gardulski (1980) examined the structural petrology of the pegmatite and found that there is a much greater scatter of low densities in c-axis distribution as contrasted with diagrams for the quartzite (Fig. 17). She attributed this to the high content of feldspar. As a brittle mineral under the conditions of deformation, the feldspar would protect certain zones of quartz from excessive strain. Attitudes of maxima of c-axes of quartz in the pegmatite are concordant with those for quartzite specimens.

Striated surfaces, spaced centimeters apart, pervade many parts of the tectonitic pegmatite exposed in the study site. The surfaces display quite a range of orientations, but mostly they strike east-west and dip at gentle to moderate angles, either south or north (Fig. 18a). The strain significance of these surfaces is an accommodation to extensional stretching displacements and normal-slip. Striae on these surfaces show a strong preferred orientation, defining a great circle which trends north (Fig. 18b). The striae lie within the place defined by the axes of greatest and least elongational. Some surfaces are spoon-shaped, with dip-slip striae in the depression and strike-slip striae on the walls. S.L. Beard prepared a structure profile (Fig. 19) parallel to the lineation direction along the pegmatite/quartzite contact (see A-A^1, Fig. 10). A main purpose was to evaluate the distribution and dip-directions of striated surface as a function of the pinch-and-swell morphology of tectonitic pegmatite. The carefully constructed structural profile shows the stretched nature of the upper surface of the pegmatite(s). Joints in the pegmatite are well developed and systematically aligned. E.Y. Anthony, D.A. Currier, and K.F. Inmann analyzed the jointing and documented the presence of orthogonal sets, one striking N-S, the other striking E-W (Fig. 20). Fracture density in the pegmatite averages about 0.1 cm^{-1} (data from Anthony, Currier, and Inmann).

Décollement.

The mylonitic terrane of the Coyote Mountains is tectonically juxtaposed against the Roadside Formation and Sand Wells red beds by the Ajo Road décollement (Fig. 7). Slickensides plunge steeply at 40-60° to the north on the fault surface, and normal movement is indicated from the stratigraphic and lithologic relations. The décollement strikes west-northwest and dips north at an average of 45° (Davis, 1980b). In detail it is curviplanar with a sinuous trace along the mountain-pediment interface.

The hanging wall rocks consist of the Roadside Formation overlain by the Sand Wells Formation. These rocks are everywhere moderately fractured, but in the vicinity of the décollement they are shattered and assume a pasty, gougy appearance. At one exceptionally fine exposure, a sliver of intensely sheared Roadside Formation crops out along the fault surface as the lowest member of the upper plate. It is overlain by similarly deformed red beds of the Sand Wells Formation. At all other exposures of the fault in the map area, the Roadside Formation is absent and the red beds occupy the lowest structural position in the hanging wall.

The footwall rocks by the Ajo Road décollement consist primarily of

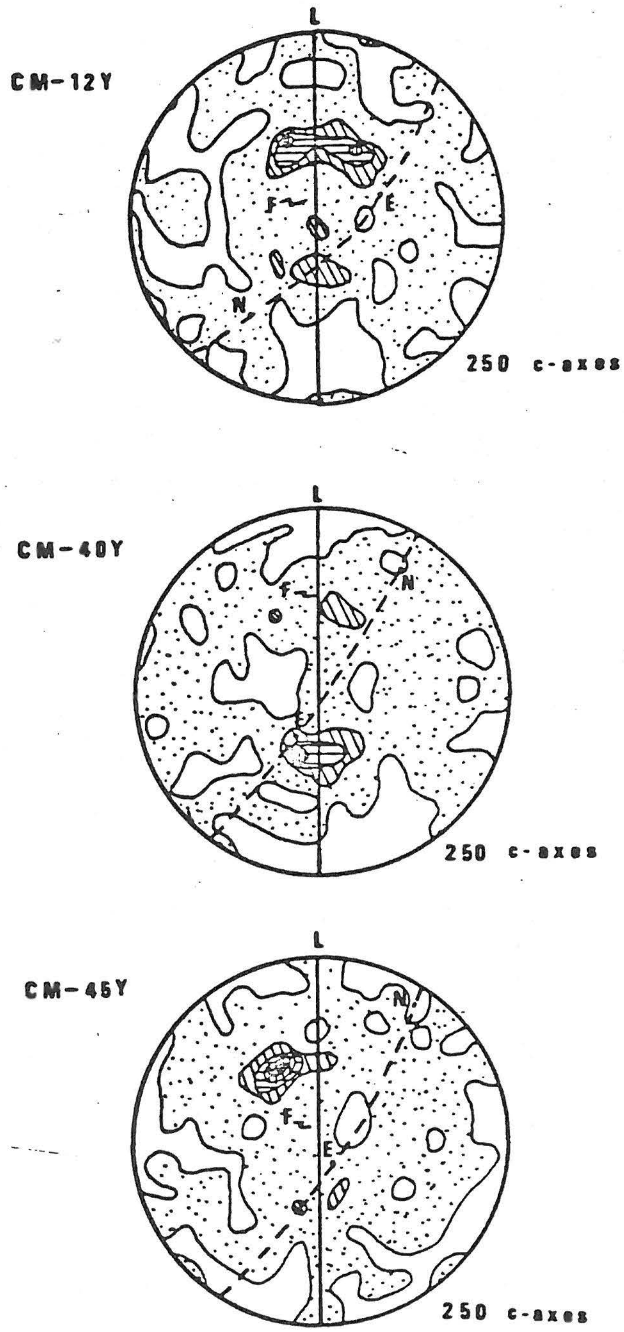

Figure 17. Lower hemisphere pole-density diagrams of c-axis orientations of quartz in pegmatite, Tohawaw Canyon.

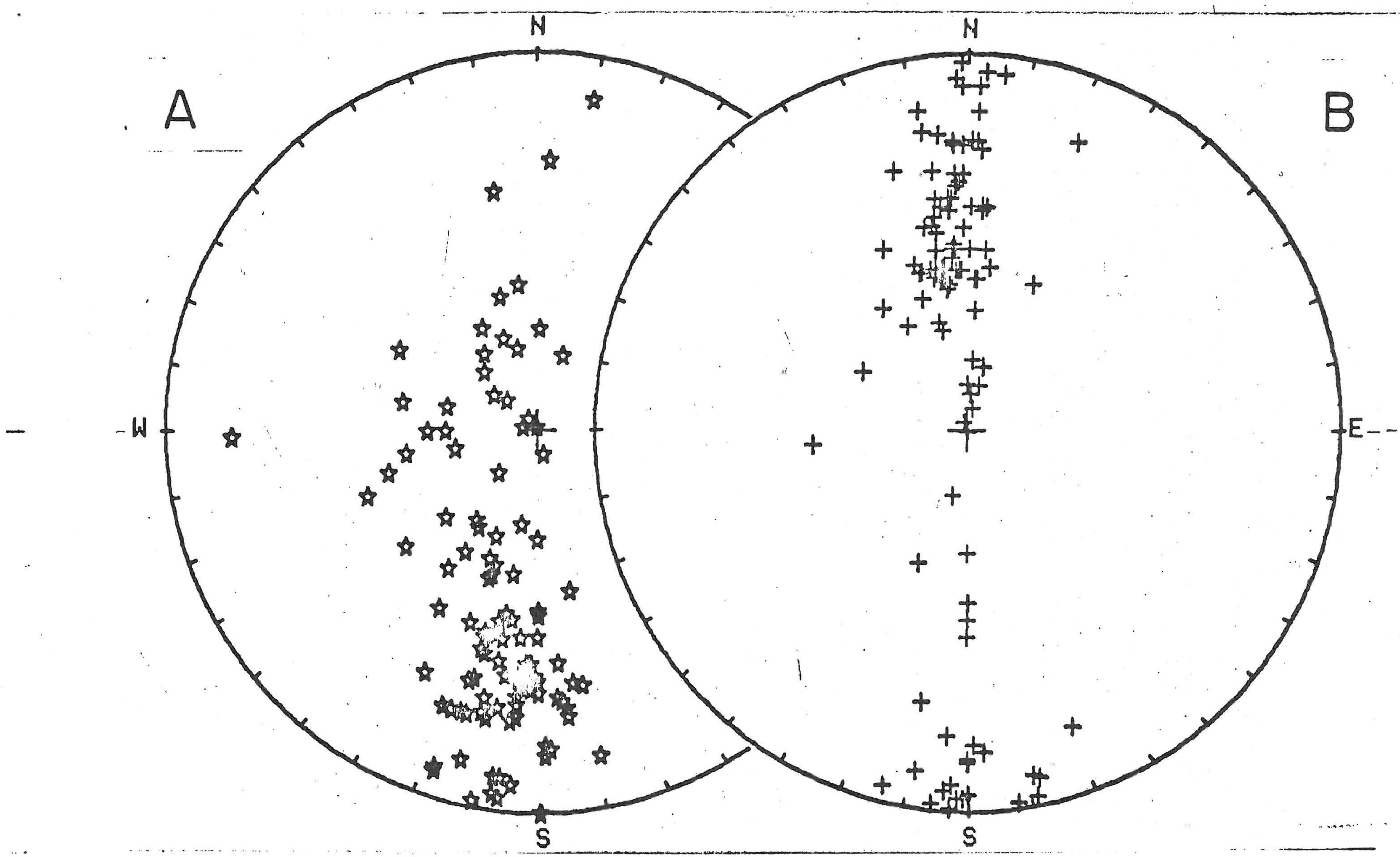

Figure 18. Lower hemisphere equal-area projections of (A) striations and (B) striated surfaces.

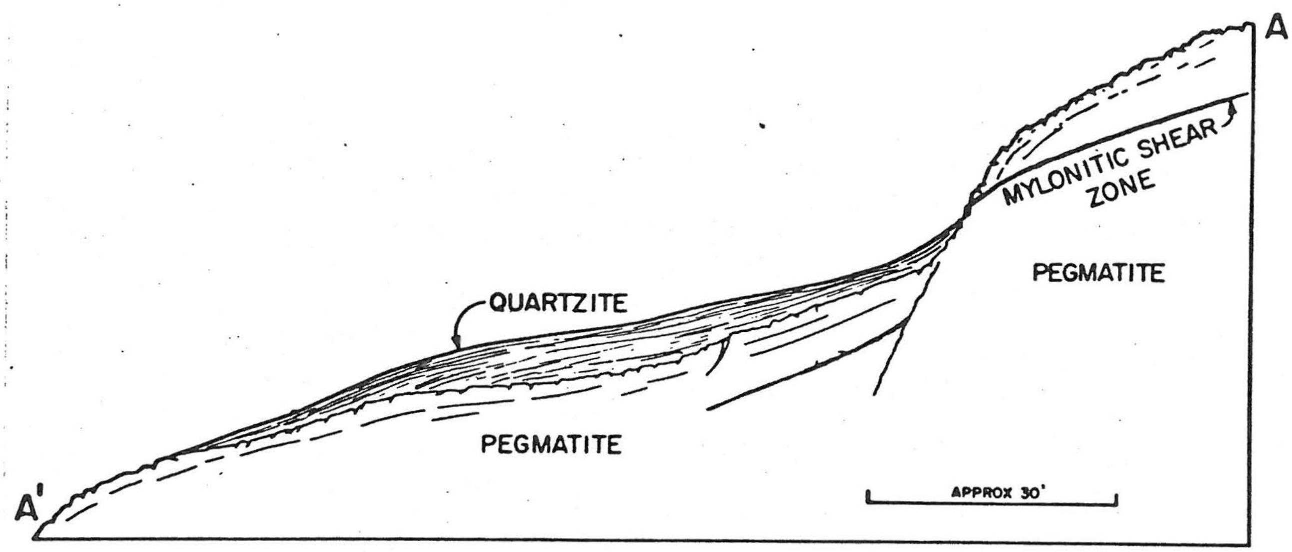

Figure 19. Structure profile showing traces of striated surfaces at outer margin of pegmatite sill(s). Tectonite quartzite is plated on the top of the pegmatite.

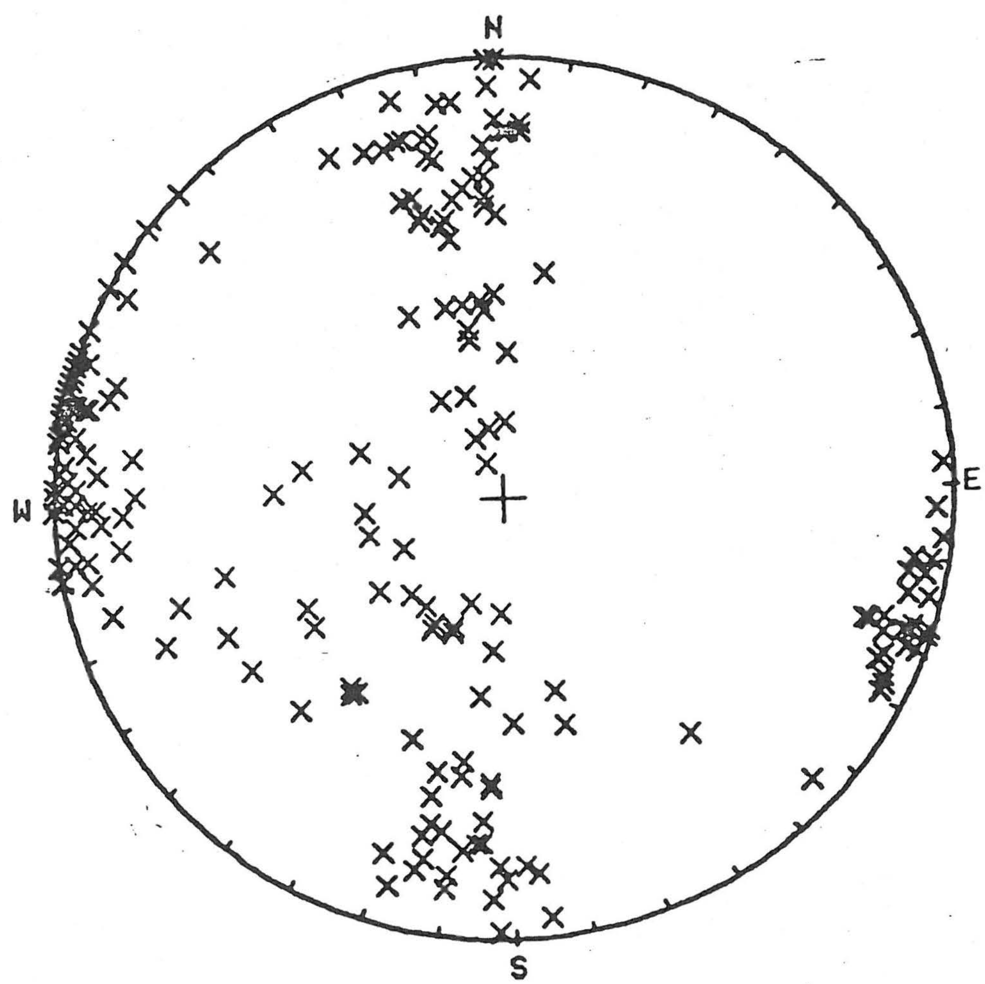

Figure 20. Lower hemisphere equal-area projection of poles to joints, Tohawaw Canyon.

Pan Tak granite and pegmatite, and lamprophyre dikes. Microbrecciation and a brittle, mylonitic foliation marks the décollement zone, and it overprints the earlier, more ductile, mylonitic foliation in the Pak Tak. The décollement-related imprint is characterized by intense brecciation of the footwall rocks, and faint foliation associated with this faulting is measurable in outcrop. The outcrop expression of the footwall near the décollement is that of a brown-to-green, chloritic, tabular mass of rock whose protolith may be determined only from field relations. The mineralogy in hand sample is indeterminable, and alteration and brecciation are so extreme that one of the most notable features of the rock is its fissile but shattered character. The intensity of microbrecciation in these rocks decreases rapidly down and away from the décollement, although a clear cataclastic overprint is visible through at least 100 m of rock.

In thin section, the rocks of the décollement zone include true mylonites, and subangular to subrounded fragments floating in comminuted groundmass attest to intense movement on the fault. Brittle mylonitization and microbrecciation can be seen to overprint previously deformed granite and pegmatite. Locally, quartz veins lace the mylonite, and the quartz is unstrained and unrecrystallized.

TANQUE VERDE MOUNTAINS

Introductory Comments.

The geology of the Tanque Verde Mountains has been mapped and discussed by Drewes (1975, 1978) (Fig. 21). Rocks in the mountain comprise an exceptionally good exposure of metamorphic core complex terrane (Davis, 1980b). The site of interest is located on the northwest flank of the mountain within the general area of the Loop Drive of Saguaro National Monument (East) (Fig. 22). Mylonitic tectonite, décollement zone, and detachment rocks are all well exposed. Analysis of the internal fabric of these bears importantly on kinematic interpretations of the structural evolution of metamorphic core complexes. Especially instructive is the physical and geometric nature of rocks within the décollement zone.

Protolith.

Mylonitic tectonite on the northwest flank of the Tanque Verde Mountains has been derived from two different plutonic rocks (Fig. 21). A coarse-grained mylonitic gneiss, mapped by Drewes (1978) as Precambrian Y Continental Granodiorite, is cataclastically deformed Precambrian quartz monzonite. The mylonitic gneiss is identical texturally and compositionally to a gneissic phase in the foothills of the nearby Santa Catalina Mountains, a rock now recognized as overprinted 1.4 b.y. "Oracle Granite". Creasy and Theodore (1975) originally recognized that the coarse-grained quartz monzonitic gneiss was derived from the Oracle Granite. Shakel and others (1977) extracted zircons from the gneiss and, using Ur-Pb isotopic analysis, found them to yield an age date of 1440 ± 10 m.y. The progressive transformation of the unfoliated Precambrian porphyritic quartz monzonite to coarse-grained mylonitic tectonite can be observed at a number of locations in southern Arizona (Davis and others, 1975; Banks and others, 1977).

The second variety of mylonitic gneiss in the Tanque Verde Mountains is fine- to medium-grained two-mica garnet-bearing quartz monzonitic augen

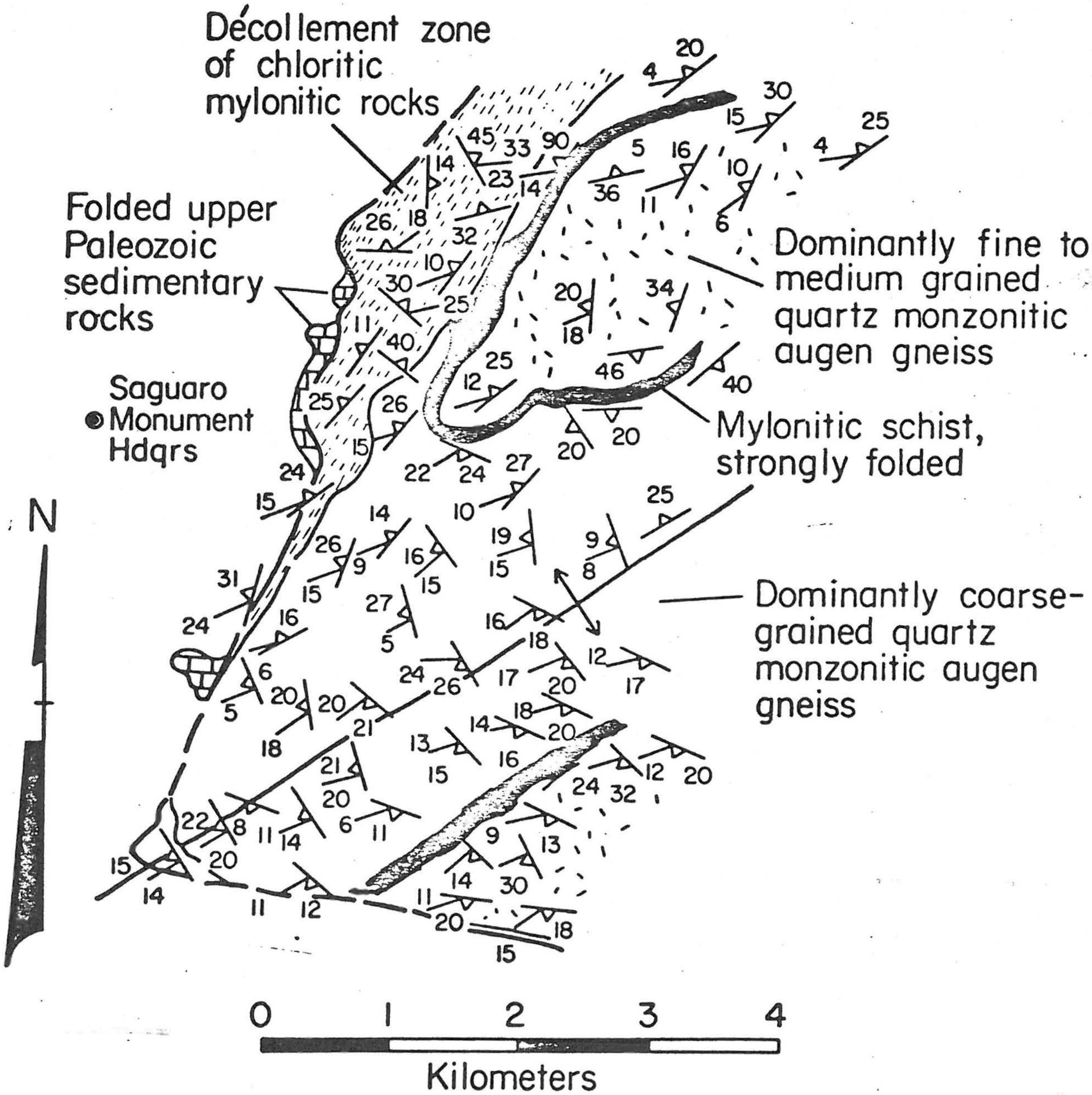

Figure 21. Geologic map of Tanque Verde Mountain area.

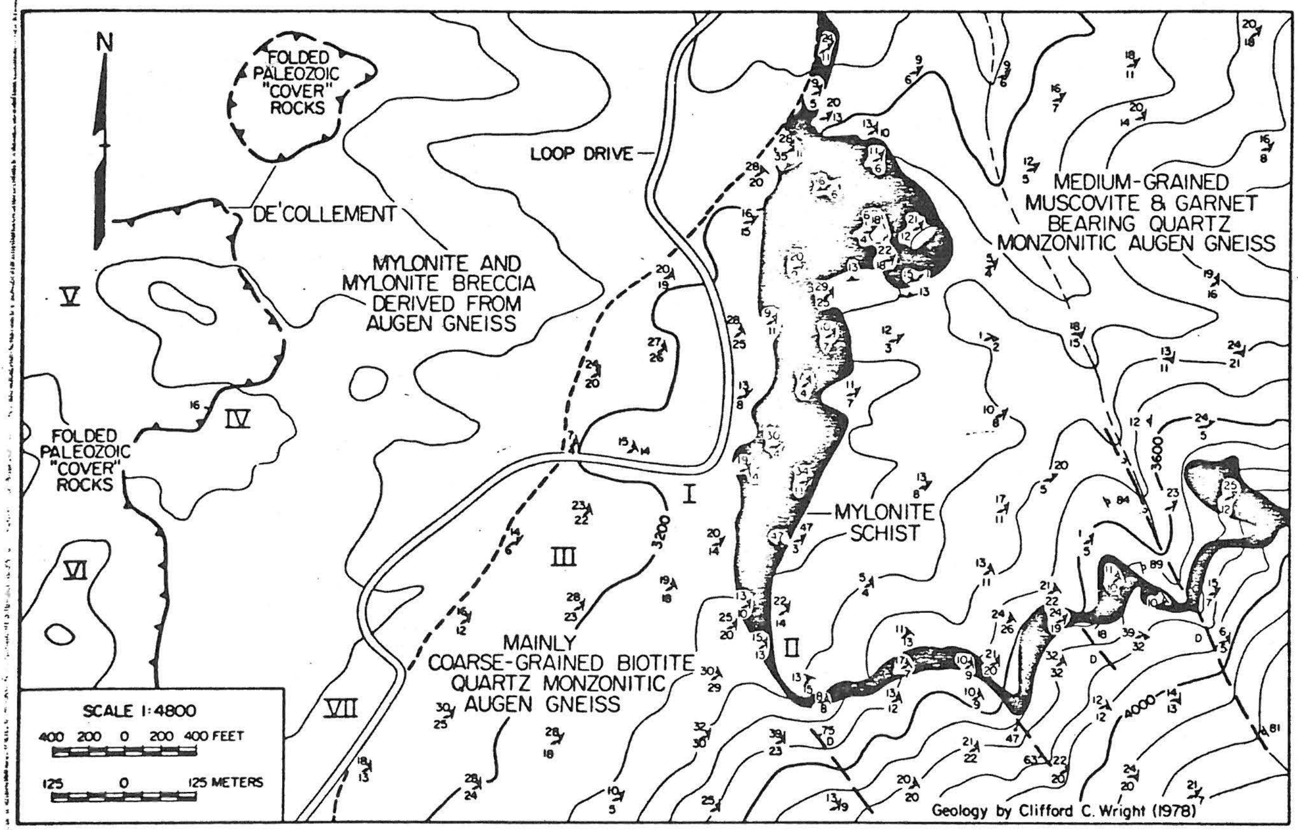

Figure 22. Geologic map of Tanque Verde Mountain area, prepared by C. Wright, senior thesis, Carleton College.

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Décollement Zone.

At Saguaro National Monument, in the vicinity of the eastern stretch of the Loop Drive, the contact between mylonitic tectonite and the décollement zone is abrupt. The coarse-grained mylonitic gneiss is overlain in fault contact by microbrecciated and rotated mylonitic gneiss (Fig. 25). The position of the fault boundary is distinguished by (1) abrupt change in orientation of foliation and lineation, and (2) alteration, shattering, and microbrecciation of the mylonitic gneiss. Normal coarse-grained mylonitic augen gneiss below the décollement zone consists of large (up to 4 cm) feldspar porphyroclasts of rounded to elliptical shapes in a matrix of mainly quartz and feldspar. Microscopically, the feldspar porphyroclasts show pull-apart structures with infilling by quartz and chlorite. Quartz occurs in several modes: feldspar-free zones of recrystallization in which quartz is undulatory and very fine grained; zones of flattened, elongate, extremely undulose quartz (L:T = 20:1) adjacent to feldspar porphyryoclasts; and zones of polygonal quartz in "pressure shadows" next to feldspar augen. Fracturing is minor and alteration of feldspar is insignificant in the nonmicrobrecciated tectonite gneiss.

By way of contrast, the microbrecciated tectonite gneisses of the décollement zone are marked by extreme facturing, brecciation, microfaulting, and alteration. The foliation visible in normal gneiss is nearly obliterated by anastomosing bands of microbreccia which cut through the rock. Microfaulting has caused rotation and tilting of micro-fault blocks of up to 60° or 70°. Some parts of this rock are so mylonitized that only a dense aphanitic mass remains. Veinlets of chlorite, epidote, biotite, and opaque minerals are abundant and lace the rock in apparently all orientations. Grain size is greatly reduced.

Different grades of microbrecciated mylonitic tectonite within the décollement zone have been mapped by Davis, and their distribution is shown in Figure 25. The intensities of microbrecciation are described as weak (w), moderate (m), and strong (s). Gardulski studied the petrography of representative examples of each of these field-based intensity classes. Where microbrecciation is weak, feldspar porphyroclasts remain very large (1 to 2 cm) but they are quite fractured. The porphyroclasts are rounded and not uncommonly elliptical. Some show significant pull-apart and chlorite veining. The infilling chlorite displays strongly anomalous interference colors, unlike the finer-grained alteration chlorite found in the matrix. Quartz grains above and below the feldspar porphyroclasts are markedly elongate (10 to 1 ratios) and/or very fine-grained (.001 mm). Extinction in the quartz is much more undulose than for quartz in feldspar-free areas. Away from the feldspar porphyroclasts, the quartz tends to be equidimensional and approximately .01 mm in grain size. Subgrain boundaries are abundant and original grain boundaries are usually vague because of recrystallization. Foliation is deflected around the large feldspar porphyroclasts. Quartz grains in pressure shadows of the feldspar porphyroclasts are relatively coarse. Thin laminae of micas and opaque minerals display anasomosing patterns around feldspar porphyroclasts and lenses of quartz grains.

The moderately microbrecciated rocks are marked by a number of petrographic changes which serve to distinguish them from normal or weakly microbrecciated tectonite gneiss. Quartz is strongly elongate (approximately 10:1) except where shielded by porphyroclasts of feldspar. Grain size of the quartz averages .01 mm. The feldspar porphyroclasts display selvages of recrystallized feldspar. Such feldspar is very fine-grained, approximately .0005 mm. The regions of feldspar recrystallization lie above and below the porphyroclasts, in pull-apart areas, and sometimes in pressure-shadow zones. There is a near total absence of biotite, opaque minerals, epidote, and apatite, and thus the foliation is almost completely defined by elongate quartz grains. Muscovite is coarse grained (.2 to .3 mm) and is recrystallized. The relative abundance of muscovite and the decline in amount of plagiocase feldspar implies that alteration and neomineralization may have been important processes at this stage of deformation.

Strongly microbrecciated tectonite gneiss shows almost complete overprinting of the original foliation. The transition from moderate to strong microbrecciation is marked by decrease in size of feldspar porphyroclasts to an average of .15 to .25 mm. There is a corresponding increase in percentage of recrystallized fine-grained, slightly altered feldspar in the groundmass (up to 1/3 total feldspar). There also appears to be a greater amount of very fine-grained white mica, probably muscovite, and this phase helps to define the foliation. The feldspar porphyroclasts which do survive are noticeably more fractured than in the moderately microbrecciated tectonite. Quartz grains are extremely elongate adjacent to the feldspar porpyroclasts. Chlorite content shows an increase and is commonly associated with opaques and epidote. Veinlets of chlorite and epidote plus calcite are moderately abundant.

Where the tectonite gneiss is indeed strongly microbrecciated, the fabric of the samples is one of intense deformation. There are abundant veinlets of chlorite, epidote, and biotite, and these follow innumerable microfaults which pervade the rock. The average size of feldspar chips is about .1 mm, with the largest feldspars only .3 mm or so. The fine-grained groundmass has an average grain size of only .0005 mm. It consists of quartz and feldspar and comprises an average of 50% of the rock. The larger feldspar chips as well as lenses of coarser quartz (.005 to .03 mm) float in the matrix. Original foliation is absent except where preserved in some micro-fault blocks. Many of the strongly microbrecciated rocks have a 'swirled' appearance due to the abundance of anastomosing fractures and microfaults which lace the rocks. Tiny breccia zones are common, as are veinlets of chlorite and opaques. Although foliation may be preserved in some of the microscopic fault blocks, there are abrupt changes in orientation of the foliation from block to block.

In outcrop, the rocks of the décollement zone are fine-grained and pervasively overprinted by shattering along closely-spaced fractures. Lineation and foliation of the original tectonite gneiss can be recognized, but these are generally masked by the microbrecciation. On fresh surfaces the rocks are sometimes brilliantly blue-green, but their typical weathered color is a drab brown.

Mapping the internal structural fabric of the microbrecciated mylonitic tectonites is very difficult, but at the same time very revealing. Measurements of relict foliation and lineation in the microbrecciated tectonite disclose that the décollement zone is made up of innumerable fault-bounded blocks of a rotational nature. One is struck by the similarity of rotational faulting both on microscopic and macroscopic scales. At the Saguaro National Monument site, relict foliation in the décollement zone generally departs radically from the normal north-northeast strike to an average which is northwest (see Figs. 21 and 25). The typical east-northeast lineation trend for normal tectonite shifts to other orientations, notably southerly. Locally the foliation is vertical, and it is possible that relict foliation in some blocks has been overturned, although this is difficult to prove.

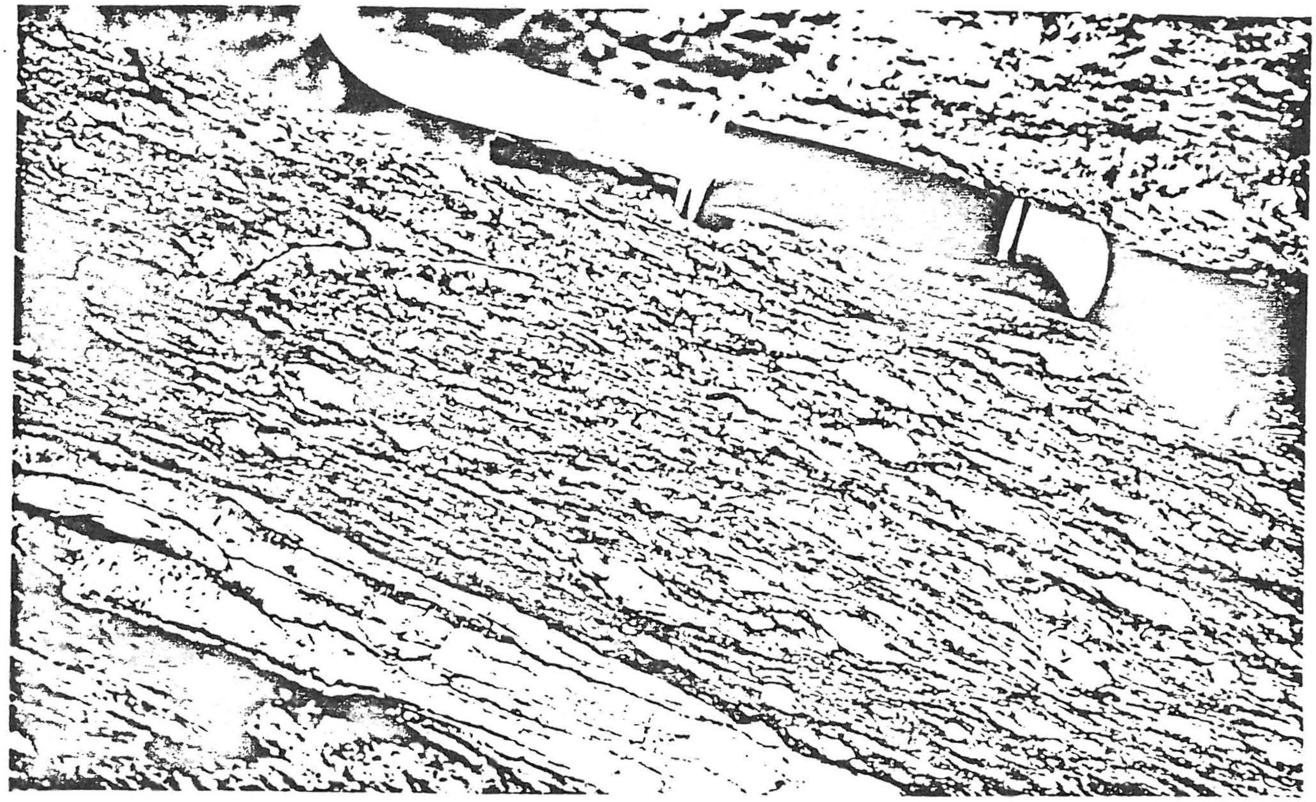

Figure 23. Foliation in tectonite gneiss.

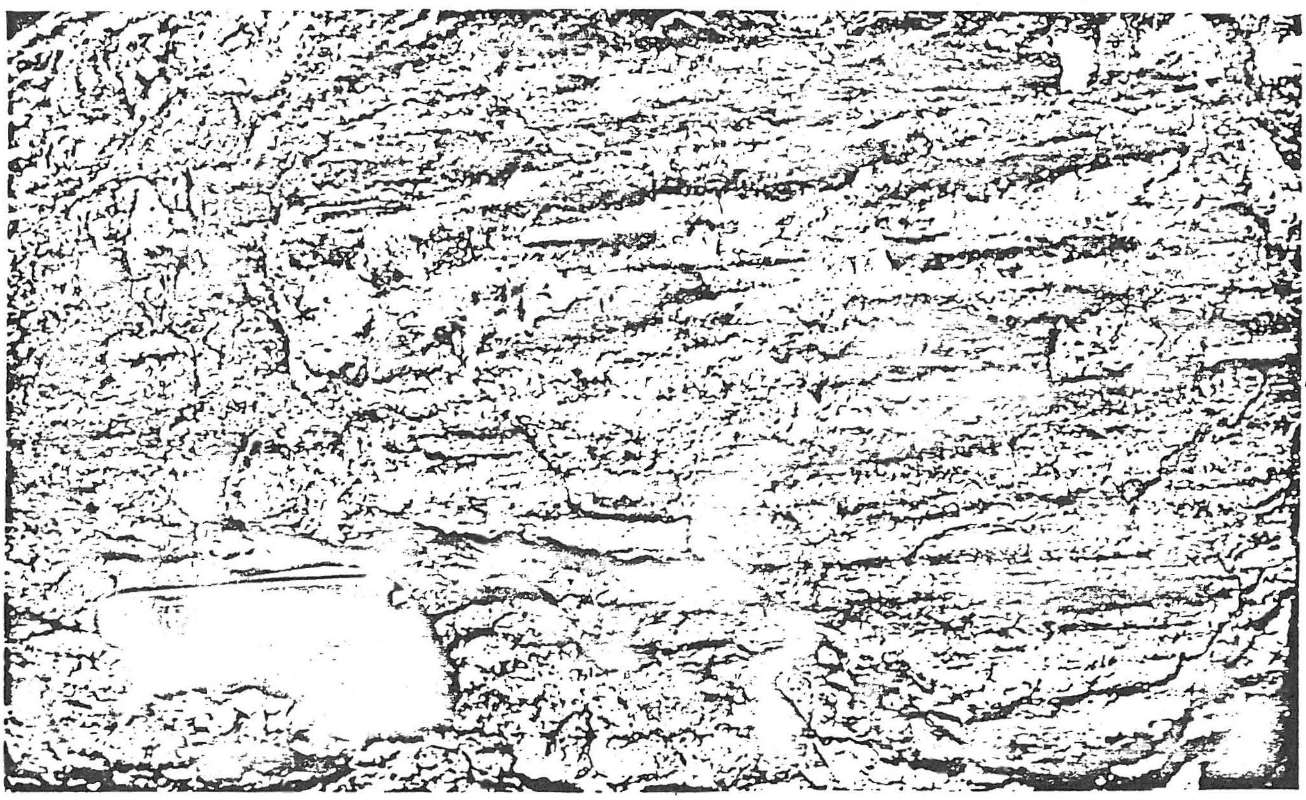

Figure 24. Lineation in tectonite gneiss.

Davis, et. al. (17)

The surprising discovery afforded by detailed structural mapping is that the changes in foliation-lineation attitude are not random, but rather, are very systematic. Domains can be identified within which foliation and lineation are uniform in orientation. These are separated from adjacent domains by inferred or poorly exposed fault boundaries. Where fault displacements are observable they typically show normal separation.

The structural characteristics which make the décollement zone so distinctive formed in the late stages of and/or after the formation of mylonitic tectonite. The microbrecciated mylonitic tectonite which makes up most of the décollement was produced during imbricate faulting and rotation of the mylonitic tectonite. In parts of the décollement zone where mylonitic tectonite is only modestly overprinted by microbrecciation and faulting, there is bimodality of strike orientation of the relict tectonite foliation: some remains in normal attitude, some is rotated to a strike which is nearly perpendicular to original strike.

In effect the décollement zone appears to be a zone of coalescence of gently to moderately dipping faults, some of which may be the lower reaches of listric normal faults. Multiple superposed fault movements locally achieved radical rotation of originally gently dipping foliation. Although the microbrecciation and rotation post-dated the formation of the original mylonitic tectonite, the tectonic movements which produced microbrecciation and rotational faulting seem to be coordinated with those which produced the original lineation. This is shown by the tendency of the strike of rotated foliation to become aligned at right angles to the trend of lineation in underlying mylonitic tectonite. Like that of the mylonitic tectonite, the strain significance of the formation of microbrecciation and rotational faults in the décollement zone is accommodation to near vertical flattening and east-northeast elongation.

Faulting, rotation, and microbrecciation were carried out at a time and at a structural position such that no upper-plate detachment strata became interleaved with microbrecciated mylonitic tectonite of the decollement zone. The very fine-grained microbreccia ledge beneath the décollement surface caps the underlying décollement zone rocks, typically in angular discordance. It too is probably the coalescence of faults (Shackelford, 1977, 1980). Rocks in the capping ledge are so thoroughly microbrecciated that no relict foliation can be seen.

Detachment Rocks.

The décollement surface at the Saguaro National Monument site strikes north-northeast and dips gently northwest. Overlying the décollement are detachments of non-tectonite Precambrian, Paleozoic, Mesozoic, and Tertiary rocks. In the immediate area of interest, the upper-plate detachment rocks are Pennsylvanian(?)-Permian limestones and mudstones, intricately folded and faulted but not generally metamorphosed.

Rocks and structures of the Paleozoic and Mesozoic detachments of the Tanque Verde and Rincon Mountains have been described and discussed by Davis (1973, 1975). Davis proposed a gravitational-tectonic deformational model to explain the formation of the overturned to recumbent fold structures which occupy "cover rocks" exposed from place to place around the base of the gneissic complex. Such folds are very well exposed in Saguaro National Monument (East) and a map revealing some of the complexity of internal deformation appears as Figure 26. The model for gravity-induced folding of such rocks argued against the concensus interpretation that the folds were produced by

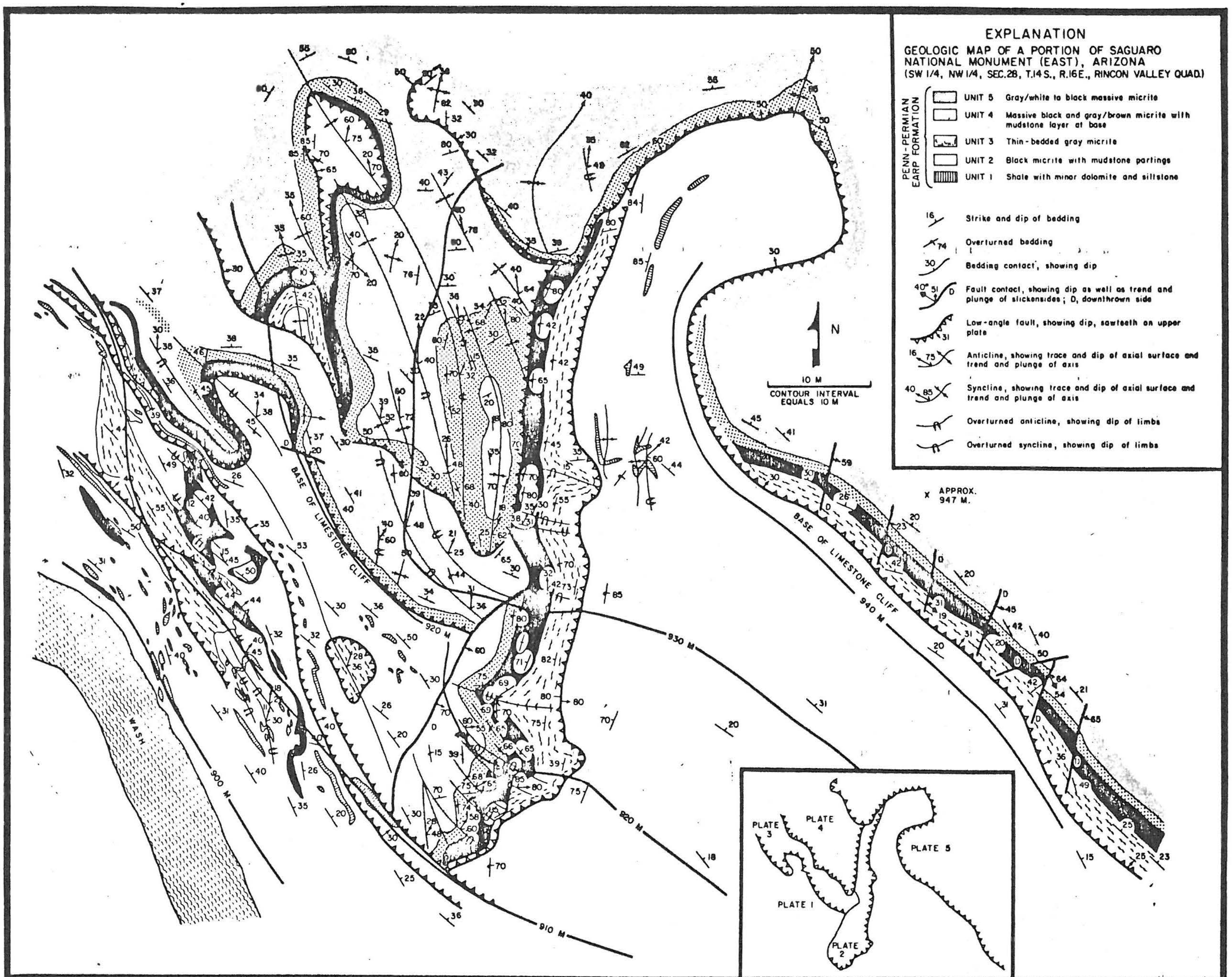

Figure 26. Map of folds in detachment at Saguaro National Monument (Davis and Frost, 1976).

northeast-directed Laramide regional overthrusting.

The gravity-induced folding model of Davis (1975) is here reevaluated in the light of new data amassed over the past five years. Folds in these rocks may well be the result of southwest-directed simple shear, not radially directed gravity gliding. Simple shear and attendant folding accompanied, at least in part, the development of lineated mylonitic tectonite.

In effect, the orientation of all folds in Paleozoic and Mesozoic tectonite and detachment rocks in the Rincon Mountains area are compatible with a model of overall southwestward directed simple shear. This is most keenly reflected in orientations of axial planes of all folds. The modal strike of axial planes is at right angles to penetrative lineation in mylonitic tectonite. Furthermore, the dip of axial planes is consistently northeastward. Such a dip-direction vergence, strongly supports southwest-directed shear. Unlike the axial planes, the axes of the folds show differences from domain to domain, but the axes orientations coupled with style of asymmetry are consistent with southwestward movement. Within the overall southwest-directed movement pattern, there tends to be southwestward divergent flow, with components of movement down the dip of the major antiformal arches like Tanque Verde antiform. This divergence contributed to the "apparent" radial pattern which Davis (1975) emphasized.

Many kinds of relationships can be cited which document the parallelism of movement plan in tectonite and detachments. The décollement zone in the Coyote Mountains contains slickenside striae which are parallel to the mineral lineation in underlying tectonite. Large grooves mapped by Drewes (1978) in the décollement zone of the southern Rincon Mountains near Colossal Cave and on the nose of Tanque Verde Mountain trend N60°-70°E, parallel to mineral lineation in underlying tectonite. The grooves and mullion are likewise parallel to slickenside striae in granite and limestone in the southern Rincon Mountains (Davis and others, 1974). Also, sedimentary strata in detachments, like those in Happy Valley on the eastern side of the Rincon Mountains, strike north-northwest on the average, perpendicular to the east-northeast transport direction.

To sum up, the presence of upper Paleozoic detachment strata atop microbreccias and mylonitic tectonite derived from Precambrian protolith is the net effect of sustained shearing and faulting along a common line of movement. Whether these distinctly different structural assemblages were produced during a continuum of deformation or as a result of discrete superposed events is not presently known. The mylonitic gneisses represent a deep exposure of a regional shear which accomodated normal-slip displacement in a southwestward direction. Not only was Precambrian plutonic rock converted to mylonitic tectonite, but overlying sedimentary rocks of younger Precambrian and Paleozoic age were also converted to tectonites. During prograde metamorphism of the sedimentary rocks, southwestward-directed simple shear resulted in folding and partial transposition of the sequence. The result of the shearing was profound vertical flattening and east-northeast extension (elongation) of all rocks in the zone of tectonite formation. The net effect of the thinning and flattening was a lowering of strata which had been in upper structural positions (e.g., upper Paleozoic strata or Cretaceous strata) to closer proximity with mylonitic tectonite derived from Precambrian protolith (Davis, 1977). At some stage during the change from moderately ductile to moderately brittle deformation, earlier formed mylonitic tectonite was itself disrupted by the shearing process. As a consequence, the décollement zone was formed as a zone of coalescing imbricate faults. The presence of hydrothermal waters during retrograde metamorphism, microbrecciation, and shattering resulted in the thorough, chloritic alteration. The final emplacement of detachment strata onto the top of the décollement zone carried already-folded strata along listric(?) normal faults. Some newly formed folds may have developed as well, but this is difficult to demonstrate. The final emplacement of detachment strata occurred in mid-Miocene time, apparently between 20 and 15 m.y. ago.

Awareness of the kinematics of southwestward-directed movement permits the Rincon Mountains metamorphic core complex to be viewed in an illuminating way. Using J. Hoover Mackin's (1950) down-structure method for viewing geological maps, the map of the Rincon Mountains, prepared by Drewes (1978), can be converted into a structural profile. This is shown in Figure 27. The décollement can be seen to be a broadly arcuate mullion-like fault surface separating mylonitic tectonites below from a variety of upper-plate detachment rocks above. Mylonitic tectonites are seen to be "thin" lens-shaped structural units, some of which are reduced to mylonitic schists at their edges and terminations. On Tanque Verde ridge, the gneisses derived from Tertiary granodiorite climb as a tabular body up-structure, perhaps revealing a distorted vestige of the original intrusive relationship of the granodiorite into Precambrian quartz monzonite. The strong tectonite overprint is such that foliation and lineation cut through the granodioritic gneissic phase discordantly. The deformation thus was passive.

The major sites of formation of mylonitic schist are at contacts between dissimilar gneissic phases and at the deep mullion-like trough in the Rincon Valley re-entrant. The symmetrical distribution of mylonitic schist about the Rincon Valley trough suggests that strain was preferentially concentrated along this "keel" of the detachment system.

Structural units in the upper-plate detachments, whether they be composed of sedimentary or volcanic rocks, are thin and lensoidal, thoroughly pervaded by extensional faults and tear faults. From Tanque Verde ridge southeastward, four detachment domains are recognized: these contain Cretaceous sedimentary rocks, Precambrian Rincon Valley granodiorite, Paleozoic sedimentary rocks, and Oligocene (?)-Miocene sedimentary rocks. Strain styles within the detachments are dissimilar: Paleozoic and Cretaceous strata display exquisite overturned to recumbent folds; Precambrian granodiorite is so shattered that "it would be impossible to find within it a headstone for a grave" (E.B. Mayo, pers. comm., 1972); and the Oligocene(?)-Miocene strata display faulting and rigid-body rotation to northwesterly strikes. In Rincon Valley (Drewes, 1978), the detachment of Rincon Valley granodiorite rests on tectonite derived from Paleozoic rocks, remnants of earlier ductile penetrative simple-shear deformation. Just southeast tectonite gneiss derived from the Tertiary granodiorite contains interlayers of tectonite derived from Paleozoic impure carbonates and quartzites (Drewes, 1978) (see Fig. 27). These probably are overprinted roof pendants, disclosing that the Tertiary granodiorite was originally intruded to structural/stratigraphic levels <u>above</u> the Precambrian basement.

Although the Rincon Valley granodiorite has been interpreted to be a Laramide thrust plate (Drewes, 1978), Figure 27 reveals that it can be viewed simply as one of a variety of country-rock elements that were emplaced by detachment faulting as young as mid-Miocene. Due to frictional shear as the detachments were emplaced, narrow antiformal crests of mullion ridges like Tanque Verde Mountain, and the smaller mullion counterparts in the vicinity of Colossal Cave (Fig. 27), were subjected to high stress and attendant development of cataclastic fabrics, especially microbreccia and "ultramylonite".

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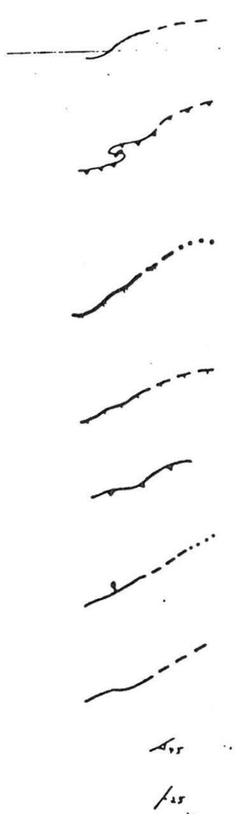
TECTONIC MAP OF THE NORTHEASTERN RINCON MTNS., ARIZONA

KEY FOR MAP AND CROSS SECTIONS

STRUCTURE SYMBOLS



Depositional or intrusive contact; dashed where position and/or interpretation is less certain.

Basement-carapace plated fault contact; dashed where position and/or interpretation is less certain. Sawteeth on carapace (upper plate), hooked sawteeth on overturned basement (lower plate).

Detachment fault, a low- to moderate-dipping dislocation surface; dashed where position and/or interpretation is less certain, dotted where concealed. Double tick marks on upper plate (allochthon).

Low-angle denudation faults of the allochthon; dashed where position and/or interpretation is less certain. Tick marks on upper plate.

Thrust fault. Open sawteeth on upper plate.

Listric (?) normal fault; dashed where approximately located and/or interpretation is less certain, dashed where concealed. Ball and bar on downthrown block.

High-angle normal fault; dashed where approximately located and/or interpretation is less certain.

Strike and dip of foliation in metamorphic basement.

Strike and dip of bedding.

ROCK UNITS

Post-Allochthonous Units

Qal - Alluvium.

QTg - Basin fill gravels associated with Basin and Range faulting, these units are in general flat-lying.

Tsp - Paige gravels (=Mid-Miocene Big Dome Formation?), consolidated alluvial conglomerates, fanglomerates, sandstones, and rare tuffaceous sandstones.

Allochthonous Units

Tm - Mineta Formation, Oligocene (=Pantano Formation), sequence of bright red to purple conglomerates, sandstones, limestones, and minor volcanics.

Kbl - Bisbee Group, predominately Glance cobble- to boulder-conglomerates with lesser red sandstones and shales.

Pzu - Pennsylvanian-Permian Paleozoic strata, (predominately Horquilla and Earp Fms), these rocks may be intensely fractured and/or brecciated, but are unmetamorphosed.

Pzl - Cambrian-Mississippian Paleozoic strata, (Belsa, Abrigo, Martin, and Escabrosa Fms), these rocks may be intensely fractured and/or brecciated, but are unmetamorphosed.

Xj - Johnny Lyon Granodiorite (=Rincon Valley Granodiorite or Happy Valley), predominately light-colored porphyritic quartz monzonite through granodiorite (a 1625 m.y. granite).

Xp - Pinal Schist, greenschist facies pelitic schists, sandstones, and metabasic rocks, Precambrian metamorphism.

Core Rocks

Th - Happy Valley Quartz Monzonite, 28 m.y.b.p. biotite K-Ar date.

Tw - Wilderness suite, two-mica garnet-bearing granite (45-35 m.y.b.p.).

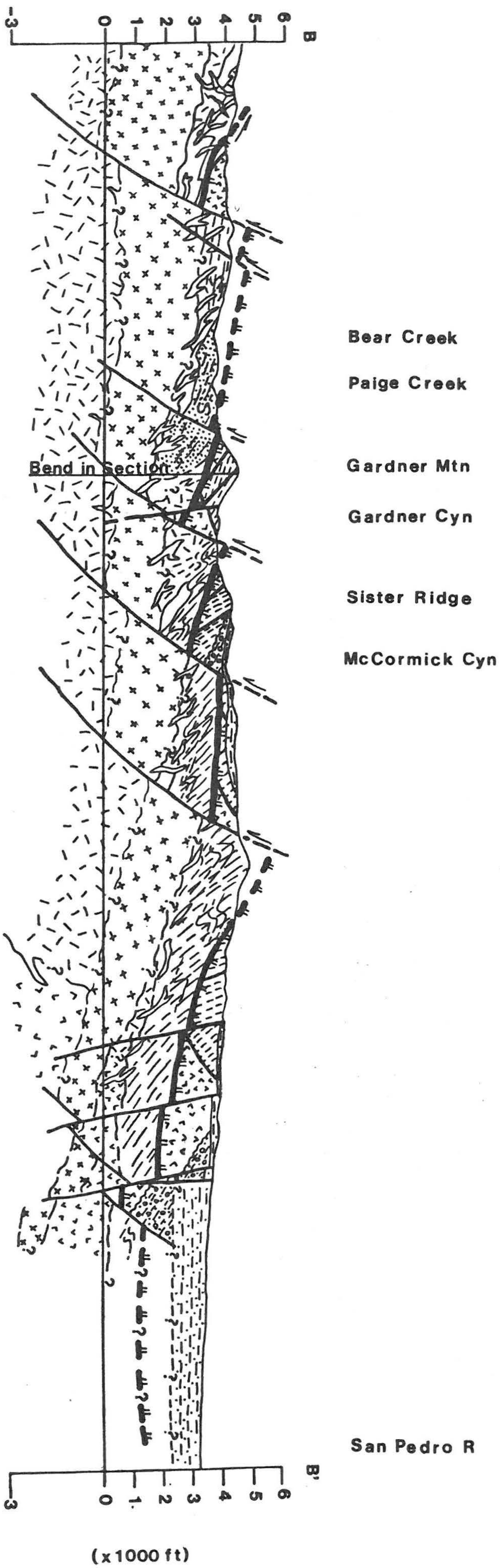
Pzm - Metamorphic tectonite carapace of deformed Paleozoic strata, includes marbles, calc-silicate marbles, quartzites, and minor pelites. Locally includes meta-sedimentary units derived from Precambrian Apache Group rocks.

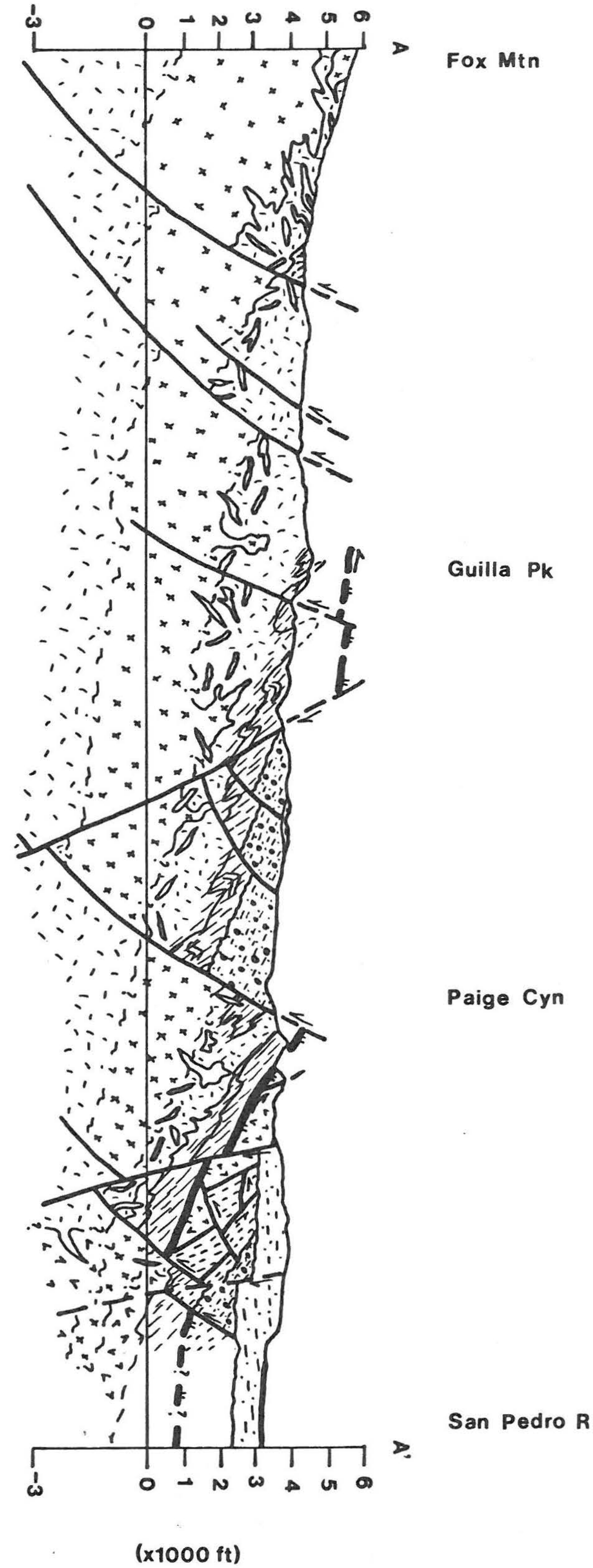
dbm - Amphibolite (metamorphosed Precambrian diabase).

grm - Granite gneiss (may be a metamorphosed Precambrian or possibly younger pluton).

Ycm - Continental Grandodiorite (=Oracle Granite?) mottled, porphyritic quartz monzonite through granodiorite, metaplutonic and locally deformed into gneiss.

Xpm - Pinal Schist, affected by Tertiary dynamothermal metamorphism.


B
Bear Creek
Paige Creek
Bend in Section
Gardner Mtn
Gardner Cyn
Sister Ridge
McCormick Cyn
San Pedro R
B'
-3 0 1 2 3 4 5 6
(x1000 ft)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Guilla Pk
Paige Cyn
San Pedro R
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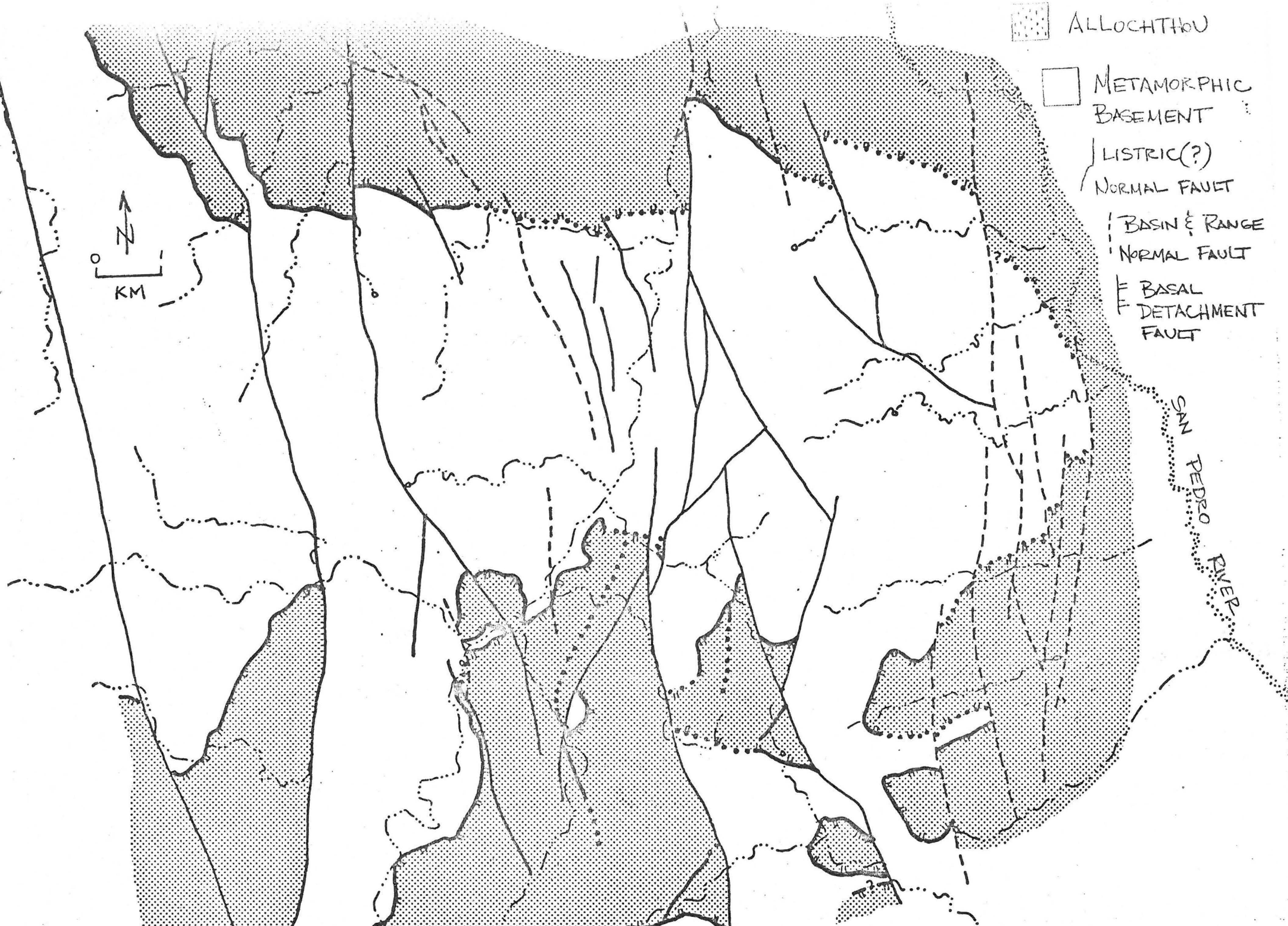
FIELD NOTES AND SKETCHES

FIELD NOTES AND SKETCHES

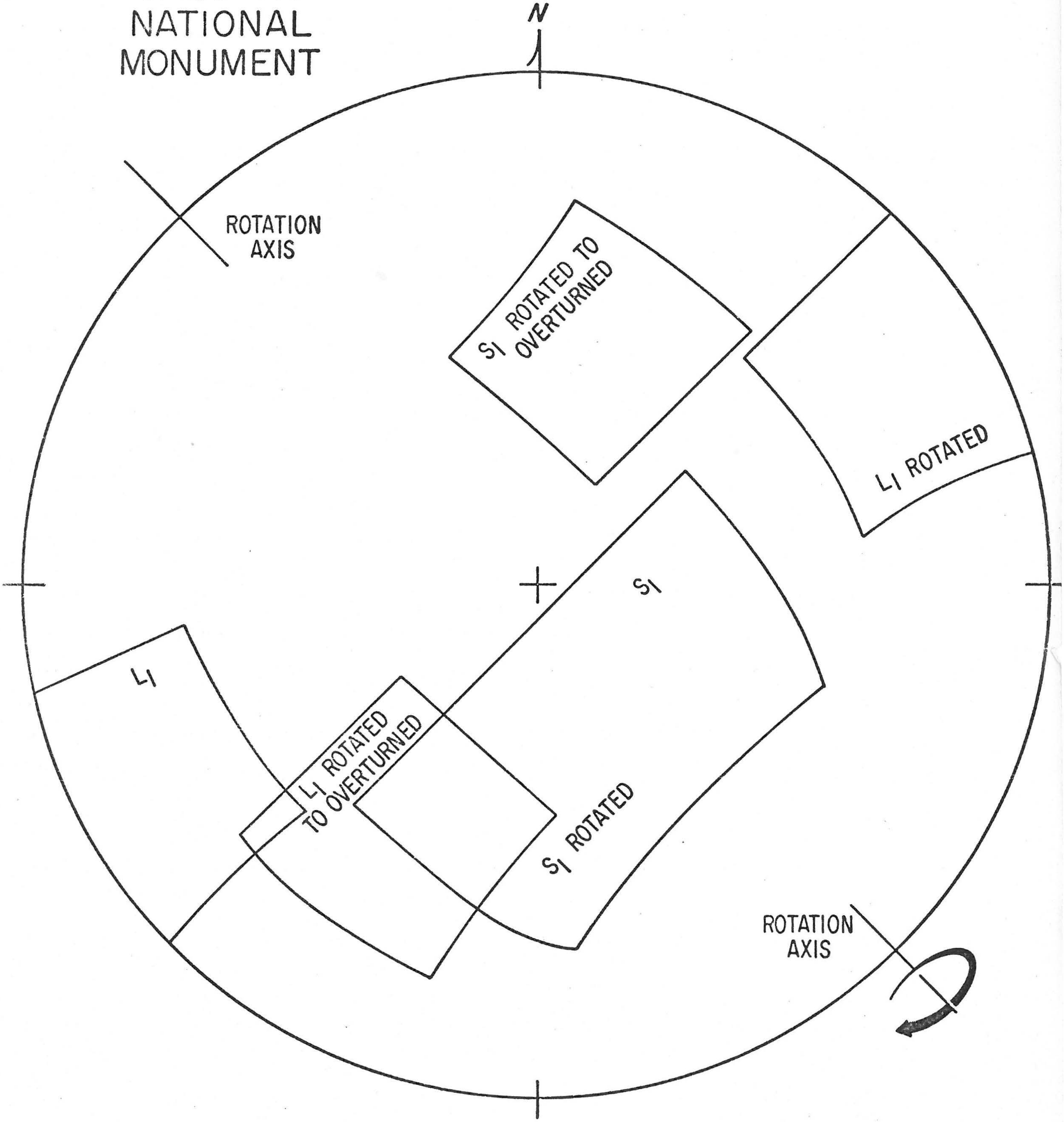
FIELD TRIP

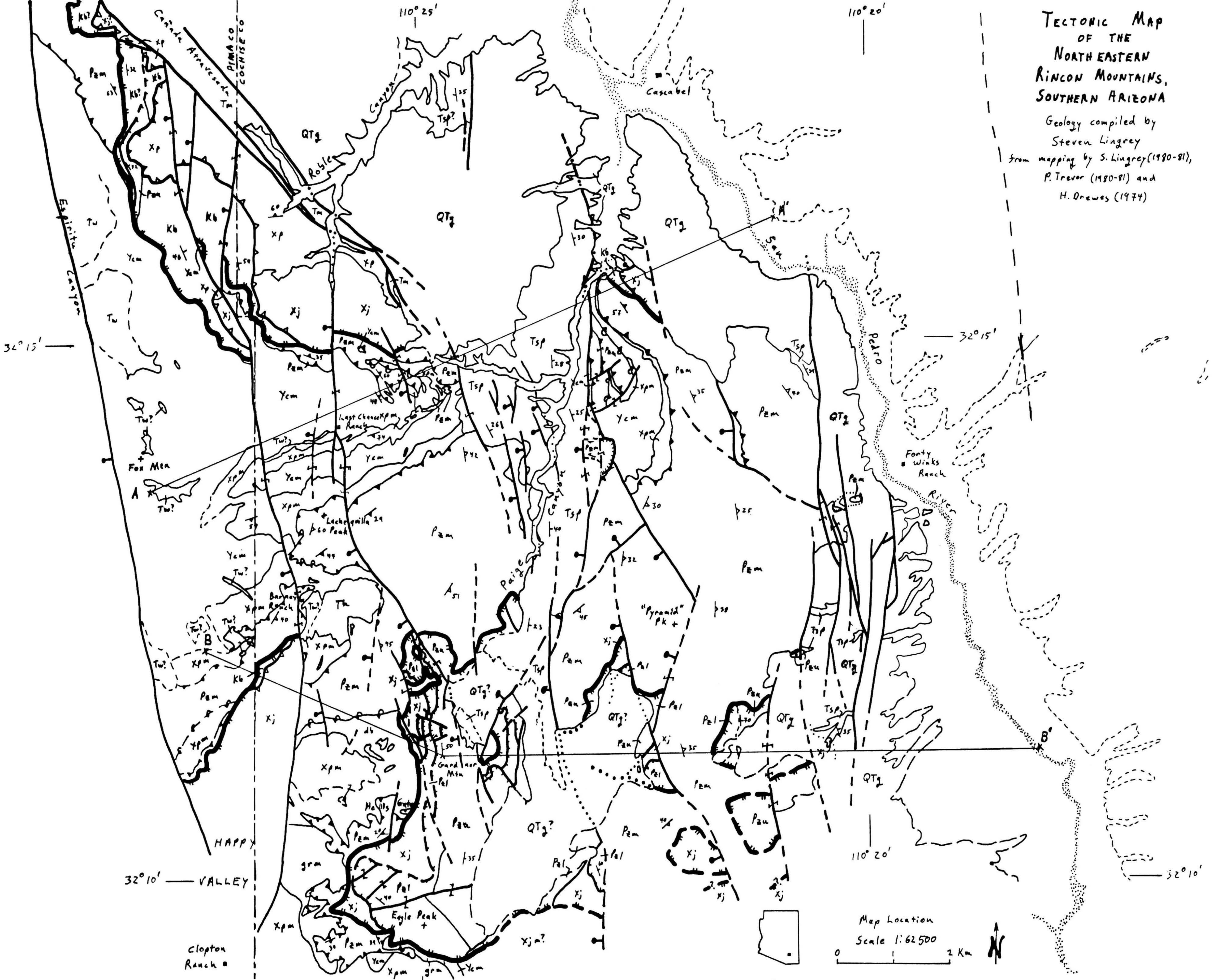
HANDOUTS

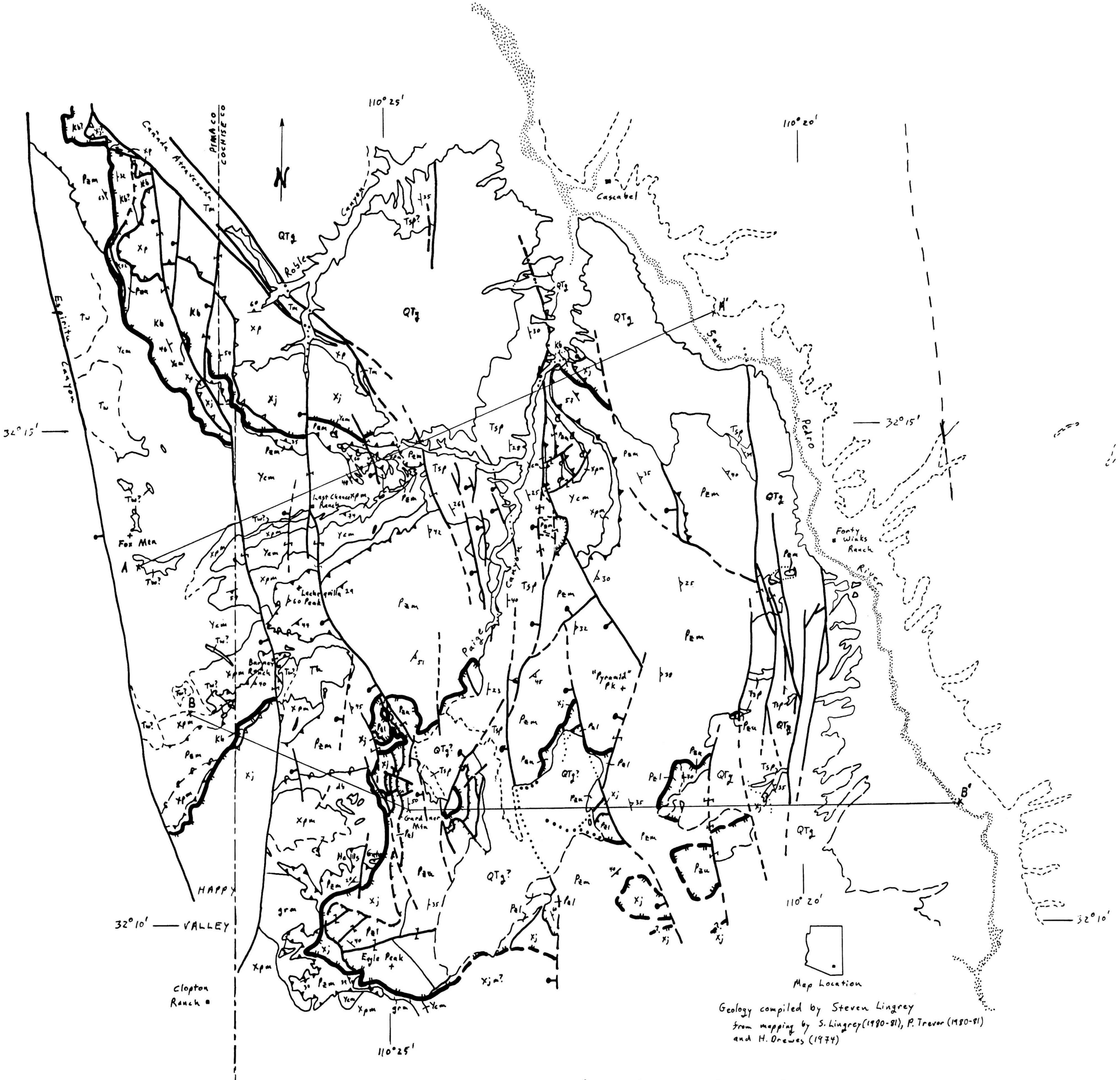


BASAL DETACHMENT SURFACE & ALLOCHTHON IN THE NORTHEASTERN RINCON MOUNTAINS.

SAQUARO NATIONAL MONUMENT




TECTONIC MAP OF THE NORTHEASTERN RINCON MOUNTAINS, SOUTHERN ARIZONA
Geology compiled by Steven Lingrey from mapping by S. Lingrey (1980-81), P. Trevor (1980-81) and H. Drewes (1974)
110°25'
110°20'
32°15'
32°10'
Cascabel
Cañada Atravesada
Roble Canyon
PIMA CO
COCHISE CO
Espiritu Canyon
San Pedro River
Last Chance Ranch
Fox Mtn
Lechuguilla Peak
Barney Ranch
Forty Winks Ranch
"Pyramid" Pk
Paige Canyon
Gardner Mtn
Happy Valley
Eagle Peak
Clopton Ranch
A
A'
B
B'
Map Location
Scale 1:62500
0
1
2 Km
N



TECTONIC MAP OF THE NORTHEASTERN RINCON MOUNTAINS, SOUTHERN ARIZONA

Scale 1:62500

0 1 2 Km