STRUCTURAL GEOLOGY OF THE MT. BIGELOW-
BEAR WALLOW-MT. LEMMON AREA,
SANTA CATALINA MOUNTAINS, ARIZONA

by

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I hereby recommend that this dissertation prepared under my direction by Charles J. Waag
entitled STRUCTURAL GEOLOGY OF THE MT. BIGELOW-BEAR WALLOW-MT. LEMMON AREA, SANTA CATALINA MOUNTAINS, ARIZONA
be accepted as fulfilling the dissertation requirement of the degree of Doctor of Philosophy

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ABSTRACT

In the vicinity of Mount Bigelow, Bear Wallow, and Mount Lemmon, the Precambrian Apache Group and Cambrian rocks are intensely deformed. This deformation is manifest as faulting, décollement and folds of varied styles. Analysis of the fold styles, symmetries, and orientations reveals at least two and probably three systems of folds. Folds of the \( F_1 \) system are synmetamorphic, deform the sedimentary layering \( (S_1) \), and have a genetically related axial plane foliation designated \( S_2 \). The intersection of the two \( S \) surfaces forms a stretched mineral lineation parallel to the "b" tectonic axis; however, stretched mineral lineation parallel to the "a" axis is also present in the folds of the \( F_1 \) system.

Folds of the \( F_2 \) system are superimposed upon folds of the \( F_1 \) system, deform both \( S_1 \) and \( S_2 \), and were formed in the waning stages of, or after, the main metamorphic event. Microcrenulations parallel to the "b," and stretched mineral lineation parallel to the "a" and "b" tectonic axes are common. Locally stretched mineral lineation parallel to the "b" tectonic axis of the \( F_1 \) system is deformed by \( F_2 \) folds.

The \( F_1 \) and \( F_2 \) systems of folds are generally homaxial, and knee-"isoclinal" similar folds are common to both
systems. These combined styles as well as other folds were formed by gravitational gliding, and express the ductility contrast and mean ductility of the rocks during deformation. Interpretations of the kinematics involved in the formation of the folds indicate movements down structural gradients established by centers of active upwelling. One such center of upwelling was in the Catalina Gneiss south of Mount Bigelow. The Leatherwood Quartz Diorite pluton was also reactivated and served as a second center of upwelling producing southward and southwestward tectonic "currents" in the Bear Wallow and Mount Lemmon areas, respectively. These centers of upwelling seem to have been active at essentially the same time; however, upwelling within the Catalina Gneiss probably remained active for a longer period and reached a higher elevation.

K-Ar dates from micas in the Catalina Gneisses yield an average age of 26.8 ± 1.7 m.y. This date indicates that uplift in the Santa Catalina Mountains had progressed to the point that metamorphism had terminated by late Oligocene-early Miocene time and thus dates the $F_2$ generation of folds.

The general homoaxiality of the folds, and similarity of styles in the $F_1$ and $F_2$ systems suggests that the same centers of upwelling generated both systems. Divergence of the axes is attributed to changes in the tectonic
slope, and to minor shifts in the centers of active upwelling with time. Generation of the folds of the $F_1$ and $F_2$ systems by one continuous episode is favored, and a late Eocene or early Oligocene time is tentatively suggested for the beginning of the generation of the $F_1$ folds.

A third system of relatively broad, open, upright folds formed by vertical uplift and lateral crowding during the final stages of activation of the Catalina Gneiss and emplacement of the Catalina Granite is also identified ($F_3$). The activation of the gneiss and the intrusion of the granite were probably controlled by previously established fracture directions, which localized heat introduced into the area. The resultant actively rising welts and attendant gravity tectonics share many similarities with mantled gneiss and migmatite domes.
INTRODUCTION

Purpose of the Investigation

The Precambrian and early Paleozoic rocks of the Mount Bigelow-Bear Wallow-Mount Lemmon area have undergone one or more periods of deformation. This deformation is manifest as faulting, decollement, and folds of varied styles, symmetries, and orientations. The objective of this study is to determine the mechanism and sequence of formation of the structures, to define in detail the contact relationships between the Catalina Gneiss and the overlying rocks, and to relate the structures to movements in the Santa Catalina Mountains.

Location and Topography

The study area is approximately 18 miles northeast of Tucson (Figure 1), covers 5 square miles, and surrounds Mount Bigelow and Mount Lemmon, the two highest peaks in the Santa Catalina Range. Relief is generally rugged, and heavy pine and aspen forest covers much of the area. Good outcrops are largely restricted to road cuts and cliff faces; however, smaller exposures are locally revealed through a heavy mat of pine needles and soil.
Figure 1. Location map showing areas of present study
Methods of Study

The fieldwork for this study was done during the summer of 1966, the spring and fall of 1967, and the spring of 1968. Approximately six months were spent in the area.

The geologic data were recorded on maps of various scales. The choice of scale was at least in part dictated by the complexity of the structure to be recorded and the quality of the outcrops available. The main road cuts were mapped using a Brunton Compass and an open-site alidade on scales ranging from 50 feet to the inch to 200 feet to the inch.

Enlargements of 1966 and 1958 Forest Service aerial photographic coverage were also used as field sheets. The scales of the enlargements are variable, but approximate 300 feet to the inch. Many data sites were located on the enlargements with the aid of 9 inch by 9 inch stereo pairs of the same coverage. The data thus recorded were then generalized and compiled on the final map scale of 400 feet to the inch.

The base maps for the final compilation (Figures 2 and 3, in pocket) were redrafted from photographic enlargements of the terrain and culture overlays of the Bellota Ranch and Mount Lemmon, Arizona, quadrangles. The scale of the overlays from which the enlargements were made is 1:48,000.
Twenty thin sections of rocks collected in the field were prepared and examined with a petrographic microscope. Eighteen additional specimens were slabbed, polished and stained with sodium cobaltinitrite and amaranth solutions for feldspar determinations. X-ray diffraction was used to determine the composition of the finer grained rocks in the area.

Much of the structural data gathered in the field was plotted on Schmidt equal area projections as an aid to interpretation. As an additional aid to understanding the relationship between the lineation and the fold axes, and the mechanism of folding, a number of hand specimen sized folds were collected, cut, and polished.

Previous Investigations

The Geologic Map of Pima and Santa Cruz Counties, Arizona, prepared by the Arizona Bureau of Mines covers the entire study area on a scale of 1:375,000. This and discussions of the general geology of the Santa Catalina Mountains by Bromfield (1952) and DuBois (1959) contain the only published information on the area of investigation.

Peirce (1958) mapped and described the rocks in the Bear Wallow and Mount Bigelow areas as part of his doctoral dissertation on a much larger area in the Santa Catalina Mountains. The large size of his area, and the lack of
accurate topographic maps or aerial photographs for control, allowed Peirce to make only a limited structural study.

Wood (1963), in a study of the metamorphic effects of the Leatherwood Quartz Diorite upon its wall rocks, mapped a portion of the Bear Wallow area near Butterfly Peak. This same locality and an additional small segment to the east were remapped by Hanson (1966) in his study of the structure and petrography of the Leatherwood Quartz Diorite.

Pilkington (1962) mapped the area east of Mount Bigelow including Kellogg Mountain. A small portion of Kellogg Mountain is included in the present study.
ROCK UNITS

Precambrian Apache Group

Pioneer Formation

The basal unit of the Apache Group, the Pioneer Formation, was originally designated the Pioneer shale by Ransome (1903) for exposures in the Ray quadrangle, Arizona. Since that time, the unit has also been referred to by many as the Pioneer Quartzite and the Pioneer Formation. Most recently, Creasey (1967) in a report on the geology of the Mammoth quadrangle, Arizona, used the formation designation and divided the formation into three members. In ascending order the members are the Scanlan Conglomerate, a pebbly sandstone or quartzite middle member, and a tuffaceous upper member. Shride (1967) and Willden (1964) recommended that the Scanlan be included as the basal part of the Pioneer Formation, but chose to consider the Scanlan a bed rather than a member. In this study, the Scanlan is considered the basal member of the Pioneer Formation, but the middle and upper members of the Pioneer recognized by Creasey (1967) are not distinguishable. Rocks probably equivalent to them are here informally referred to as the phyllite member.

Within the area covered by this report, the Scanlan has been positively identified by the writer only on Mount Kellogg. Elsewhere the Scanlan is missing, or for various
reasons can only be tentatively identified. At Marshall Gulch, for example, a metaconglomerate that is probably Scanlan crops out within a zone of mixed pegmatite and gneiss. At this place, the clasts of the conglomerate are coarsely recrystallized, flattened, and stretched, and the usual quartz sericite matrix has been recrystallized and "flooded" with feldspar from the adjacent pegmatite and gneiss. Biotite and muscovite are abundant in the recrystallized matrix. In places the rock is so gneissose that its original identity is almost completely destroyed. The apparent absence of the Scanlan throughout most of the area may be in part the result of poor exposures and of facies change; however, lack of deposition, structural complications, and metamatism, transfusion, or granitization are probably principally responsible for its absence.

The Scanlan metaconglomerate exposed on the southwest slope of Mount Kellogg occurs as a lens several hundred feet long and perhaps as much as eight feet thick. Exposures are poor, large displaced blocks cover much of the outcrop, and metaconglomerate rubble is strewn over much of the slope between the outcrop and the Mount Lemmon Highway. The rock is composed of pebbles and cobbles of quartz and quartzite, and rarely of chert. The clasts range from white to dark gray and are set in a light brown to gray schistose matrix of quartz, sericite, and minor feldspar. Most clasts are extremely stretched into slightly flattened cigar-shaped
ellipses having axial ratios averaging about 10:2:1. Some of the stretched clasts have spine-like tips, attesting to a plastic condition during deformation. Recrystallization pervades the clasts and matrix quartz alike, and imposes a secondary foliation on the metaconglomerate. This secondary foliation is also common in the overlying phyllite member, and is difficult to distinguish from the original stratification.

The basal part of the Pioneer is apparently not everywhere so obviously conglomeratic, but in places may be represented by a quartzitic phyllite that contains disseminated quartz blebs or lenses. These blebs may be metamorphosed clasts, or else lenses that resulted from quartz segregation during metamorphism. Similar relationships are discussed in more detail in the description of the Barnes metaconglomerate.

The phyllite member of the Pioneer Formation weathers comparatively rapidly and occupies a narrow rubble-covered zone along the south flanks of Mount Kellogg and Mount Bigelow. The phyllite is overlain discontinuously by Barnes metconglomerate and commonly rests directly upon the Catalina Gneiss. The lower contact is very irregular and is generally intrusive, but in some places seems to be gradational and has a granitized character. The member is composed of gray, green, and reddish-brown iron-stained
quartz-sericite phyllite and phyllitic quartzite. An abundance of muscovite and sericite gives the rock a satiny sheen. Most samples show microcrenulations or chevron folds as large as an inch in amplitude.

Generally the abundance of folds, the amount of recrystallization and quartz segregation, and the coarseness of the grain size within the phyllite member increase with proximity to the contact with the Catalina Gneiss. These features probably represent increased metamorphism near the contact; they become especially pronounced in isolated remnants south of the main contact, and along the main contact at places where a thin veneer of Apache Group phyllite or quartzitic phyllite overlies the gneiss. This increased metamorphism also results in a subtle change in the texture of the member from phyllitic to schistose and locally gneissose.

The rocks of one such remnant of Apache Group cropping out along the Mount Lemmon Highway above Boy Scout Spring were regarded by Peirce (1958) as Pinal Schist. Peirce concluded that the intricate folding and abundance of quartz segregations resulted from an additional phase of metamorphism not experienced by the Apache Group in the area.

This conclusion is not consistent with field observations made during this study and as outlined above; for
example, that the presence or abundance of meso- and micro-crenulations, chevron folds and quartz segregations is the result of increased metamorphism locally near the contact. In addition, structural evidence indicating a separate event seems to be lacking.

Dripping Spring Metaquartzite

The Dripping Spring Quartzite was named by Ransome (1903), who described a lower division of 175 feet of massive beds of streaked buff and pink quartzite, and an upper division of 225 feet of thinner bedded, hard laminated quartzite. The section Ransome described is exposed on Barnes Peak; however, the name he adopted was from the Dripping Spring Mountains, where similar exposures are known. In the type section on Barnes Peak, the Dripping Spring is underlain by Barnes Conglomerate and overlain by Mescal Limestone.

Shride (1967) suggested that the Barnes Conglomerate should be included as a basal bed of the lower member of the Dripping Spring Quartzite. Granger and Raup (1964) gave the Barnes Conglomerate formal recognition as the basal member of the Dripping Spring Quartzite.

Creasey (1967) in his work in the Mammoth quadrangle along the northeast flank of the Santa Catalina Mountains similarly divided the Dripping Spring into three members: the basal Barnes Member; a middle member composed of pink
to maroon medium-grained feldspathic quartzite; and an upper member of thinly interbedded fine-grained micaceous quartzites and shale. Creasey noted that the matrix of the Dripping Spring is partially recrystallized, indicating a mild metamorphism probably brought about by intrusion of pre-Paleozoic diabases in the section.

Higher rank metamorphic equivalents of these rocks can be recognized in the present area of investigation approximately nine miles south of their exposure in the Mammoth quadrangle.

**Barnes Metaconglomerate.** The Barnes metaconglomerate occurs as a series of discontinuous outcrops in the Mount Bigelow-Mount Kellogg vicinity, and in a small exposure at the base of a waterfall on the west slope of Mount Lemmon.

Typically, the unit contains stretched pebbles, granules, and rare cobbles in a matrix of quartz, feldspar, sericite, and epidote. The clasts are generally white to gray quartz and quartzite, and scattered chert and jasper fragments in a pink to green matrix. Where the Barnes is in proximity with the Catalina Gneiss or igneous rocks, biotite is also a common matrix constituent. Hematite is a common component, and where present stains the rock red.

The amount of clast elongation or flattening varies considerably with the locality, and no statistical estimates of deformation were made. Nowhere within the area of
investigation did the writer find what seemed to be undeformed clasts. Furthermore, present axial ratios are difficult to obtain, because exposures are poor, and recrystallization has been extensive and tends to obliterate clast boundaries. In fact, deformation of the clasts may be so great that the clasts are barely recognizable. Where such extreme stretching and flattening exist, the clasts become discontinuous layers of clean quartzite intercalated with more continuous bands of epidote and limonite-stained feldspar and sericite.

The Barnes metaconglomerate is poorly stratified, although the stratification in some places is accentuated by the elongation and flattening of the clasts. Some specimens display a secondary foliation at an angle to the original stratification. This foliation is pervasive; it is caused by recrystallization and passes through clast and matrix alike.

In the vicinity of Mount Bigelow, typical Barnes metaconglomerate can be traced laterally over a distance of a few tens of feet into a reddish-brown quartz-muscovite phyllite that contains isolated lenses and eye-shaped pods of quartz. This phyllite apparently represents a fine-grained, less-conglomeratic facies of the Barnes.

Observed thicknesses of the Barnes in this area range from slightly less than one foot to a maximum of
eight feet. It is not generally a ridge-former, but occurs inconspicuously at a break in slope beneath the bluff-forming middle member of the Dripping Spring metaquartzite.

**Middle Member.** The middle member of the Dripping Spring is a light gray to pink feldspathic metaquartzite that weathers reddish-brown to pink. Biotite, muscovite, red garnet, and hematite are common constituents. The biotite and muscovite are ordinarily present along laminae and cross-laminae and tend to accentuate these structures both on fresh and weathered surfaces. Hematite and garnet occur both concentrated along the laminae and scattered randomly throughout the rock.

In hand specimen, the metaquartzite appears to be well sorted and fine-grained; however, most specimens exhibit recrystallization and elongation of the grains. The extent of fabric deformation is related to the complexity of the larger structures in the area, and in some places a secondary foliation is developed. In these places, the attitude of the bedding is difficult to determine unless compositional layering is present. Fortunately, bedding in the middle member is well developed with individual beds ranging from one to four feet in thickness; quartzitic phyllite partings not uncommonly separate these beds.

The middle member of the Dripping Spring metaquartzite is the most resistant unit within the Apache Group and
is typically a bluff-former. It is well exposed in high cliffs along the north slopes of Mounts Bigelow and Kellogg (Figure 4), where it is complexly folded and intruded by pegmatite and aplite.

No meaningful estimate of the total thickness of the middle member can be made in this area because of the complex structure and dilation of the section by intruded diabase, pegmatite and aplite. However, nine miles to the north in Peppersauce Canyon, Creasey (1967) records a thickness of 204 feet for the middle member. This thickness can probably be applied to the study area with reasonable reservation, but at least one notable difference in lithology does exist. On Mount Bigelow, 70 feet stratigraphically above the Barnes metaconglomerate member, a second conglomerate is present. The upper conglomerate bed is six to eighteen inches thick and is composed of white metaquartzite clasts in a light brown matrix of fine-grained quartz, feldspar, and muscovite. The conglomerate bed is overlain by "typical" middle member feldspathic quartzite and directly underlain by five to six feet of quartzitic phyllite. Beneath the phyllite approximately 20 feet of dark amphibolite is exposed. The dark amphibolite is probably a metamorphosed diabase, indicating that originally the upper conglomerate bed was at least some 20 feet closer to the Barnes Member. The upper conglomerate bed is exposed just to the right of the steps leading to the lookout tower on Mount Bigelow.
Figure 4. Sphinx Rock on the north slope of Mount Bigelow

Large recumbent fold in the middle member of the Dripping Spring metaquartzite. Bands of aplite and pegmatite along layering tend to emphasize the structure. Note hinge of fold outlined by the dark phyllite in the pinnacle in the right foreground. Facing east.
Upper Member(?). Rocks thought to belong to the upper member of the Dripping Spring Quartzite crop out in Bear Wallow, in the vicinity of the observatory and from Turkey Flat eastward to Mount Bigelow. Unfortunately, all of the evidence for this correlation is circumstantial. None of the exposures can be traced to known sections, nor is the lithology sufficiently distinctive to preclude the possibility that these rocks are metamorphic equivalents of the Pioneer Formation. The strength of this correlation then lies in the apparent stratigraphic relationships of these rocks with overlying and underlying units.

In the Turkey Flat vicinity, the rocks are underlain by fine-grained, light brown feldspathic metaquartzites of the middle member. On the north side of the ridge north of Bear Wallow, and elsewhere along the ridge to the east, the same rocks appear to be overlain by Mescal marble. The complexity of the structures, particularly in the Bear Wallow area, makes it difficult to determine the exact relationship of the units.

The upper member is composed of interbedded sericite phyllite, quartzite, and thin conglomerate beds. The contacts between the sericite phyllite and metaquartzite are locally abrupt, but are commonly gradational. As a result, quartz-sericite phyllite and sericite-metaquartzite make up a large percentage of the member.
The sericite-bearing rocks generally have a satiny luster, and shades of gray, green and reddish-brown are common. In addition to quartz and sericite, accessory chlorite, hematite, or epidote are normally present. Less commonly, garnet is a constituent. Quartz generally occurs in thin lenses or stringers; chlorite as flakes or in small pods, and hematite as finely disseminated blebs or crystals. Some of the hematite has cubic crystal outlines, and may be pseudomorphous, replacing pyrite. Microcrenulations or puckering of the foliation are locally marked by quartz segregations.

The metaquartzites within the upper member are generally quite clean considering their intimate intercalation with phyllites. They are ordinarily white, fine-grained with a sugary texture, and become heavily iron-stained upon weathering. Locally the metaquartzites are conglomeratic at their bases.

At least one conglomerate occurs within the member, and is exposed locally on the cabin-studded ridge above the Upper Bear Wallow Campground (Figure 5). The conglomerate contains stretched and flattened pebbles in a fine-grained matrix of quartz, sericite, and feldspar. Some of the clasts show evidence of rolling as well as stretching.

Mescal Limestone

Ransome (1915) first applied the term Mescal Limestone to exposures of a thin bedded, cherty, dolomitic
Figure 5. Metaconglomerate

Stretched and flattened clasts in an unnamed metaconglomerate of the upper member(?) of the Dripping Spring metaquartzite. The metaconglomerate is exposed in a cabin driveway above Upper Bear Wallow Campground. Facing northeast.
limestone in the Mescal Range of Arizona. Shride (1967) in his study of the Younger Precambrian of southeastern Arizona, redefined the Mescal and divided the limestone into three members: a lower thin to thick bedded carbonate member; a stromatolite-bearing middle member; and an argillite upper member.

Impure recrystallized limestones and calcareous, cherty metaquartzites form a discontinuous series of exposures along the north side of the ridge north of the Upper Bear Wallow Campground, and along the Potato Patch Trail south of Butterfly Peak. No attempts are made here to correlate these calcareous rocks with the members as defined by Shride (1967); however, the rocks are assigned to the Mescal Limestone.

The impure recrystallized limestones are massive, thinlily laminated and weather to a dull gray. On fresh fractures, the rock shows alternating white and light yellowish-green laminae. The white laminae are nearly pure calcite, whereas the light yellowish-green laminae contain an abundance of quartz, tremolite, and epidote. The quartz, calc-silicate-rich laminae are more resistant; differential weathering causes exposed surfaces to appear jagged and show irregular wavy banding. Some purer limestones in the section lack laminae, but contain small lenses, nodules or rounded segregations of quartz. The impure limestones are commonly rather intensely deformed by folding or shearing.
The calcareous, cherty metaquartzites of the Mescal are light gray to green, slabby, and thinly laminated to thinly bedded. The chert and very fine to fine-grained quartz and calcite are concentrated in alternating bands, and similarly tend to weather differentially. Many of the laminae show strong intralaminal deformation. Where exposed, the metaquartzites generally crop out in a cliff 20 to 30 feet high and seem to underlie the more massive impure marbles.

Amphibolites and Biotite-Hornblende Phyllites

The metamorphic equivalents of the Apache Group are interfingered and interfolded with amphibolites and dark green to black biotite-hornblende phyllites. The thicknesses of the dark phyllites and amphibolites range from a few inches to a maximum of several hundred feet. These dark rocks occur principally within the Dripping Spring metaquartzite; however, they are also present in the Pioneer Formation, along the borders and within the Catalina Gneiss, and beneath the rocks of the Cambrian System. Most of the thick units grade from amphibolite in the center to finer grained phyllites along the borders. The amphibolites generally show a remnant diabasic texture, and almost certainly represent metamorphosed diabase or lamprophyre sills. The diabasic texture is most pronounced in the centers of the
units and fades out gradually in the fine-grained phases along the borders. Conversely, foliation is better developed in the biotite-hornblende phyllites. Meso- and micro-crenulations are common in both rocks.

The mineralogic constituents of the phyllites listed in their approximate order of abundance are plagioclase, biotite, hornblende, quartz and lesser amounts of sericite, cordierite, epidote, chlorite and microcline. Mafics generally make up more than 50 percent of the assemblage. The amphibolites have a similar mineral assemblage except that hornblende is more abundant than biotite. The plagioclase in the amphibolites is both twinned and untwinned, and rarely shows faint zoning.

**Cambrian Rocks**

Rocks thought to be metamorphosed equivalents of the Bolsa Quartzite and Abrigo Formation occur in widespread outcrops in the Mount Lemmon portion of the area of investigation (Figure 3). Minor areas of exposure of these Cambrian rocks are also present in the Butterfly Peak vicinity, and on the north slopes of Mount Bigelow (Figure 2).

**Bolsa Metaquartzite**

The Bolsa Quartzite is of Middle Cambrian age, and was first named by Ransome (1904) for exposures near Bisbee, Arizona. In the study area, the Bolsa is especially well exposed along the highway below the Mount Lemmon Ski Lodge.
From the ski lodge, the metaquartzite can be traced along the north side of the ridge that extends eastward from the lodge, and thence through a discontinuous series of outcrops northward to Marble Peak. In the Marble Peak vicinity, as in the study area, the Bolsa is overlain by the Abrigo. More important, however, on Marble Peak the Abrigo is less metamorphosed than in the study area and is overlain by younger Paleozoic rocks. This combination of circumstances allows a more certain identification than is possible in the study area where the metamorphic grade is considerably higher and the younger Paleozoic rocks are missing.

The Bolsa is generally white to medium gray, but may be pink or light brown particularly in the bottom half of the section where it is slightly feldspathic. It is characteristically well bedded; bedding thicknesses range from less than one inch to nine feet, and average from two to four feet. Laminae of quartzitic phyllite are common between the beds. These quartzitic phyllite laminae tend to weather more rapidly than the metaquartzite layers and accentuate the bedded character of the unit. Cross-strata are common, especially in the thicker and coarser grained beds. Sorting in the Bolsa is poor, especially in comparison to the metaquartzites of the Dripping Spring Formation. Medium-sized grains predominate; however, fine and coarse grains are both concentrated in layers and interspersed in the
unit. Numerous lenticular metaconglomerate beds less than one inch to as much as four inches thick, and containing principally granules and small pebbles, are also present. Many of the metaconglomerates rest on erosional surfaces, and most are gradational into overlying metaquartzite.

Aside from the thin interstratified phyllites and some basal phyllitic metaquartzites, the Bolsa is generally rather pure, containing greater than 90 percent quartz. In hand specimen, muscovite and biotite seem to be the most common accessory minerals with lesser amounts of epidote, amphibole, garnet, hematite, and magnetite. Feldspars are present locally, but are largely limited to the lower half of the unit.

The quartz grains of the Bolsa tend to be recrystallized, tightly sutured, and the rock generally fractures conchoidally through the grains. Elongation or flattening of grains has locally accompanied recrystallization and normally results in the formation of a secondary foliation. This secondary foliation is common in the Bolsa where it is in contact with the Leatherwood Quartz Diorite or the Catalina Gneiss. Locally at the contact with the Leatherwood, the quartzite has a baked, almost hornfelsic appearance with no evidence of a secondary foliation or remnant bedding.

The thickness of the Bolsa metaquartzite within the area ranges from approximately 70 feet near Lemmon Rock Lookout to 237 feet along the road below the Mount Lemmon
Ski Lodge. Meaningful estimates of the true thickness are difficult to obtain because of structural complications and generally poor exposures. At the ski lodge, the Bolsa is cut by numerous normal faults that cause a duplication of an unknown amount of the section. At this location, moreover, the actual base of the Bolsa may not be exposed. Approximately 15 feet of phyllites and phyllitic metaquartzites underlie the typically massive metaquartzites and crop out in the driveway of the ski lodge. These phyllites and phyllitic metaquartzites are probably basal Bolsa units and may be equivalent to part of the sequence of maroon sandstones and siltstones underlying the Bolsa Quartzite in Peppersauce Canyon.

In his work in Peppersauce Canyon, Stoyanow (1936) considered the maroon rocks underlying typical Bolsa to be part of the Apache Group, but he left them unnamed. Creasey (1967) considered the uppermost 93 feet of the maroon rocks, including two feet of basal conglomerate, as part of the Bolsa Quartzite. Creasey noted that the basal conglomerate contained diabase clasts. He reasoned that the diabase clasts were probably derived from the erosion of Precambrian diabase intrusions and concluded that the rocks above the conglomerate bed should be included as part of the Bolsa Quartzite.
The correctness of this conclusion is not likely to be determined within the study area; however, the phyllites and phyllitic metaquartzites exposed beneath the Bolsa metaquartzites at the ski lodge appear to be gradational and conformable with the metaquartzites. Some of the phyllitic metaquartzites are conglomeratic and contain granules and pebbles of quartzite in a matrix of quartz, biotite, hornblende and plagioclase.

Dark green to black amphibolites and biotite-hornblende phyllites are exposed beneath the metasedimentary phyllites and phyllitic quartzites included here as basal Bolsa metaquartzite. The amphibolites and biotite-hornblende phyllites are extremely well foliated, but in thin section seem to exhibit residual igneous texture. The exact contact relationships of the metasediments and the dark amphibolites and biotite phyllites are not clear, but the two compositionally different rocks appear to be interfingered. This interfingering and the remnant igneous texture suggest an intrusive origin for the dark amphibolites and biotite phyllites. If the dark rocks were indeed intruded into the metasediments then the phyllitic quartzites and metasedimentary phyllites exposed below the ski lodge need not be the very basal Bolsa units.

On the west side of Mount Lemmon, the Bolsa is underlain by the Dripping Spring metaquartzite and is overlain by the Abrigo Formation. At this location, a minimum of 140
feet of Bolsa metaquartzite is exposed in steep cliffs, but the basal contact of the formation is covered. This section is also structurally complicated and contains abundant pegmatite and aplite intrusives.

Elsewhere throughout the Mount Lemmon area, the Bolsa is underlain by intrusive or gneissose rocks, and part of the section seems to be missing. In the deep ravine north of the ski lodge, for instance, the Bolsa overlies and is intruded by the Leatherwood Quartz Diorite. A similar contact relationship is present in Carter Canyon. Near Lemmon Rock Lookout, the Bolsa is intruded by pegmatites from the Catalina Gneiss.

In the Bear Wallow-Mount Bigelow area, the Bolsa metaquartzite overlies, but is separated from the Mescal Limestone by amphibolite.

Abrigo Formation

Ransome (1904) applied the term Abrigo limestone to rocks of Middle and Late Cambrian age exposed on Mount Martin near Bisbee, Arizona. Stoyanow (1936) correlated the Middle and Upper Cambrian rocks of Peppersauce Canyon with the type section, but subdivided the unit and introduced new terminology. To replace the term Abrigo limestone, Stoyanow proposed four new formations. In ascending order they are: the Santa Catalina formation, the Southern Belle quartzite, the Abrigo formation and the Peppersauce sandstone.
Creasey (1967), in his work in the Mammoth quadrangle, redivided the Abrigo section in Peppersauce Canyon into three members: a basal Three C Member, equal to Stoyanow's Santa Catalina formation; a middle Southern Belle Member; and an upper Peppersauce Member that includes Stoyanow's Abrigo formation and Peppersauce sandstone. Creasey notes that the term Santa Catalina was preempted by Blake (1908), and that the name Three C was chosen to replace the occupied term.

In the study area, rocks thought to be metamorphic equivalents of the Abrigo Formation have been subdivided into five informal members. The divisions are based upon general field appearance and examination of the rocks in hand specimen and thin section. In ascending order the units are as follows: lower biotite-quartz phyllite member; calc-silicate member; "sand-pea" metaquartzite member; upper biotite-quartz phyllite member; and white metaquartzite member.

The lower and upper biotite-quartz phyllite members are composed of biotite-rich metaquartzites and quartz phyllites. Phyllites predominate in the members and grain sizes are generally fine; however, coarse-grained and conglomeratic metaquartzite stringers are not uncommon, particularly in the upper member (Figure 6). Chlorite and epidote are generally present and lend a light green cast to the otherwise gray rock. In fresh specimen, these rocks are remarkably tough, but become friable upon weathering. Compared to the other
Figure 6. Upper biotite-quartz phyllite member of the Abrigo Formation

Similar fold and metaconglomerate stringer in the Abrigo Formation exposed along the highway between the Mount Lemmon Ski Lodge and Air Force Station. Facing N 50° W.
members, the biotite metaquartzites and biotite-quartz phyllites weather easily and generally form slopes.

The contact of the lower biotite-quartz phyllite member with the underlying Bolsa is gradational and is arbitrarily placed at the top of the uppermost massive white to light brown metaquartzite. Above this, thin bedded metaquartzites and phyllitic metaquartzites are intercalated with biotite-quartz phyllites. The upper and lower biotite-quartz phyllite members are each 25 to 35 feet thick.

The calc-silicate member overlies and is gradational with the lower biotite-quartz phyllite member. The calc-silicate is generally massive, very thinly laminated and contains discontinuous dark and light green bands. The bands are commonly less than one-half inch thick, and their color reflects relative amounts of actinolitic hornblende and epidote associated with quartz, calcite, diopside, idocrase and garnet. Although generally massive on fresh surfaces, the member develops wavy bands and lenticular vugs through differential weathering. These wavy bands and lenticular vugs are characteristic of the unit in outcrop. The member also contains several lenses of white to yellowish-green calc-silicate. These lenses are several inches to approximately 15 feet thick, and although the lenses are quartz- and tremolite-rich in places, they may be composed of essentially pure wollastonite and euhedral idocrase. The calc-silicate
member is mutually intertongued with the "sand-pea" metaquartzite member.

"Sand-pea" metaquartzite is a field term adopted because of the peculiar appearance of the rock especially on weathered surfaces. The metaquartzite is generally massive, laminated, and mottled light brown and light gray to olive green. It characteristically weathers to an uneven surface that contains rounded pea-sized nodes of sand grains (Figure 7). Not uncommonly the "sand-peas" are accompanied by crystals of idocrase as large as one-half inch in length.

In thin section, the rock shows roughly circular aggregations of medium-grained idocrase, garnet, diopside and quartz set in a finer grained matrix of quartz and garnet. Within the circular aggregations the idocrase and garnet are generally euhedral to subhedral and surrounded by anhedral garnet and highly sutured quartz. Some of the garnet is anisotropic, commonly grading from isotropic in the center of the crystal to anisotropic near the borders.

Because of mutual interfingering of the "sand-pea" metaquartzite with the calc-silicate member, and because of structural complications, a meaningful thickness for either member is difficult to determine. However, 150 feet to 200 feet is a reasonable estimate for the aggregate thickness of both members. Both members, particularly the "sand-pea" metaquartzite near its top, have thinly interstratified
Figure 7. "Sand-pea" metaquartzite member of the Abrigo Formation

Note mottled color on fresh surface and pea-sized nodes which result from weathering.

Figure 8. Intense deformation in graphitic phyllite

Interbedded graphitic phyllite of the white metaquartzite member of the Abrigo formation exposed on the Mount Lemmon Air Force Station.
layers of sericite phyllite. These phyllites are commonly light gray and range from six inches to several feet in thickness.

In much of the area, the "sand-pea" metaquartzite is overlain by a thin metaconglomerate. The metaconglomerate bed ranges from a few inches to six feet in thickness, and at least locally rests disconformably on phyllites of the "sand-pea" member. In other places the basal contact of the metaconglomerate seems to be gradational with the underlying "sand-pea" metaquartzite.

The metaconglomerate is commonly composed of clasts of white metaquartzite and vein quartz in a matrix of sericite and medium- to coarse-grained quartz. Clasts range from granules to pebbles and generally show elongation or flattening. Quartz grains and sericite also tend to be strung out parallel to the long axes of the pebbles, but recrystallization is not as extensive as in the Barnes or Scanlan metaconglomerates. The metaconglomerate tends to break around rather than through the clasts as is ordinarily the case with the Precambrian metaconglomerates. This characteristic is probably due to the high percentage of sericite in the matrix as well as the lack of pervasive recrystallization. The upper contact of the metaconglomerate with the upper biotite-quartz phyllite member seems to be abrupt, but is conformable.
The white metaquartzite member is the uppermost member of the Abrigo recognized within the study area. The member is composed of two white metaquartzites interbedded with phyllites and phyllitic metaquartzites. The metaquartzites are each approximately 15 feet thick. The thickness of the phyllites and phyllitic metaquartzites within the member probably equals or exceeds this total; however, the white metaquartzites are distinctive and can be traced over much of the summit areas of Mount Lemmon and Loma Linda.

The white metaquartzites are massive to blocky, laminated, and although rather poorly sorted, contain comparatively few impurities. Sericite and fine-grained biotite are the most common contaminants, but local disseminated grains of hematite are also present. Weathering of the biotite and hematite causes the otherwise white rock to become heavily iron-stained. Fine grain sizes predominate, but interspersed medium and coarse grains are common. In places, lenticular beds or pods of granules and pebbles occur.

Deformation within the white metaquartzite member appears to be more intense than in the underlying more massive members of the Abrigo. This deformation is manifest principally in the phyllites interbedded within the member, but it is also recorded in the metaquartzites. Similar to the Dripping Spring and Bolsa metaquartzites, where deformation in the white metaquartzites is intense, a secondary foliation develops through recrystallization and stretching.
or flattening of grains. Unraveling the structures in these white Abrigo units, however, is particularly difficult because of the general lack of distinctive compositional layering.

The phyllites and phyllitic metaquartzites of the white metaquartzite member are generally light to dark gray and olive green. Locally, the phyllitic metaquartzites are conglomeratic and, like conglomerates elsewhere described, display stretched and flattened clasts. In general, the phyllites resemble those previously described in other sections with perhaps one noteworthy exception. Some of the phyllites in the upper member are graphitic; the result perhaps of metamorphism of carbonaceous sediments in the Abrigo. Where present, the graphitic phyllites are commonly intensely deformed (Figure 8, page 31).

Cretaceous(?)-Tertiary(?) Rocks

Leatherwood Quartz Diorite

The term Leatherwood Quartz Diorite was used by Bromfield (1952) to refer to the stock-like intrusive located on the northeast flank of the Catalina Mountains. According to Hanson (1966, page 15), the Leatherwood is typically medium to dark gray, medium-grained, massive to weakly foliated and ranges from granodiorite to quartz diorite in composition. Hanson estimates the average composition of the rock as plagioclase (42 percent), quartz (22 percent), biotite (19 percent), epidote (10 percent), microcline (5 percent),
hornblende (1 percent); plus traces of apatite, calcite, chlorite, fluorite, opaques, sericite, sphene, and zircon.

Along the northern boundary of the Mount Lemmon area, the Leatherwood Quartz Diorite intrudes rocks of the Cambrian System. This contact is poorly exposed, but it is marked nearly everywhere by a dark, fine-grained phyllite. In thin section, the phyllites have a distinctly cataclastic texture; however, some sections show a faint remnant igneous texture. Biotite, the most common constituent of the phyllite, is accompanied by lesser amounts of plagioclase, hornblende and quartz. The quartz may comprise as much as 15 percent of the rock in places.

The origin of the phyllites is uncertain. Perhaps they, like the phyllites and amphibolites that interfinger with the Apache Group, are metamorphosed diabase or lamprophyre intrusives. It may be, however, that these phyllites are a sheared fine-grained border phase of the Leatherwood Quartz Diorite. The latter interpretation gains some support from the fact that foliation is notably present in the borders of the Leatherwood "stock." The foliation is generally accompanied by fairly intense shearing, and the intensity of development of these structures increases as the contact is approached from within the quartz diorite pluton.

In the stream channel below Lemmon Spring, a rock that is tentatively identified as Leatherwood Quartz Diorite intrudes the Apache Group and the Cambrian rocks. Although
this exposure is more than 4,000 feet from the nearest known Leatherwood, there is little likelihood that the correlation is in error. The megascopic and microscopic composition and texture of the rocks are remarkably similar. At the Lemmon Spring location, the Leatherwood Quartz Diorite(?) forms the core of a northeast trending anticline. Foliation in the intrusive is fairly well developed and is generally conformable with the foliation in the adjacent wall rocks. Apophyses of Leatherwood Quartz Diorite(?) are common in the wall rocks immediately adjacent to the intrusive.

In the Bear Wallow area, Leatherwood Quartz Diorite intrudes biotite phyllites and amphibolites interlayered with the Dripping Spring metaquartzite. The contact is poorly exposed, but appears to be gradational. In general, the thinly foliated phyllites grade into more amphibole-rich phyllites and coarser grained amphibolites. Attendant to the increase in grain size, the plagioclase content seems to increase slightly. Pegmatite and aplite dikes, sills, and pods are very common in the biotite phyllites and amphibolites.

Catalina Granite

The Mt. Lemmon area is bounded along the north-northwest border by a granite which forms an impressive north-northwest trending spine- or fin-backed ridge named Reef of Rock. This granite was referred to by Moore and others (1949) as the Catalina Granite. The spines are
formed by differential weathering along the foliation, and along pronounced joints striking northwest and northeast.

Within the area of this study the Catalina Granite is light gray to light brown, medium-grained, and contains phenocrysts of orthoclase. The rock commonly displays one dominant foliation direction, but almost everywhere two additional less pronounced directions of planar mineral parallelism can be recognized. The dominant foliation generally strikes N 40° to 60° E, dips 40° to 60° southeasterly, and is accentuated by pervasive shearing parallel to the mineral parallelism. A second foliation, especially well marked by biotite, strikes N 70° to 85° W and dips 30° to 55° N. The third direction of mineral parallelism strikes northeast subparallel to the most pronounced foliation direction, but dips to the north. It is not really clear whether the foliation noted here as a third direction does indeed represent a separate element, or whether it reflects folding of one of the other two planar surfaces. Nowhere in the outcrops visited, could either of these possibilities be convincingly demonstrated. It is noteworthy, however, that the three directions of foliation can be recognized in outcrops one-half mile apart.

Within the area of study, the Catalina Granite is in contact only with the Abrigo Formation. The nature of the
contact is variable and is dependent principally upon the composition of the wall rocks adjacent to the granite. Where the granite is in juxtaposition with the calc-silicate or "sand-pea" members of the Abrigo Formation, the contact is clearly intrusive. In these areas, dikes and sills of granite are common in the wall rocks, and the contact between the rock types is distinct. In the stream channel below Lemmon Spring, the contact relationships indicate that the granite was quite mobile at the time of intrusion. At this location, large blocks of calc-silicate have been dislodged from the walls and "swept" along by the intrusive. The structures in the blocks are discordant with those in the wall rocks, and a poorly developed flow foliation in the granite forms swirls around the discordant blocks. In other places where the contact is clearly intrusive, little evidence of disruption of the wall rocks is present, and the intrusion seems to have been passive.

By contrast, where the Catalina Granite is in juxtaposition with the biotite-quartz phyllites of the Abrigo Formation, the contact is gradational (Figure 3). In these areas the transition from phyllite to granite takes place throughout a zone approximately 500 feet wide. As the transition zone is approached from exposures of the Abrigo Formation, the phyllites display more coarsely recrystallized textures and a subtle increase in quartz and feldspar content. Farther into the zone, the texture is schistose,
and quartz- and feldspar-rich bands are common. In the center of the transition zone, gneiss predominates, but inclusions of biotite phyllite and schist comprise approximately 30 percent of the rock. Within the transition zone immediately adjacent to the granite, the gneiss is noticeably more homogeneous, less well foliated, and contains fewer and generally smaller inclusions. Several hundred feet within the granite, the inclusions are present only as thin biotite clots or schlieren. The structures within the phyllite, the transition zone, and the granite are essentially concordant. The foliation in the phyllites of the Abrigo Formation can be traced into the gneiss, and the phyllite and schist inclusions in the granite are generally oriented parallel to the dominant foliation direction.

Catalina Gneiss

The Catalina Gneiss, first named by DuBois (1959), forms the southern boundary of the area of investigation. Along this boundary, the gneiss is generally granitic in composition, medium- to coarse-grained, well foliated, and contains abundant pegmatite. The contact of the gneiss with the Apache Group and rocks of the Cambrian System is irregular in detail; however, structures within the adjacent rock units are roughly conformable. In most places, the contact is distinct and clearly intrusive, and apophyses of Catalina Gneiss in the adjacent wall rocks are common.
Elsewhere, the contact relationships appear gradational and locally structures can be traced from the Apache Group into the gneiss. Gradational contacts are most common where pegmatite is abundant. In the vicinity of Lemmon Rock Lookout, for instance, a pegmatite "front" several hundred feet wide is present between the Catalina Gneiss and the Bolsa metaquartzite. Adjacent to its contact with the pegmatite, the Bolsa exhibits pronounced recrystallization and is uncommonly rich in feldspars. In some places, the feldspars comprise 25 to 30 percent of the rock, and both orthoclase and plagioclase are present. The feldspars are very fine-grained, pervade the matrix, and fill incipient fractures within the metaquartzite. Beyond the pegmatite front, scattered inclusions of metaquartzite, biotite metaquartzite, and biotite phyllite occur along the foliation in the gneiss. Large blocks of medium-grained, gray, biotite-rich gneiss are also included in the coarser grained host rock. The darker biotite-rich gneiss bears a remarkable resemblance to the well foliated phase of the Leatherwood Quartz Diorite. The only notable difference seems to be that the included blocks are richer in potassium feldspar. Stained samples of the biotite gneiss inclusions indicate that the rock contains approximately 20 percent potassium feldspar. Hanson (1966) estimates that the microcline content of the Leatherwood Quartz Diorite averages about five percent.
Along the Mount Lemmon Highway northwest of the entrance to Spencers Picnic Ground, the contact seems gradational. At this location, exposures of Catalina Gneiss and weathered phyllites of the Apache Group are separated by a zone of hybrid rock that has affinities with both end members. The zone of hybrid rock is approximately 150 feet wide and contains weathered biotite-feldspar-quartz schist intimately interfingered with phyllites of Apache Group aspect. The compositions of the weathered schist and the Catalina Gneiss are similar; however, the texture of the schist resembles the Apache Group phyllites. Chevron folds and micro-crenulations, common in the phyllites, are also found in the weathered schist. The foliation in the phyllite, schist and gneiss is roughly conformable; however, the schist also appears to be sheared in the plane of the foliation. The shearing connotes some movement within the transition zone to induce the cataclasis, but a fault of major displacement does not seem to be indicated. Approximately 20 feet beyond the transition zone a thin sill of Catalina Gneiss intrudes the white metaquartzites of the Dripping Spring Formation. Southeast of these exposures, below the summit of the ridge, the contact is locally gradational. In places, layers of biotite phyllite alternate with layers of medium-grained gneiss containing porphyroblasts of potassium feldspar and plagioclase.
Quartz Latite Porphyry

Along the northern edge of the area, in the vicinity of Mount Bigelow, the Apache Group and rocks of the Cambrian System are in intrusive contact with a metamorphosed porphyritic rock of roughly quartz latite composition. The quartz latite is light gray to light brown and contains abundant relict phenocrysts of plagioclase and quartz. The plagioclase phenocrysts are more common and form 30 to 40 percent of the rock. Quartz phenocrysts make up approximately 5 to 15 percent of the rock, and although less abundant, they are generally larger. The average diameter of the phenocrysts is approximately 2 mm, but crystals as large as 5 mm are not uncommon. The phenocrysts are generally rounded or lenticular, but many of the quartz crystals are euhedral. In thin section, the rock shows a distinct recrystallized cataclastic texture, and the phenocrysts appear broken and rounded. The plagioclase in the phenocrysts is generally twinned, and rarely shows a faint zoning. By contrast, twinning in the matrix plagioclase is rare.

The matrix of the quartz latite is composed principally of quartz, plagioclase, biotite and microcline, with minor muscovite, epidote and sericite. Although the matrix is generally fine-grained, the biotite is medium-grained and preferentially oriented. The oriented biotite imparts a pronounced schistose texture to the porphyritic rock. Warping
or bending of the biotite flakes around the relict phenocrysts is common. Locally, the quartz latite porphyry contains scattered inclusions of dark, fine-grained, biotite phyllite. The inclusions are commonly planar and lie within the plane of the foliation.

Rocks of strikingly similar description have been mapped by several other workers in the Catalina Mountains. Pilkington (1962), in his work to the east of the present area of study, mapped fairly large sill-like bodies of quartz latite porphyry. The rocks described here appear to be the westward extensions of those bodies. Hanson (1966) considered the portion of this schistose porphyry exposed within his area of study as an earlier minor intrusion genetically related to the Leatherwood Quartz Diorite. He proposed that the cataclastic deformation noted was caused by the emplacement of the later Leatherwood intrusion. Unfortunately, Hanson makes neither reference to the quartz latite porphyry described by Pilkington, nor does he allude to a possible correlation between the two units. In addition to these writers, Moore and others (1949), Wallace (1955), Peirce (1958), and Creasey (1967) all make reference to intrusive rocks of similar composition and texture.
STRUCTURE

Description of the Structural Elements

Planar Elements

As noted in the section on rock description, most of the Apache Group and Cambrian rocks have retained recognizable sedimentary bedding and laminae. These inherited fabric elements are the earliest formed planar features evident within the rocks and are here designated $S_1$. $S_1$ is most easily recognized where compositional or lithologic layering is present. One must use care, however, not to confuse surfaces of original compositional differences with transposed foliation (Turner and Weiss, 1963). In most places, the imposed foliation shares the attitude of the remnant structure resulting in a composite fabric. Where the composite fabric is present, the remnant structures are intensified. The imposed foliation results from the preferred orientation of recrystallized minerals and is designated $S_2$. $S_2$ is generally parallel to the limbs of "isoclinal" similar folds folding $S_1$, but passes undeformed through the hinges of the folds. Thus $S_1$ and $S_2$ are generally parallel, and $S_2$ is axial plane foliation or cleavage.
Locally, \( S_2 \) is kink folded resulting in a third foliation parallel to the axial planes of the kink folds. This axial surface is designated \( S_3 \).

Linear Elements

In this study, lineation is restricted to linear parallelism penetrative in hand specimen. Within the rocks of the area, lineation thus defined is present as stretched and elongated mineral grains, intersecting \( S \) surfaces, and as microcrenulations or crinkles on \( S \) surfaces. In most places it is not possible to distinguish between lineation that results from true elongation or stretching of mineral grains and lineation that results from the intersection of \( S_1 \) and \( S_2 \). The difficulty arises because \( S_2 \) is formed by recrystallization with attendant flattening and elongation of minerals. Thus, where \( S_2 \) intersects \( S_1 \), an apparent stretched mineral lineation results. The distinction is especially difficult where \( S_2 \) is parallel, or nearly parallel to \( S_1 \). Therefore, in this report, the lineation resulting from true mineral elongation, and that lineation resulting from the intersection of \( S_1 \) and \( S_2 \) are referred to simply as stretched mineral lineation.

Microcrenulations or crinkle folds are also common in the rocks of the area. The lineation formed by the crests and troughs of these microcrenulations on \( S_2 \) is distinguished from the stretched mineral lineation in this study.
Cleavage mullions were noted in several places within the area. The mullions are present in the hinges of folds in $S_1$, and are formed by slippage of $S_1$ layering along axial plane foliation ($S_2$) (Figures 9 and 10). The mullions are generally lenticular, but some have a poorly developed sigmoidal shape. The sigmoidal shape is probably caused by refraction of the foliation or cleavage as it passes through the sedimentary layering ($S_1$) or by slippage of $S_1$ along $S_2$.

Stretched and flattened clasts within metaconglomerates and conglomeratic metaquartzites are common in the area. As described earlier, the stretching and flattening appear to result from both recrystallization and cataclasis. Most clasts show a pervasive stretched mineral lineation parallel to their long axes. The extent of clast deformation is variable, and appears to be related to the intensity of the metamorphism, the availability of fluids, and the intensity of folding in the locality.

The planar and linear structural elements within the Leatherwood Quartz Diorite, the Catalina Granite, the Cataline Gneiss and the quartz latite porphyry were described in the descriptive section on these rocks and will not be discussed further.

**Description of the Folds**

Folds in the area of investigation range from a fraction of an inch to several hundred feet in amplitude and wave
Figure 9. "Isoclinal" similar fold \( (F_1) \)

Photograph and sketch of an \( F_1 \) fold in the Dripping Spring metaquartzite south of Butterfly Peak. Lithologic layering \( (S_1) \) is folded by \( F_1 \), and the axial plane foliation \( (S_2) \) passes through the hinge of the fold essentially undeformed. Facing S 60° E. (See close-up below)

Figure 10. Cleavage mullions

Close-up of the upper limb of the fold shown in Figure 9 showing sigmoidal cleavage mullions of white metaquartzite \( (S_1) \) in more phyllitic layers of the Dripping Spring. Alignment of mullions delineates \( S_1 \) subparallel to surface upon which Brunton Compass rests. \( S_2 \) is horizontal but becomes refracted slightly on passing through \( S_1 \).
Figure 9. "Isoclinal" similar fold ($F_1$)

Figure 10. Cleavage mullions
length. Most are nonplane noncylindrical, but some are nonplane cylindrical and plane cylindrical (Turner and Weiss, 1963). The folds are generally asymmetrical; however, symmetrical forms are present on both large and small scales. Orientations are varied, recumbent and overturned folds with axial planes dipping less than 30 degrees predominate. Upright to slightly overturned orientations are present, but are largely restricted to broad open folds. A wide variety of fold styles is found in the area, and individual folds may be of a single style or composed of a combination of two or more styles.

Styles and Mechanisms of Formation

"Isoclinal" Similar Folds. "Isoclinal" similar folds (Turner and Weiss, 1963, page 115) are the most common, and most of the rocks show some evidence of pervasive folding of this style (Figures 11 and 12). The evidence may be manifest as fold hinges, or simply as convergent foliation and an essentially penetrative lineation. The evidence for this style of folding is less pronounced in the metaquartzites than in the phyllitic metaquartzites, quartzitic phyllites and phyllites. This relative paucity of evidence reflects the fact that compositional or lithologic layering is generally poorly developed in the metaquartzites, but it also seems to indicate a relatively lower intensity of deformation within the metaquartzites. In general, the metaquartzite
Figure 11. "Isoclinal" similar folds

Tightly folded strata of the upper member(?) of the Dripping Spring metaquartzite along the main road below the loop in Bear Wallow. Facing east.
Figure 12. Tightly appressed "isoclinal" similar folds

"Isoclinal" similar folds in the upper member(?) of the Dripping Spring metaquartzite along gravel road below the observatory. Facing northwest.
layers have acted as strong, less ductile layers in comparison to the phyllitic strata.

In "isoclinal" similar folds involving intercalated phyllitic and metaquartzitic layers, the phyllitic layers thicken more in the hinge of the fold than the quartz-rich layers. Most of the folds display an axial plane foliation ($S_2$) that passes undeformed through the hinge. The foliation seems to be pervasive, but in some layers, especially those rich in quartz, the foliation cannot be traced through the fold with any degree of certainty. Many of the folds exhibit small thrust faults of minor displacement. The thrust faults commonly dip at a slight angle to $S_2$ and pass through the fold hinge just outside the point of maximum curvature. The relative thickening in the fold hinge of the phyllite-rich layers as compared to the more quartz-rich strata is significant. It indicates differential rates of flowage of material from the limbs to the hinge of the structures during deformation. This differential flowage reflects the degree of ductility of the deformed materials and indicates greater ductility in the more phyllitic rocks. If, as it appears, the various rock layers acted with differential ductility, then the boundaries between the layers are effectively velocity discontinuities. This probably means that slippage as well as flowage was an important mechanism along this discontinuity. This conclusion is strengthened by a stretched mineral lineation which is present on the limbs of
some of the "isoclinal" similar folds. The lineation is developed parallel to the "a," or tectonic transport direction.

In most of the "isoclinal" similar folds, flowage of material appears to have been largely confined within the flexed layers. Mechanics of flowage and slippage between the flexed layers have been termed flexural flow and flexural slip by Donath and Parker (1964). The following chart (Figure 13) reproduced from Donath and Parker is especially useful in considering the behavior of the materials deformed within the folds. The authors develop a classification of folds based upon the mechanisms that they believe produce folds. In the chart, the fields of folding are shown according to the authors' classification and are related to ductility contrast and mean ductility of the deformed rocks.

Most of the pervasive "isoclinal" similar folds in the interbedded phyllitic rocks of the area would be termed flexural flow folds according to the classification of Donath and Parker (1964). As mentioned above in the description of the folds, axial plane foliation is pervasive in most of the structures. According to Donath and Parker, cleavage may develop in flexural flow folding where there is high ductility contrast and appreciable flow in the hinge. This explanation probably partially accounts for the axial plane foliation ($S_2$) noted; however, in some of the folds displacement of cleavage mullions indicates passive movement along $S_2$ (Figures 9 and 10). This evidence for passive folding need not constitute a contradiction to the conclusion that flexural
Figure 13. Fields of folding related to mean ductility and ductility contrast

Source: Donath and Parker (1964)
folding was of principal importance in the interbedded rocks but probably reflects conditions of a transition zone between flexural flow and passive folding. This would be especially true where the rocks have a moderate to low ductility contrast (Figure 13). Indeed, many of the more homogeneous rocks, including some of the metaquartzites, contain evidence of passive folding, and it is in these rocks that low to moderate ductility contrasts might be expected.

The fine-grained white to light brown Dripping Spring metaquartzite cropping out on Mount Bigelow, and along the gravel road southeast of the observatory, displays a very pronounced foliation. Locally, especially near the observatory, subtle sedimentary laminae are also detectable. The laminae intersect the foliation at a wide range of angles and delineate a number of "isoclinal" similar folds in the laminae \( S_1 \). In some places the laminae are marked by curviplanar surfaces. Care must be used in selection of these surfaces, however, as many of the joints in the area are also curviplanar.

On Mount Bigelow, "isoclinal" similar folds in the same metaquartzite are more difficult to recognize, but an abundance of quartz veins or segregations is present. The quartz veins form intricate patterns in the outcrops and intersect the foliation at a wide range of angles. The vein quartz is recrystallized, and the apparent long axes of the quartz crystals lie within the foliation no matter what the
attitude of the vein. Additionally, incipient displacement of the quartz veins along the foliation is revealed in some veins. This evidence warrants the suggestion that flow and slip across the sedimentary laminae and along \(S_2\) (passive folding) were important mechanisms in the deformation of these metaquartzites. The suggestion is strengthened by inspection of Figure 13. The reader will note that where ductility contrast is low, the field of passive folding extends slightly into the moderate range of mean ductility. Owing to the relatively more homogeneous composition of the metaquartzites compared to the more phyllitic rocks, it seems reasonable to assume that a low ductility contrast may have existed in the metaquartzites during deformation.

Some of the very thinly laminated cherty phyllites in the upper member of the Dripping Spring metaquartzite display "isoclinal" similar folds that clearly indicate passive flowage. The phyllites contain interstratified laminae rich in chert, feldspar, and hornblende. Color differences in the laminae permit tracing of delicate flame-like structures in the phyllites. Foliation is moderately well developed and passes through the hinge of the folds. Not uncommonly, however, the foliation is deformed.

Flowage appears to have been principally passive; that is, along the foliation \((S_2)\) and across the laminae \((S_1)\). Locally, flowage seems to have occurred across the foliation and caused active deformation of both \(S_1\) and \(S_2\).
Differential flowage parallel to the axial surface has also been effective in formation of these folds. Most of the folds show remarkable changes in configuration in cross sections cut at closely spaced intervals (Figure 14).

**Knee-"Isoclinal" Similar Folds.** "Isoclinal" similar folds are found in combination with knee folds (Hills, 1963, page 342). This combination occurs over a broad range of scales and wavelengths and amplitudes range from several inches to over 100 feet (Figure 15). The combined forms are most common in interbedded metaquartzites and phyllitic rocks where a considerable range of ductility contrast probably existed during deformation. The metaquartzites generally form the knee fold portion of the combination and seem to have deformed principally by slippage between the flexed layers. In contrast, the phyllitic rocks form the "isoclinal" similar portion of the overall fold, and appear to have formed by flowage between the flexed layers (flexural flow).

**Concentric Folds.** Concentric folds are abundant within the study area (Figure 16). They occur as individual folds, but are more common in simple combination with "isoclinal" similar folds, or in a more complex combination of styles in disharmonic folds. Like knee folds, they are generally found in interbedded metaquartzites and phyllitic rocks and are normally confined to the metaquartzite layers.
Hand specimen sized folds in the very thinly laminated cherty phyllites of the upper member(?) of the Dripping Spring metaquartzite. Distance from front of specimen (left) to back (right) is approximately one inch.

(Photographs by Norman E. Lehman)
Figure 15. Knee-"isoclinal" similar fold

Combined fold styles in thinly interbedded phyllites and metaquartzites of the Dripping Spring Formation along the main road in loop at Bear Wallow. Facing east. Sketch accentuates knee and "isoclinal" similar portions of fold.
Concentric folds within a larger disharmonic fold in the upper member(?) of the Dripping Spring metaquartzite along the gravel road above the Upper Bear Wallow Campground.

Intense folding in the Dripping Spring metaquartzite immediately overlying metadiabase(?) in Sabino Canyon. Facing northeast. Trend of fold axis is N 80° E, 15°. Upper layers have moved south-southeast relative to lower layers.
Slip between flexed layers (flexural slip) is the principal mechanism of formation of the concentric folds.

**Disharmonic Folds.** Disharmonic fold styles on a mesoscopic scale are especially common in the area. Like the more regular knee-"isoclinal" similar combined forms, the disharmonic folds occur where a contrast in lithology is present. Flowage and slippage between the flexed layers are again important, however, active flowage of material across the layering plays a critical role in development of disharmonic folds. This active flowage along and across ruptured layers that have deformed by flexural slip results in extensive redistribution of material. As a result, the geometry in the hinge of the fold commonly changes rather abruptly from layer to layer (Figure 17, page 59). These features reflect a high ductility contrast in the deformed rocks. Donath and Parker (1964) have termed such deformation quasi-flexural folding. Figure 18 is a sketch of disharmonic, overturned and recumbent folds in interbedded metaquartzites and phyllites of the upper member of the Dripping Spring metaquartzite.

**Chevron Folds.** Chevron folds or kink folds are common in the phyllites and amphibolites of the area. The folds range from microcrenulations to chevrons several inches in wave length and amplitude. This style of fold is especially abundant where phyllites are in contact with the Catalina
Figure 18. Disharmonic, overturned and recumbent folds

Sketch of folds in interbedded metaquartzite and phyllite of the upper member of the Dripping Spring metaquartzite along the gravel road above the Upper Bear Wallow Campground. Average trend and plunge of folds is approximately S 75° W, 3°. Facing N 70° E.
Gneiss. In these places the chevrons are pervasive, and fold S$_2$. The chevrons are generally asymmetrical away from the gneiss; however, reversals in the direction of asymmetry and symmetrical folds are by no means lacking.

Chevron folds are also abundant in the dark biotite phyllites and amphibolites thought to be matamorphosed diabase or lamprophyre sills. In thin section, the kink folded phyllites display extremely well developed biotite or muscovite laminae alternating with laminae rich in quartz and plagioclase. The quartz and plagioclase, in contrast to the coarsely crystalline muscovite and biotite, are generally very fine grained and cataclastically deformed. Flexural slip or bend gliding appears to have been the dominant mechanism effective in formation of the chevron folds.

**Intrafolial Folds.** Intrafolial folds (Turner and Weiss, 1963, page 116) occur in sufficient numbers throughout the area to warrant mention here. They are most common in thinly laminated rocks, particularly in the Dripping Spring Formation, and the calcareous, cherty metaquartzites of the Mescal Limestone. The intrafolial folds are generally several inches in amplitude and wave length, and are sparsely scattered throughout the rock.

They appear to result from bedding slip which causes folding in the strata overlying the slip surface (Figure 19). Folding usually dies out within a few inches above the slip
Figure 19. Knee-intrafolial fold

F$_2$ folds in the Dripping Spring metaquartzite exposed above Butterfly Peak. Trend of the fold axis is S 80° E, 11° upper layers have moved south-southeast relative to the lower layers. Facing east.

Figure 20. Ptygmatic folds of pegmatite

Ptygmatic folds in biotite-hornblende phyllite exposed in bank of Mount Lemmon Ski Lodge parking lot. Relative movement of upper layers is northeastward. Facing N 45° W.
surface. Flexural slip and flexural flow are both important mechanisms in the formation of intrafolial folds.

**Ptygmatie Folds.** Ptygmatie folds of pegmatite and aplite are common in the dark biotite phyllites and amphibolites in the area. They are found in greatest abundance adjacent to the Catalina Gneiss and the Leatherwood Quartz Diorite. The ptygmatie folds are both continuous and discontinuous, and typically display thinning in the limbs and thickening in the hinge of the structures.

Figure 20 (page 63) shows ptygmatie folds in a pegmatite vein within dark, biotite-hornblende phyllite. In this example at least, the folds seem to have been formed by differential slippage (passive slip) along the foliation. The fact that foliation is not developed in the pegmatite vein probably indicates that the vein material was very ductile during deformation.

**Relations of the Structures in Space and Time**

Folds and Linear Elements

In the preceding section, the geometry and mechanisms of formation of the structural elements have been described. This investigation has revealed that these structures are not all of the same age, but that some predate others in formation. In this section, the structures will be discussed with respect to their formation in space and time. For ease of
presentation, the sequence of development of the folds and linear elements in the Bear Wallow-Mount Bigelow area and the Mount Lemmon area are discussed separately.

Mount Bigelow-Bear Wallow Area. The earliest formed folds are designated F₁. Folds of the F₁ system fold the sedimentary layering (S₁), and are synmetamorphic. Foliation developed during this metamorphism appears to be essentially parallel to the axial planes of the folds in S₁ and is designated S₂. S₂ therefore passes essentially undeformed through the hinges of the folds of the F₁ system. The style of F₁ is commonly "isoclinal" similar, but knee- and concentric-"isoclinal" similar folds are abundant (Figure 16).

Folds of the F₁ system seem to be pervasive in the more phylilitic rocks of the Apache Group. They are less well developed in the metaquartzites, and in the rocks of the Cambrian System, particularly in the Bolsa metaquartzite. Where recognized, F₁ are commonly small and range from an inch to several feet in amplitude (Figures 11 and 12). This system of folds is present on a larger scale however, as is evident from Figure 9. The larger scale F₁ are generally not recognized because of the limited size of the outcrops in the area, and the inaccessibility of some of the larger exposures. There is also a strong probability that some of the larger F₁ have been mistakenly identified as F₂. Unfortunately, positive identification of F₁ can only be made where
it can be demonstrated that $S_2$ passes through the hinge of the fold relatively undeformed. No other feature is indicative of the $F_1$ system.

In the Mount Bigelow-Bear Wallow area, folds of the $F_1$ system generally trend northeast or east-southeast, but reversals of plunge are common. Asymmetrical forms seem to be the most abundant, and where observed, the folds are generally recumbent or reclined with the axial planes dipping less than 20 degrees. Obviously, however, since the dip of the beds within the area ranges from 0 to 90 degrees, upright and overturned, folds with steeply inclined and overturned axial planes are present.

Folds that deform both the foliation ($S_2$) and the sedimentary layering ($S_1$) are designated $F_2$. The folds of the $F_2$ system postdate the metamorphic event which generated the $F_1$ system, but they may, in part at least, have been formed during the waning stages of the event. $F_2$ are generally larger than the recognized $F_1$ and range from a few inches to several hundred feet in amplitude (Figure 4). They are commonly the most obvious folds in the area. Knee and knee-"isoclinal" similar folds are the most common styles in the $F_2$ system. Outcrops that reveal $F_2$ folding $F_1$ are rare. Fortunately, one such exposure is present along the Mount Lemmon Highway in the loop at Bear Wallow (Figure 21). Asymmetrical forms with a reclined, recumbent, or overturned orientation are the most common.
Figure 21. Superposed folds

"Isoclinal" similarly folded (F₁) strata of the Dripping Spring metaquartzite, refolded by an F₂ knee fold exposed along the main road in the loop at Bear Wallow. Sketch delineates tight isoclinal fold hinges labeled F₁. Note flat faults indicating southward movement of the upper sheets. Facing east.
The trend of F_2 is somewhat variable and ranges from a few degrees south of west to north-northwest. The pattern of trends is fairly regular, however, (Figure 22, in pocket) and changes from N 70° W north of Mount Kellogg to N 45° to 85° W north of Mount Bigelow, and then from N 35° W to N 70° W in the vicinity of Turkey Flat. North and west of Turkey Flat, the regularity of the trend is broken, but in Bear Wallow, the main trend of F_2 turns west-southwest, or east-northeast, and continues to Spencer Peak. The regularity of the trend is again disrupted in Bear Wallow by local perturbations. North and northwest of Bear Wallow, in the vicinities of Butterfly Peak and Soldier Camp, F_2 continue their northwestward trend.

Throughout much of the area, F_1 and F_2 are so closely similar in trend and style they are difficult to distinguish in outcrop. In Bear Wallow, for instance, the fold systems are essentially homoaxial (Figure 21). Also, as noted earlier, the foliation in many of the quartzitic layers is poorly developed, and it is frequently difficult, on a mesoscopic scale at least, to determine whether the foliation is folded by the fold.

In the vicinity of Spencer Peak and Bear Wallow, folds thought to belong to a third system are recognized. These folds are commonly broad and open in style, upright, and symmetrical to slightly asymmetrical. The trend of the folds is generally a few degrees on either side of south.
Like F2, these structures fold S1 and S2, but theoretically would deform the real or imaginary S3 elements generated by the F2 event. Unfortunately, folds definitely superposed on the F2 system have not been observed, however, outcrops containing three axial directions have been found. One such outcrop is exposed behind the southeasternmost cabin on the south bank of Bear Wallow Creek. In this outcrop small "isoclinal" similar folds are refolded by an open style, asymmetric fold plunging 3°, S 5° E. The "isoclinal" similar folds are abundant in the outcrop. In the flank of the cross-fold the "isoclinal" similar folds plunge 35°, N 70° E, and higher in the sequence away from the immediate influence of the cross-fold, the "isoclinal" similar folds plunge 8°, N 40° E. In the same outcrop, but poorly exposed, two knee folds plunging 35°, S 50° E were noted.

The axis of the southward plunging open fold is marked by microcrenulations, and the east flank of the fold contains a stretched mineral lineation that plunges 33°, N 45° E. The interpretation, made here, is that the northeast plunging folds are F1, and the stretched mineral lineation is a "b" lineation to that generation of folds. The open fold is interpreted as F3 and the microcrenulations are a "b" lineation to that generation of folds. The poorly exposed knee folds then, on the basis of the style and the southeast trend, are tentatively assigned to the F2 system. There is some concern on the part of the writer at this
point in the study that what appears to be a third set of
folds may, in fact, be a different style, trend, and ori­
entation of \( F_2 \). However, on the basis of the multiple oc­
currences of outcrops displaying three axial directions, and
because of what seem to be important differences in the
style, symmetry, orientation, and trend, this fold system
is designated \( F_3 \).

The stretched mineral lineation within the area can­
not be assigned to a single time of formation, but appears
to have formed during several stages of deformation. As
noted earlier, stretched mineral lineation is formed by
either the intersection of \( S_1 \) and \( S_2 \), or by true or "sim­
ple" mineral elongation or growth. That lineation formed
by the intersection of \( S_1 \) and \( S_2 \) is a "b" lineation gener­
ated by \( F_1 \) folding and may be termed \( L_1 \).

The stretched mineral lineation that is formed by
"simple" mineral elongation, on the other hand, may be gen­
erated by any of the three fold events. Also, where formed
by "simple" mineral stretching, the lineation may be paral­
lel to either the "a" or "b" tectonic axis. On the north­
west side of Spencer Peak, just below the summit, stretched
mineral lineation is present in the "a" direction in one
"isoclinal" similar fold and in the "b" direction in an­
other fold of like style, orientation, and trend. These
two folds are only 10 feet apart and involve the same lay­
ers of the Dripping Spring metaquartzite.
In Bear Wallow, folds of the $F_2$ system containing stretched mineral lineation in both the "a" and "b" tectonic transport direction were observed. The "a" lineation is formed in the limbs, and the "b" lineation is formed in the hinge of the structure. Stretched mineral lineation with this dual tectonic orientation in $F_2$ folds seems to be present where there is a scarcity of platy minerals. The explanation for this dual orientation may be, that where biotite and muscovite are effectively absent the movement during folding would not be as concentrated along the foliation, and the normal microcrenulations would be less likely to form. By contrast, the "a" lineation observed may result because slippage and mineral growth are more evenly distributed throughout the flexed layers. The "b" lineation in the same fold probably forms by arcuation in the hinge. Therefore, although most of the stretched mineral lineation is probably a "b" lineation related to the $F_1$ fold system, such a relationship cannot be automatically assumed.

In some places folded lineations are revealed, and the identification of the fold system can offer clues to the time of generation of the lineation. In a few scattered outcrops just above the Bigelow Lookout Road near the turnoff from the Mount Lemmon Highway, the Catalina Gneiss is interlayered with dark, biotite phyllites of the Apache Group. In these exposures, both rocks show a pronounced
stretched mineral lineation plunging generally northeastward. One exposure shows the stretched mineral lineation folded by a northwest plunging fold. The fold, although poorly exposed, appears to be similar in style and overturned to the southwest. On the exposed hinge of the fold, the lineation reverses its plunge from the northeast to the southwest but undoubtedly resumes a north-northeast plunge on the overturned limb. The fold clearly folds the foliation in the gneiss and biotite phyllite, and belongs to the Fg system. Since the lineation predates the fold, it is designated L1.

In the Mount Bigelow area, microcrenulations subparallel to F2 are common. These microcrenulations are believed to have been generated by puckering and crinkling of the foliation that is folded by F2 and are accordingly designated L2. Where microcrenulations and stretched mineral lineations are present in the same exposures, the angle between their trends ranges from a few degrees to 120°. Because of the similarities of folds of the F1 and F2 systems, in areas where both the stretched mineral lineation and microcrenulations are present, the microcrenulations have been relied upon to indicate the trends of the F2 folds.

Microcrenulations also occur as a "b" lineation subparallel to F3 in the Spencer Peak and Bear Wallow areas. These structures do not differ in character from L2 and can only be identified by their association with F3. The
intersection of \( L_2 \) and \( L_3 \) should form interference microcrenulations much like interference micro-ripplemarks. Unfortunately, nowhere within the area were such features identified with certainty.

The trends of the long axes of 13 stretched and flattened clasts were measured in the Barnes and Scanlan metaglomerates on Mount Bigelow and Mount Kellogg. The trends ranged from N 25° W to N 65° W with the greatest frequency near N 55° W. The surfaces of the clasts are marked by a stretched mineral lineation which is parallel to the long axes of the clasts. The orientation of these linear elements is also subparallel to microcrenulations (\( L_2 \)) in the Pioneer Formation. Because of the subparallelism of the clasts, stretched mineral lineation, microcrenulations, and \( F_2 \) folds, it is concluded that the clast elongation and orientation was accomplished by the event which generated the \( F_2 \) system of folds.

**Mount Lemmon Area.** In the Mount Lemmon area intense folding is most common in the phyllitic rocks, especially in the biotite-quartz phyllite members of the Abrigo Formation and in the phyllites underly the metaquartzites of the Bolsa Formation. In general, the Bolsa appears to have behaved competently during deformation, and intense folding within the unit seems to be relatively local.
The relationships of the fold systems to each other, and to the linear structural elements in this part of the area, are more cryptic than in the Bear Wallow-Mount Bigelow portion. Again, several fold systems are indicated, but in the Mount Lemmon area evidence is more indirect. Only one exposure was found in which two systems of folds were present within a single outcrop. This exposure is just north-east of the Lemmon Rock Lookout.

As in the Mount Bigelow-Bear Wallow area, F₁ is again used to designate folds that belong to the earliest known fold system. F₁ folds the sedimentary layering (S₁), but S₂ passes through the hinge of the fold relatively undeformed. Folds of the F₁ system are generally "isoclinal" similar in style, asymmetrical and overturned, recumbent or reclined. The axial planes commonly dip gently northeast, and the folds plunge a few degrees northwest or southeast.

A second system of folds, with trends subparallel to the F₁ system, folds S₂ and S₁. Distinguishing between folds of the F₁ and F₂ systems is especially difficult in this area because S₂ is only faintly developed in many of the rocks. The trends of F₂, although subparallel to F₁, are more scattered and commonly range from north to west, but folds plunging to the southeast are also abundant. Folds of the F₂ system are generally concentric or knee in style, asymmetric and upright to overturned to the southwest. Chevron folds also are common in the F₂ system.
In addition to F₁ and F₂, a system of northeast trending folds is evident in the Mount Lemmon area. The folds of this system occur as upright, broad gentle open warps, and as upright to slightly overturned tight forms with steeply dipping limbs. The variation in style probably results from exposure of the folds at different structural levels. The folds are commonly asymmetric, but the direction of asymmetry is variable and seems to be related to location.

The position of this system of folds in the sequence of fold generations is not certain, but most of the evidence available indicates that they postdate those systems designated F₁ and F₂. In outcrops of the Bolsa metaquartzite on the northwest side of Mount Lemmon, "isoclinal" similar folds of the F₁ system appear to plunge down the dip of the southeast limb of a large northeast trending anticline. In the same outcrops, secondary foliation in the limb of a northeast trending fold is rotated from a dip of 45° to the horizontal. The foliation does not appear to be related to the axial plane of the northeast trending fold and is assumed to belong to an event that predates the northeast flexure. Elsewhere throughout the area, this worker has received the impression that the folds of the F₁ and F₂ systems may be "draped" across this orthogonally trending system. The folds of the northeast trending system are herein designated F₃.
The \( F_3 \) folds range in amplitude from a few inches to several hundred feet. Most of the tighter folds observed in the field fall within the lower portion of the size range given. Approximately 50 percent of the fold trends shown on Figure 23 (in pocket) are observed folds; that is, the axes could be located and measured in the field. The remaining 50 percent are approximate fold trends established on the basis of opposing dips. The fact that a proportionately larger number of the axes of the folds belonging to the \( F_3 \) system could not be located, expresses their broad, open style.

Stretched mineral lineation in the Mount Lemmon area, as in the Mount Bigelow-Bear Wallow area does not appear to be generated by a single fold event, but bears genetic relations to at least two generations of folds. In some folds of the \( F_1 \) system, the stretched mineral lineation is parallel to the "b" tectonic axis. Elsewhere in \( F_3 \) folds, the stretched minerals appear to be parallel to the "b" tectonic axis of that system. The two observations are not mutually exclusive, but determination of the true genetic relationship of the stretched mineral lineation in a particular outcrop is complicated by the nearly orthogonal array of the \( F_1 \) and \( F_2 \), and \( F_3 \) fold systems.

Microcrenulations formed by puckers in the foliation are also present in the Mount Lemmon area. The microcrenulations generally plunge to the northwest, and commonly occur.
in proximity and subparallel to folds of the $F_2$ system. These microcrenulations then seem to be a "b" lineation generated by the event which formed the $F_2$ system of folds and are herein referred to as $L_2$.

In the Mount Lemmon area, stretched and flattened clasts are present in the Abrigo Formation and in the Barnes metaconglomerate. The clasts are commonly oriented with their longest axes trending northeast or southeast (Figure 23). The major maximum of clast orientation in Figure 23A lies parallel to the predominant direction of stretched mineral lineation. In fact, most of the individual clasts show a pervasive stretched mineral lineation parallel to the longest axis. The clast elongation and orientation, like the development of the stretched mineral lineation, apparently result from more than one deformational event. Thus, these data cannot be automatically related to a specific time in the deformational history of the rocks nor to a constant orientation with respect to the tectonic axes. For instance, the major maximum on Figure 23A might indicate an alignment of pebbles parallel to the "b" tectonic axis of the $F_3$ system of folds, or to the "a" axis of the $F_1$ system. Equally ambiguous interpretations might be made from the minor maximum. As a matter of further confusion, the major and minor maxima may be related to different tectonic axes of the same fold generation. This latter suggestion is supported by the
observation that in several places, orthogonally oriented clasts occur a short distance apart in isoclinally folded phyllites (Figure 6). In one outcrop northwest of the radar station, stretched clasts oriented parallel to the "b" tectonic axis were found in a fold of the F₁ system.

Fractures

Décollement. Décollements parallel or subparallel to S₁ and S₂ are present in the Bear Wallow area. They are probably much more common and extensive within the entire study area than noted; however, the combination of complex folding and poor exposures makes it difficult to detect these features. Where recognized, the décollements are generally marked by nearly flat, curviplanar, shear zones or gouge zones, one to six inches thick. The faults commonly accompany large recumbent folds, and are imbricated. The imbricate sheets range from a few inches to several tens of feet in thickness. The décollements not uncommonly splay into several smaller wedges along the leading edge of the slip sheet.

Décollements with attendant folding are especially common along the contacts between the metasedimentary rocks of the Apache Group and the interlayered metadiabases(?) The metadiabases(?) appear to have acted with relatively
higher ductility during deformation and allowed the overlying less ductile layers to slide as a unit or sheet. Evidence of this form of deformation is well exposed along the Mount Lemmon Highway below the loop in Bear Wallow.

Many of the décollements recognized within the area seem to have been formed by the event which generated the \( F_2 \) system of folds. The décollements are coincident with the folding and commonly cut the overturned limbs of recumbent, concentric or knee-"isoclinal" similar folds of the \( F_2 \) system. In general, the rocks during the \( F_2 \) episode seem to have deformed initially by isoclinal folding. Ensuing stress-buildup probably caused attenuation of the overturned limb, and eventually stress relief was manifested as décollement. Subsequent movement along the slip surface was accompanied by drag folding and, undoubtedly, larger scale folding continued within the main body of rocks and within the sheet. As the folding again reached the isoclinal stage described above, another décollement appears to have formed.

Although many of the décollements noted in the field appeared to be generated during \( F_2 \) folding, the system of folds associated with other décollements could not be determined. Because of the similarity of the folds of both
systems, however, it would seem strange indeed not to have had décollement generated by $F_1$ folding also.

**Bedding Plane Slips and Wedging.** Bedding plane slips are probably very common throughout the study area. Similar to the décollements of larger size, the bedding slips are difficult to recognize. Again, the difficulty in detection arises from a paucity of good exposures, a lack of marker beds in many of the units, and the complexity of associated structures.

Where bedding slip is revealed, it is normally marked by wedging (Cloos, 1964). The wedges are most common in the interbedded sequences of phyllites and metaquartzites, and in metaquartzites containing thin phyllitic partings (Figure 24). Not uncommonly wedging in one layer is accompanied by folding in the adjacent strata.

In some places, bedding slip is accompanied by folding, without evidence of wedging (Figure 25). Small bedding plane displacements are also revealed by offsets in cross-cutting dikes or veins, and many of the bedding surfaces in the area are striated, indicating movement along the surface.

Bedding slip and wedging probably initiated the deformation and continued to be an important process throughout the entire structural history of the area. Few criteria for precise judgments as to the time of development of bedding slip and wedging are available; however, the mere
Figure 24. Wedging

Bedding slip resulting in wedging in the Bolsa metaquartzite on the northwest flank of Mount Lemmon. Facing northeast.
Figure 25. Sketch of bedding slip accompanying folding

Knee-"isoclinal" similar fold with accompanying bedding slip in the Bolsa meta-quartzite on the northwest slope of Mt. Lemmon. Trend of fold axis is S 65° E, and upper layers have moved south-southwest relative to the lower layers. Facing southeast.
presence of heterogeneities within the rock, be they bedding or foliation, invites suggestion that these surfaces have played a long and active role in the deformation.

Wedging is most commonly recognized in conjunction with disharmonic folds of the $F_1$ and $F_2$ systems. Many of the wedges appear to develop in the early stages of deformation and act as loci for subsequent folding in adjacent layers. Wedging is also common in less obviously deformed units but there wedging is generally more difficult to recognize.

The wedge shown in Figure 24 is exposed in the vertical limb of an $F_3$ fold in the Bolsa metaquartzite. The metaquartzite layer forming the wedge can be traced through a more gentle $F_3$ flexure to the right of the photograph, suggesting that the wedge predates the $F_3$ system of folds. This suggestion is strengthened by the close spacial association of the wedge and the bedding slip fold of the $F_1$ system shown in Figure 25.

The most recent evidence of bedding slip was observed on the northwest side of Mount Lemmon within the limits of the radar station. In this place, the strata dip in the same direction as the slope of the mountain, and have been "daylighted" by erosion. Dip slip has occurred along the bedding and has opened fissures as much as one foot in width. These fissures are presently filled with uncemented rubble and soil.
Thrust Faults. Thrust faulting within the area of investigation is generally of minor importance, but small thrusts accompany most of the fold styles, and all of the identified fold systems. Small thrusts are present in the hinges of many of the "isoclinal" similar folds and probably formed during the late stages of folding. They appear to represent a change in the mechanism of deformation resulting from a decrease in the overall ductility of the material. Faulting in the disharmonic folds may have occurred relatively earlier, during the main stage of folding, and probably reflects a higher ductility contrast between the various strata. This suggestion is supported by selective faulting of relatively stronger layers involved in the overall disharmonic folds.

One thrust with approximately 100 feet of displacement was noted in the flank of a large fold of the F_3 system. The fault is accompanied by smaller scale F_3 folding in the footwall, but the thrust plate appears relatively undeformed. Unsheared pegmatite is intruded along the fault. The lack of deformation in the plate would seem to indicate that the fault developed early in the F_3 event. The ensuing lateral movement which caused F_3 folding in the footwall may have been more vertically directed in the final stages allowing the pegmatite to enter the fault zone. A major thrust south of Butterfly Peak in the Bear Wallow area deforms the foliation, offsets rocks folded by the F_2 system, and may
represent a late or post-$F_2$ upward adjustment of the southern block.

**Joints.** Joints are ubiquitous throughout the study area. The intensity of development is variable and joint spacing ranges from less than an inch to several tens of feet. The closer spacings are by far the most common, and in many parts of the area, sheeting is penetrative throughout the entire outcrop. Several strike directions are especially pronounced, but the direction of greatest intensity is variable from one part of the area to another. In general, the following directions are represented at all locations by either a major or a minor maximum: N 10° to 20° W; N 10° to 20° E; N 60° to 70° E; and N 40° to 50° W. The dip of the joints within the area is commonly greater than 60 degrees.

The character of the joint surface and the continuity of the joints seem to be related to the direction of strike. Commonly, the joints striking north-northwest and north-northeast are continuous and exhibit a smooth regular surface. The northeast and northwest striking joints are discontinuous, en echelon, and exhibit more irregular, jagged or curvilinear surfaces. Some of the joint surfaces are slickensided, but insufficient data were gathered to determine if their orientations are preferred or random.
The most pronounced joint set in the Mount Bigelow area strikes N 10° to 20° E. This set, shown as a major maximum in Figure 22B almost perfectly bisects a line drawn between the major maximum and minor maximum on the orientation diagram. The orientation diagram shows the plunge of folds in approximately the same area. Folds of the \( F_1 \) and \( F_2 \) system are essentially homoaxial in this area, and both systems are included in the diagram.

Data from joints and \( F_2 \) folds in the outcrops inside the loop at Bear Wallow are presented in Figure 22A. In comparing these data from the two areas, it is interesting to note that a change in the trend of the \( F_2 \) fold axes is attended by a swing in the prominent joint set so as to preserve their orthogonal relationship. On the basis of these data, it is suggested that the pronounced N 10° to 20° E and N 10° to 20° W joint sets within the area are transverse joints genetically related to the formation of the \( F_2 \) fold system. The minor maximum in the rose diagram of Figure 22B appears to represent axial or longitudinal joints related to the same generation of folds. A corresponding minor maximum is not obvious in Figure 22A, but may be represented by the weakly developed joint set shown striking approximately N 55° W.

To attempt to corroborate the suggested relationship between the \( F_2 \) fold axis and the major joint set, data were gathered in the area between Mount Bigelow and Bear Wallow,
where the fold axes have more irregular trends. The trend and plunge of 44 folds, and the attitude of 300 joints were measured in the outcrops along the gravel road below the observatory. As noted earlier, superposed folding could not be definitely identified in these outcrops, but one exposure revealed a northwest plunging F₂ fold, folding the northeast plunging stretched mineral lineation. Although most of the folds could not be distinguished on the basis of folded and nonfolded foliation alone, those folds plunging subparallel to the stretched mineral lineation were designated F₁. Unfortunately, the orientation diagram makes only the barest suggestion of a bimodal distribution of the fold axes, and one received little additional encouragement for the designation of two fold systems. In comparing the orientation and rose diagrams of Figure 22, however, it seems more than a coincidence that the major maximum of the rose diagram is again found to be essentially orthogonal to the trends of the folds identified as F₂. It is also interesting to note that an orthogonal relationship exists between the minor maximum on the joint rose and the plunge direction of the folds identified with the F₁ system.

Several attempts were made to unravel the sequence of fracture development using small indicated offsets in the joint pattern. Most of the attempts revealed contradictory sequences, even over small areas, and no overall analysis was made as a part of this study.
No statistical diagrams were prepared for the Mount Lemmon area, however, joints in the north-northwest and north-northeast directions are essentially pervasive. The north-northeast striking joints are especially pronounced, and there appears to be a slight eastward swing to a new strike of N 25° to 35° E. This apparent reorientation may be a response to the generally more northwest-southeast trends of the F₁ and F₂ folds in the Mount Lemmon area. That is, it seems that the joints have maintained a transverse orientation to the fold axes. Joints striking in the northwest quadrant are also pronounced in the Mount Lemmon area.

Normal and Reverse Faults. In the area of study, many of the fractures in a transverse orientation to the F₂ fold system show displacement. Most of the faults indicate only minor displacement and cannot be traced beyond a single exposure; however, a number of the faults of greater displacement extend for several miles across the area. All of the faults dip very steeply, generally more than 75 degrees, but in some places it is not possible to determine the dip direction. Dikes of granite, amphibolite, pegmatite, or aplite fill some of the fractures and others are marked by gouge and breccia zones. The gouge and breccia zones range from several inches to as much as ten feet in width (Figure 26).
Figure 26. Normal(?) fault

Breccia zone in the Dripping Spring metaquartzite along the Mount Lemmon Highway. Facing southwest.
At least some of the faulting must have accompanied the folding in the late stages of the $F_2$ event (Figure 27). The major faults which share the important north-northeast and north-northwest directions common to joints are post-$F_2$ event and may have been caused by the event which generated the $F_3$ system of folds.

In the Mount Lemmon area, major faults occur in the east-northeast and northwest directions as well as along the more prominent north-northeast trend. Most of the faults have normal displacements and probably formed in the final stages of the $F_3$ event. It would be an oversimplification, however, to attempt to group all of these fractures into this late period as some indicate an earlier history. In some places the foliation in the wall rocks passes through dike rocks present in the fractures. Other dikes display no discernible directional fabric.
Figure 27. Normal fault and cascade folds

Fault in the Dripping Spring metaquartzite exposed along the Mount Lemmon Highway immediately west of the loop in Bear Wallow. Note the brittle fracture displayed by the metaquartzites outlining the cascade folds on the (west) down-dropped side of the fault. Layers on east side of the fault contain small "isoclinal" similar folds, average plunge S 55° W, 6°. Cascade folds on west side of fault plunge northward. Facing north.
CONCLUSIONS

Indicators of Tectonic Transport

Many of the folds throughout the area, because of their symmetry and style, allow interpretation of the kinematics involved in their formation. In the main, folds of the \( F_2 \) system are the most revealing; however, less commonly, \( F_1 \) and \( F_3 \) folds are sufficiently well exposed to permit interpretation of the relative movements involved in their formation.

The knee-"isoclinal" similar fold illustrated in Figure 15 indicates that the upper layers involved in the fold have moved relatively to the left over the lower layers. Like interpretations can be made from the larger knee-"isoclinal" similar forms such as shown in Figure 28. In this fold, the relative movement of the beds in the upper part of the photograph appears to have been to the right over those layers in the lower part. The entire fold appears to be part of a large glide sheet which has moved northeastward over the underlying rocks. Where smaller folds have been formed by movement along décollement, the sense of movement of the décollement can be inferred from the folds. In addition, some of the wedges and bedding plane slips marked by folds or striations can provide uni- or bidirectional tectonic transport data.
Large knee-"isoclinal" similar fold in the steep cliffs northeast of Turkey Flat. The decollement which cuts the limb of the fold in the foreground can be traced northward along the cliffs where it is parallel to the layers in the "isoclinal" similar portion of the fold. Facing northwest.
To provide an overall picture of the kinematics of the deformation in the area, arrows indicating the sense of movement of the upper layers of the interpretable folds are plotted on Figures 22 and 23. Most of the interpretations are based upon knee and cascade folds of the \( F_2 \) system. The "isoclinal" similar folds common in the \( F_1 \) system generally yield equivocal tectonic transport directions, and bidirectional interpretations are commonly the best that can be obtained. In portions of the area where the two fold systems are homoaxial, the writer had only limited success in identifying the folds. Therefore, unless the system of a fold could be definitely identified, the fold was generally assumed to belong to the \( F_2 \) generation.

In the Mount Bigelow-Mount Kellogg area, the tectonic transport directions indicated by the \( F_2 \) system of folds are generally northward. To the west, along the spur running north-northeast from Turkey Flat, the transport directions are consistently to the northeast. North and west of Turkey Flat, interpretable data are difficult to find, and where available give a scattering of movement directions. In the Bear Wallow-Spencer Peak portion of the area, the principal tectonic transport direction appears to be southward, although in the immediate vicinity of Bear Wallow local perturbations are present.

In the Mount Lemmon area (Figure 23) the directions of tectonic transport as indicated by folds of the \( F_1 \) and
systems and bedding slips, are generally to the southwest. Unfortunately in this area, folds which indicate the direction of tectonic transport are neither numerous, nor evenly distributed and large areas with no data or only scanty data exist.

**Causes of Folding**

The cause of the movement delineated on Figures 22 and 23 might best be approached by inspection of Figure 29. This photograph shows the knee-"isoclinal" similar fold pictured in Figure 16, but here the fold is presented in relation to the larger structure in the locality. It is evident from Figure 29 that the knee-"isoclinal" similar fold occurs as a parasitic fold on the limb of a larger anticline. Furthermore, it appears as though the knee-"isoclinal" similar fold is caused by movement of material away from the crest of the anticline and under the influence of gravity.

This same process is much more dramatically illustrated on a grand scale in the cliffs along the north slope of Mount Bigelow. In these cliffs, large knee and cascade folds and flap structures (Harrison and Falcon, 1936) are beautifully exposed (Figures 28 and 30; Figures 31 and 32, in pocket). The recumbent folds shown in Figure 31 have their greatest wave length and most regular form in the steep cliffs below the summit of the mountain. Farther to the north, the wave length decreases, and the folds become
Figure 29. Parasitic fold on the limb of an anticline

Knee-"isoclinal" similar fold on the flank of a larger anticline in Dripping Spring metaquartzite in the loop at Bear Wallow. Facing east-southeast (see Figure 16 also).
Figure 30. Cascade folds

Large folds in the Dripping Spring metaquartzite on the north slope of Mount Bigelow. Facing northwest.
more cascade-like in form. The zone between these somewhat different forms is composed of amphibolites which are more intensely deformed and apparently were much the more ductile during deformation. From the evidence available in this area, the conclusion that these folds have formed by gravity gliding from a higher area to the south seems inescapable.

Although the postulated high would be outside the study area, further evidence to indicate its existence is present on the south slopes of Mounts Bigelow and Kellogg. The contact of the Catalina Gneiss with the rocks of the Apache Group in this particular area was described earlier as intrusive. The contact is structurally conformable, convex to the north, and dips generally northward. Dips along the contact become steeper in the convexity and reach vertical to slightly overturned at its northernmost extension. The increase in the steepness of the dip of the contact and foliation is thought to reflect active upwelling in the mobile gneiss, and the convexity or protrusion is interpreted as a slightly more active bulge on the northern flank of a somewhat larger zone of active upwelling.

In the area surrounding Bear Wallow and Spencer Peak, the indicated tectonic transport directions are generally southward with exception of a limited area in the loop at Bear Wallow (Figure 22). Puzzling, and in contrast to the Mount Bigelow area just discussed, the tectonic transport
directions in this area are toward the Catalina Gneiss. Tectonic transport in this direction suggests that during the $F_2$ event at least the gneiss south of Bear Wallow was relatively inactive. This suggestion is strengthened by the more metasomatic character of the contact in this area as compared to the intrusive contact on Mount Bigelow. Southward tectonic transport also indicates a general upward inclination in the structural gradient toward the north. The fact that the strata presently dip generally southward is in keeping with this requirement; however, before attempting to define the position and nature of a feature which could provide the needed gradient for gliding, the structures within this area should be reviewed briefly.

On the western slope of Spencer Peak above Sabino Canyon, folds of the $F_2$ system indicate south-southeast transport (Figure 33). To the east-northeast of Spencer Peak, and along the Mount Lemmon Highway, similar transport directions are indicated by folds above the décollement. Some of the folds immediately above the décollement indicate transport directions which deviate considerably from the general trend. These deviations probably represent minor deflections of the sheet resulting from local increases in resistance to gliding. Large recumbent folds bounded by décollement are present along the ridge north of the Upper Bear Wallow Campground. The transport directions of
Figure 33. Knee folds of the F₂ system

F₂ folds in the Dripping Spring metaquartzite above meta-diabase(?). Note faint trace of folded F₁ folds above kinematically active pegmatite. Attitude of the F₂ fold axis is S 80° W, 4°. The upper layers have moved relatively southward over the lower layers. Facing east.
these folds, and the drag folds which accompany the décollement, are commonly difficult to determine, but seem to be southward. The fact that some of the décollement splay southward into imbricate wedges strengthens the southward transport interpretation. Additional evidence of southward tectonic transport is present along the highway below the loop in Bear Wallow, and in the vicinity of Soldier Camp.

The principal departure from the southward tectonic transport occurs in the loop at Bear Wallow, where outcrops at the eastern end of the loop in the highway show opposing transport directions. The south end of the outcrop contains knee folds with southerly transport directions, and the northern end contains cascade folds in shale that indicate tectonic transport to the north. Additional northward transport directions are indicated by some of the structures exposed for a short distance along the road to Soldier Camp (Figure 27). Approximately 1,000 feet northwest of the loop, along the dirt service road, several folds exposed in the road cut also indicate northward tectonic transport.

Deformation in much of the exposure in the loop is so intense that it is difficult to delineate the structure in detail; however, the outcrop seems to expose a local center of up-buckling and flowage. Diapir-like masses of metaquartzite in phyllite are exposed in the center of upwelling, and "isoclinally" folded metaquartzites and phyllitic metaquartzites are exposed to the south of the welt. The units
on the south flank have "cascaded" away from the rising center and formed a series of knee folds and open buckles. An obvious décollement exposed in vertical metaquartzites near the center of the exposure appears to have acted as a locus of movement from which knee folds originate. Although tectonic transport is primarily to the south, some northward flowage on the north side of the minor structures is evident (Figure 29). On the north flank of the welt, phyllites appear to have spilled northward against the main tectonic "current." This welt, unlike the zone of upwelling south of Mount Bigelow may not have "roots." More probably, it is restricted to the cover rocks and represents a "tumor" formed by flowage of ductile materials into a relatively low pressure site. Such a site of low pressure may have been caused by buckling within the less ductile rocks. Buckles might logically develop within the cover rocks as a result of differential gliding rates.

The areal extent of the welt is not known. Folds along the south side of the welt can be traced for some distance eastward, but unfortunately there seem to be no reliable criteria for sorting out the locally generated structures from those due to the general southward transport. Immediately west of the exposures in the loop, the welt is upthrown by a north-northeast striking fault. Aside from the fact that the knee folded metaquartzite exposed in the loop
can be identified in the top of the road cut on the west side of the fault, no other structural continuity is revealed.

Mescal marble has been identified along the north side of the cabin-studded ridge above the Upper Bear Wallow Campground. In the cliffs below the last two cabins on the eastern end of the ridge, the Mescal crops out beneath large recumbent folds in the Dripping Spring metaquartzite. The folds in the Dripping Spring are cut by a décollement which splays southward, and although the sequence appears in reverse order, there is no definite evidence that the section is actually overturned. A second, lower, décollement is exposed farther to the west, but cannot be traced for any distance eastward because of poor exposures.

Several alternatives may be suggested to explain the reverse order, or perhaps overturned stratigraphic sequence. The presence of a series of glide sheets in the area suggests that the Dripping Spring rests on an unexposed décollement above the Mescal and has glided into position from the north. An alternative to this solution would be to infold the Mescal into a recumbent syncline. This alternative would require northward tectonic transport, evidence of which is lacking, in the immediate area. Mescal marble, however, also crops out in the vicinity of the westernmost cabin above the loop where northward transport is indicated. Since the Mescal exposures seem to be along the projected trend of
the structural high exposed in the loop, perhaps the marble was infolded by northward transport from the locally rising welt, then overridden by the southward gliding sheets.

Although the Bear Wallow area lacks the broad panorama of arrested folds cascading from a topographically higher area, as displayed in the cliffs of Mount Bigelow, the case for gravity gliding is no less strong. Again in this area, the amphibolites and dark, biotite phyllites (metadiabases?) have acted with higher ductility than the adjacent metasedimentary rocks and have served as particularly active gliding zones. It is no coincidence then, that as the observer works his way northward up the structural gradient, down stratigraphically, metadiabases(?) are more abundant in the highly disturbed section. If, as certainly is in evidence, tectonic transport was to the south, the metadiabases(?) would have provided excellent glide zones as they became uplifted to the north.

The most obvious feature to accomplish the uplift is the Leatherwood "stock" which intrudes the metadiabases(?) along the northern border of the area. The suggestion that the Leatherwood Quartz Diorite pluton provided the needed gradient for southward gliding is strengthened by the occurrence of what seems to be intensely sheared Leatherwood in exposures along the Mount Lemmon Highway immediately north of Soldier Camp. In these outcrops, the Leatherwood (?) is involved in the deformation and indicates transport toward
the south. Weathered, less sheared and foliated Leatherwood crops out in the ravine 800 feet north-northeast of the exposure along the highway. The difference in texture may be due to more intense cataclasis nearer the border of the actively upwelling "stock." The most serious objections to the proposal that the Leatherwood "stock" served as a zone of upwelling are: (1) the presence of Apache Group and lower Paleozoic rocks dipping northward into the Leatherwood at the relatively low elevation of Butterfly Peak (Figure 2), and (2) acceptance of a more complicated hypothesis which involves opposing tectonic "currents."

Folds within the northward dipping units beneath Butterfly Peak have the same trends as those found in the southward dipping rocks south of Butterfly Peak, and where tectonic transport directions can be obtained, southward movement of the upper layers is indicated. A flat fault within the metadiabases(?) separates the rocks of opposing dips, and has thrust lower units of the Dripping Spring over the Mescal marble. The movement on the fault has caused noticeable drag of the Dripping Spring metaquartzite in the footwall, and dips change from 12 degrees 300 feet from the fault to 45 degrees adjacent to the thrust. This fault appears to be a major feature of considerable displacement and probably indicates uplift of the southern block after the F₂ folding event. It is quite doubtful, however, that the elevation difference between the Apache Group rocks in Bear
Wallow and those on Butterfly Peak could be accounted for by movement on this thrust alone. Thus, the north dipping "but­tress" of rocks on Butterfly Peak remains partially incom­patible with the postulated role of the Leatherwood "stock" and the evidence of southward tectonic transport in the Bear Wallow area.

If, as has been suggested, opposing overall tectonic transport directions are present in the Bear Wallow and Mount Bigelow areas, what of the area in between? In the cliffs along the ridge northeast of Turkey Flat, cascade and knee folds of the F₂ system are well exposed (Figure 32). Formation of these folds would require movement of the upper layers to the northeast, and the presence of a structural high in the vicinity of Turkey Flat. Figure 2 reveals a northward projection of the Catalina Gneiss following the con­tours north of Boy Scout Spring and a small patch of gneiss exposed in the ravine west of Turkey Flat. These exposures indicate that in the Turkey Flat vicinity, the rocks of the Apache Group form a very thin veneer over the gneiss. The presence of the veneer and the evidence of tectonic trans­port from this area would seem to allow the suggestion that a lobe of Catalina Gneiss similar to that on Mount Bigelow lies at a very shallow depth below Turkey Flat.

In contrast to the well formed cascade and knee folds northeast of Turkey Flat, north and west of the flat the folds are small and of mixed styles. In this area, where
tectonic transport directions can be interpreted, they generally indicate opposing directions of movement, and many of the interpretations are equivocal. The character of the structures and the lack of clarity in the transport directions indicated within this zone may reflect the opposing tectonic "currents" on either side of the zone. Such a suggestion can only be inferred from the data available. We shall, however, return to this inference further on in this section.

In the Mount Lemmon area, folds of the \( F_1 \) and \( F_2 \) systems and other structures that indicate direction of tectonic transport are neither numerous nor evenly distributed. Large areas with no data or only scantly data exist, and locally conflicting tectonic transport directions are revealed. Décollements have not been recognized in this area; however, a broad range of ductilities is indicated by the rock units. This observation invites speculation that slip sheets may not be uncommon, especially in the interbedded phyllites and metaquartzites of the Abrigo Formation. Certainly, the presence of wedging and bedding slip in the rocks encourages this view. The fact that many of the units appear to undergo considerable changes in thickness from place to place also suggests that décollements may be important.

The majority of the folds indicate that the upper layers have traveled southwestward over the underlying
strata. This indicated tectonic transport to the southwest would require a structural gradient inclined upward to the northeast. The spatial relationships of the northwest and southeast trending folds and the Leatherwood Quartz Diorite in Figure 23 mark the Leatherwood "stock" as a prime feature to have created the gradient required for gliding. Again, however, as in Bear Wallow, there is contradictory evidence in this area.

One of the most obvious objections to gravity gliding from the northeast is the fact that presently the rocks dip northeast. It would seem more logical perhaps for gravity gliding to have taken place down the present dip suggesting a rising welt to the southwest. Under these conditions, the direction of asymmetry of the folds in the area would require under gliding; that is, the upper layers become tucked under in the direction of inclination of the structural gradient. The resulting fold would be asymmetric toward the welt. Folding apparently generated in this manner has been noted in Arrastre Wash in the Tucson Mountains; however, within the area of study where folds can be definitely related to a center of upheaval, the direction of asymmetry with very few exceptions indicates that the upper layers have overridden the lower layers down the structural gradient. It must be conceded that these observations of the asymmetry of the folds, with respect to the structural gradient, were made in Bear Wallow and on Mount Bigelow. It is possible that these
relationships are reversed in the Mount Lemmon area. The writer, at this point at least, prefers to assume that the rocks behaved similarly in both areas and suggests that the present northeast dip of the rocks is due to postgliding uplift in the southwestern part of the area. The suggestion that uplift followed the development of the F₂ system of folds is not new. The reader will recall the presence of the upthrust near Butterfly Peak. The character of the uplift proposed for the Mount Lemmon area would differ only in that it probably was associated with intrusion and metasomatism rather than the rigid block uplift indicated south of Butterfly Peak. This uplift may be genetically related to the F₃ folds.

The F₃ system of folds seems to be generated by upwelling and lateral crowding by the Catalina Granite and the Catalina Gneiss. In Zone I of the Mount Lemmon area (Figure 23), the trend of the F₃ folds is subparallel to the trend of the Catalina Granite, and the northwest limbs of the folds are slightly steeper. The emplacement of the granite as described earlier was largely passive, and the pluton appears to have intruded along an already established northeast trend. In the area of study, the contact of the granite generally dips steeply northwest and seems to be controlled by northeast striking fractures and the attitude of the wall rocks. The slight westward asymmetry of the F₃
folds in Zone I is probably due to the upward and slightly southeastward advance of the intrusive.

Folds of the F₃ system in Zone III seem to have a more easterly trend than those adjacent to the Catalina Granite. This observation may be real, or only apparent as relatively few data are available in these areas. The folds of the F₃ system are not as well developed adjacent to the Catalina Gneiss (Zone III) as in Zone I, and there seems to be little or no preference in the direction of asymmetry. These observations would agree with the rather wholesale pegmatization and metasomatism associated with the Catalina Gneiss in this area. The implication is that the emplacement of the gneiss was more passive and largely by metasomatic processes in contrast to the slightly more active, intrusive-metasomatic role of the Catalina Granite.

Inspection of Figure 23 reveals a preferred location of folds of the F₃ system. The F₃ folds are most common in a "V"-shaped area (Zones I and III) oriented with the open part of the "V" to the northeast. The northwest arm of the "V" (Zone I) trends northeast and lies adjacent to the Catalina Granite. The southeast arm (Zone III) trends east-northeast and lies adjacent to the Catalina Gneiss. In contrast to the predominance of folds of the F₃ system within Zones I and III, there is a relative scarcity of folds of the F₁ and F₂ systems. Conversely in Zone II, there appears
to be a lower average density of folds of the $F_3$ system. These anomalies of fold system density, in part at least, are real and reflect the cause of folding and the ductility of the rocks during deformation. However, such factors as the amount and quality of exposure, and the relative difficulty in detecting folding within some of the rock units must also play an important role in the data distribution.

On the eastern slope of Sabino Canyon, folds of the $F_3$ system are present in the wall rocks adjacent to the Catalina Gneiss contact (Figure 22). The folds are asymmetric toward the intrusive, and appear to be formed by drag within the wall rocks caused by lateral crowding of the Catalina Gneiss. The folds are tightly appressed near the gneiss, and in general open into broader forms with increased distance from the contact. Chevron folds with like trends and orientations are also abundant, especially near the gneiss. The contact in this portion of the area is irregular, but in part is composed of regular north-northeast and north-northwest trends. Farther to the east, the same regular trends again dominate the contact, and in the vicinity of these trends, $F_3$ folds are again common. The outcrop containing three directions of fold axes discussed in detail earlier, is located still farther to the northeast and lies along the projected trend of a major north-northeast to north-northwest fault zone. The association of the north-northeast and
north-northwest trending fractures, similarly trending contacts, and the subparallel \( F_3 \) system of folds appears to be more than a coincidence. On the basis of this association, it seems reasonable to suggest that upwelling of mobile rocks along the north-northeast and north-northwest directions has caused uplift and lateral crowding with resultant deformation along these trends.

The origin and time of formation of the folds of the \( F_1 \) system are problematic. At least two possibilities deserve mention. One possibility is that these folds represent the early stages of one event that formed both the \( F_1 \) and \( F_2 \) systems. The alternative is that the \( F_1 \) folds were formed by a distinctly separate event which predated, and was separated in time from the second event which formed the \( F_2 \) system. The relative simplicity of a single event, in that it requires only one cause, or at least one activation of a source of deformation, is attractive to the writer. The fact that the fold systems are commonly homoaxial, contain the same fold styles, and that many stages in the development of these styles are revealed, permits the suggestion that the folds pictured in Figure 21 may have developed in a series of continuous stages. Figure 34 illustrates several stages in the suggested evolution of the superposed folds. According to this evolution, \( F_1 \) "isoclinal" similar folds would form by gravity gliding during the period of metamorphism. As the conditions of metamorphism waned in the cover rocks,
Stage 1, $F_1$ "isoclinal" similar folds

Stage 2, knee folds superposed upon $F_1$

Stage 3, "isoclinal" similar folds superposed upon $F_1$

Figure 34. Diagrammatic sketch of selected stages in the generation of superposed folds by one continuous episode of deformation.
upwelling may have continued at a depth causing postmetamorphic deformation at higher levels. During these late stages the imposed fabric elements became involved in the continuing deformation and the \( F_2 \) folds were formed. In some places deformation continued beyond the point at which the rocks could yield by folding, and flat glide faults of small displacements were formed (Figure 23). This outline of the evolution of the structures is oversimplified and represents only the clearly recognizable stages and conditions.

As noted earlier, in many folds it is difficult to determine whether \( S_2 \) is folded. Indeed, in several folds, it appears as though \( S_2 \) is folded in some layers, but passes undeformed through the hinge in other layers. Such a judgment is difficult to make without substantial petrofabric work. If proven to be true, the observation could strengthen the argument for one continuous generation of deformation.

In some parts of the area, the \( F_1 \) and \( F_2 \) fold systems are not homoaxial. Divergences of the axial trends of the \( F_1 \) and \( F_2 \) systems probably reflect changes in glide direction during deformation. It seems quite reasonable to expect tectonic transport directions to change as resistance to movement builds up within the gliding mass. Divergences of the axes might be expected especially in areas of opposing tectonic "currents." Earlier in this discussion, it was
inferred that the area north and west of Turkey Flat represented a zone of interfering tectonic "currents." It may be merely coincidental, but in this area there is divergence between the axial trends of the two fold systems (Figure 22).

Generation of the \( F_1 \) and \( F_2 \) systems of folds by two distinct and separate events does not necessarily present any great obstacles. Actually, the two likely alternatives need differ only in that to have two separate events, the centers of upwelling must be activated at two different times rather than undergo a single activation. The overall time for the deformation might well be the same. The fact that the systems are commonly homoaxial near the zones of upwelling, would seem to require that the same principal zones be active during both events. In this alternative, the areas in which the \( F_1 \) and \( F_2 \) axes diverge may be due either to small changes in the centers of upwelling, or to changes in the tectonic slope effecting the two events. The lack of clarity in determining whether foliation is folded in some folds might be just as easily explained by having two distinct events, if the second, milder event were sufficiently intense to cause local partial recrystal- lization.
The contact of the Apache Group with the underlying Catalina Gneiss within the area of this report is generally metasomatic and intrusive. The lower portion of the section—the Scanlan metaconglomerate, the Pioneer Formation, and the Barnes metaconglomerate—is generally thinned or missing. The presence of the Scanlan metaconglomerate locally, however, implies that the original contact was sedimentary. Unfortunately, within the area there is little indication of the original character of the pre-Scanlan basement.

Younger Precambrian rocks in relatively undisturbed, unconformable relation with the Older Precambrian basement are exposed in the Santa Catalina Mountains but commonly on the lower flanks of the range. Such contacts have been described by Wallace (1955), Banerjee (1957), Peirce (1958), Raabe (1959), Erickson (1962), Creasey (1967), Braun (1968). The Older Precambrian rocks described by these workers generally comprise two different rock types. The most common is Oracle Granite, or quartz monzonites and granites similar to the Oracle in composition and texture; the second is Pinal Schist. North of the Mogul fault near the town of Oracle, biotite from the Oracle Granite yields a K-Ar age
of 1,420 m.y. (Damon, Erickson and Livingston, 1963), as compared to an age of 1,450 m.y. obtained by the Rb-Sr method (Giletti and Damon, 1961). Reported exposures of Apache Group rocks in unconformable contact with Pinal Schist are confined to the northwest part of the range (Erickson, 1962; Wallace, 1955). Pinal Schist (?) beneath the Apache Group has been reported (Raabe, 1959) in Bullock Canyon on the east flank of the range. Banerjee (1957) presented convincing evidence indicating that the Oracle Granite was formed by metasomatism of the Pinal Schist. He suggested that the metasomatism took place after the intense folding stages of the Mazatzal Revolution.

In general, higher in the Santa Catalina Range, such as in the area of this investigation, the clearly sedimentary unconformities change to intrusive and metasomatic contacts between the Catalina Gneiss and the overlying metamorphosed Apache Group. The Catalina Gneiss at these contacts is generally a granitic gneiss (granitic gneiss-gneissic granite of DuBois, 1959), but banded augen gneiss in contact with the Apache Group has been reported by Pilkington (1962). It would appear then, that sometime after late Precambrian the Older Precambrian basement rocks—similar or dissimilar to those noted farther down on the flanks of the range—became mobile and intruded and metasomatically replaced portions of the overlying cover rocks. The time and implications
of mobilization of the basement rocks will be discussed later in the development of the history of the study area.

The rocks of the Cambrian System lie unconformably upon the Apache Group. This unconformity does not cut as deeply into the Apache Group in the Mount Bigelow-Bear Wallow area as elsewhere in the Santa Catalina Range. As a result, Mescal marble is found throughout much of the area. Sometime during the final phase of the late Precambrian, or at least before the rocks of the Middle Cambrian were deposited, the Apache Group was dilated by diabase sills and dikes. Aside from these abundant diabase intrusions in the Apache Group, there is no apparent evidence of a structural event, which involved the rocks of the younger Precambrian and not the overlying Cambrian rocks. On a regional scale, however, the Precambrian rocks appear to have undergone broad uplift or warping (Shride, 1967).

If the Troy Quartzite was deposited, it seems to have been stripped off during early Cambrian time. By Middle Cambrian, the seas again encroached upon the area, and the Bolsa Quartzite and Abrigo Formation were deposited. No Paleozoic rocks younger than the Abrigo are present within the area of study; however, rocks of the Devonian, Mississippian, Pennsylvanian and Permian Systems are present along the northeast flank of the range (Peirce, 1958; and Creasey, 1967).
The thicknesses of the Paleozoic units as shown by McKee (1951) total approximately 4,500 feet, and the thicknesses of the Apache Group sedimentary rocks listed by Creasey total approximately 900 feet. If an estimated 600 feet of diabase is added to the Apache Group section, the total rock accumulation above the Older Precambrian basement at the close of the Permian Period was approximately 6,000 feet.

No Mesozoic sedimentary rocks crop out within the area of this study, however, McKee (1951) showed a total of 15,000 feet of Paleozoic and Mesozoic rocks had accumulated by the end of the Mesozoic Era. McKee (1951) and Creasey (1967) have tentatively identified conglomerates cropping out near Peppersauce Canyon as Cretaceous in age, and McKee suggested that the conglomerates may be Lower Cretaceous. Creasey named these Cretaceous (?) rocks the American Flag Formation. He stated that it was in these rocks that Bromfield (1950) found the fresh-water pelecypod Unio and the gastropod Viviparus. The American Flag Formation is of special interest in that it contains an abundance of coarse conglomerates that probably reflect uplift in the Catalina Range during Cretaceous or perhaps pre-Cretaceous (Nevadan) time. According to Creasey (1967, page 42) in one section near Peppersauce Canyon fine- to medium-grained graywackes characterize the lower part of the section, and conglomerates the upper part. The beds become more conglomeratic
upward in the section, and the relative abundance of the different rock types among the boulders and cobbles changes. The clasts in the lower beds are largely volcanic. Higher in the section, quartzite clasts are abundant and are accompanied by sparse clasts of limestone, gneiss and granite. The matrix material and interbeds throughout the section are graywackes. The presence of volcanic boulders and cobbles in the section indicates that volcanism accompanied the uplift, and the gradual change upward in the composition of boulders suggests progressive denudation of the uplifted area.

The American Flag Formation is intruded by a granodiorite porphyry. Creasey believed that this porphyry is correlative to the San Manuel Quartz Monzonite which has been dated radiometrically (Creasey, 1967). The biotites of the San Manuel body yield K-Ar dates of 69 and 65 million years (Rose and Cook, 1965; Creasey, 1965), a late Cretaceous age. These dates strongly reinforce McKee's original suggestion that the conglomerates (American Flag Formation) are Lower Cretaceous, and probably reflect uplift in early Cretaceous or Nevadan time.

The description of the granodiorite porphyry that intrudes the American Flag Formation is strikingly similar to the foliated quartz latite porphyry described by the writer and by Pilkington (1962), and also to the cataclastic Leatherwood of Hanson (1966). The latter three units
are almost certainly the same rock, and the writer suggests that they represent a metamorphosed phase of rocks genetically related to the granodiorite which intrudes the American Flag Formation. The quartz latite porphyry may be older than the Leatherwood as Hanson suggests, but because of the rather broad areal extent of the sill-like bodies, it may be more logical to refer the foliation and cataclasis to a more widespread structural event than the intrusion of Leatherwood pluton. In the area of this study, the foliation in the quartz latite porphyry appears to be folded by \( F_2 \), and was probably generated by the main metamorphic event affecting the entire area.

If the assumptions and correlations concerning the Leatherwood and various granodiorite porphyry and quartz latite porphyry bodies are valid, then the Leatherwood "stock" cannot be older than late Cretaceous or early Tertiary. This assumption is extremely tenuous, however, since there is no evidence that precludes an earlier age for the Leatherwood pluton. A K-Ar date of 48.2 million years was obtained from coarse muscovite from a pegmatite in the Leatherwood Quartz Diorite near Summerhaven (Damon and others, 1963). This date cannot be accepted without reservation, as the investigators interpret the age as indicating incomplete degassing resulting from the large dimensions of the
pegmatitic mica (Damon and others, 1963). By their interpretation, however, the Leatherwood must then be older than the date yielded.

One of the tentative conclusions resulting from the present investigation is that the Leatherwood "stock" acted as a center of active upwelling from which cover rocks glided southward toward Bear Wallow and Mount Lemmon. This investigation also revealed that an actively rising dome south of Mount Bigelow caused gravity gliding to the north. These areas of upwelling seem to have been active at approximately the same time, and the fold systems, F₁ and F₂, formed during the gliding are concluded to be essentially synmetamorphic and late or postmetamorphic respectively. Radiometric dates obtained from micas in the Catalina Gneisses by the K-Ar method yield an average age of 26.8 ± 1.7 million years (Damon, 1967) as cited by Peterson (1968). According to Damon and others (1963), these dates probably indicate that uplift in the Catalina Range had progressed to the point that metamorphism had terminated by late Oligocene-early Miocene time. This assumes that degassing did not continue for a significant period beyond the close of the period of metamorphism. This is not an unreasonable assumption (oral communication, P. E. Damon, Department of Geochronology, The University of Arizona). If this premise is accepted, the F₂ system of folds would have formed sometime during the late Oligocene-early Miocene.
Judging from the information available, it seems reasonable to suggest that the Leatherwood "stock" was emplaced into the cover rocks during Laramide time; that is, late Cretaceous or early Tertiary. Magma may have been generated at depth and become separated or squeezed off into the cover as a discordant pluton. At the level of intrusion, the rocks appear to have been fairly cool and a narrow zone of contact metamorphism developed around the periphery of the intrusive.

Later in the Tertiary, perhaps early Oligocene, heat which had been building up deeper within the Catalina complex reached the structural level of the cover rocks. The heat and fluids probably entered along fracture intersections or other already established centers of weakness, and the cover rocks became hot, fluid-rich and quite ductile. Conditions of metamorphism, previously existent at depth, now reached into the cover rocks. The hot fluids migrating upward in the zones of weakness caused a volume increase by expansion, and by addition of material. This increase in volume resulted in a swelling in the area surrounding the zone of influx. As the swelling or active upwelling took place, the overlying material moved down the flanks of the welt under the influence of gravity. Within the area of this study, the Leatherwood "stock" and the Bigelow upwelling shed their cover, and the F₁ folds were formed. This
hypothesis suggests an early to middle Tertiary date for the $F_1$ system of folds.

The time of the actual beginning of the formation of the $F_1$ folds is difficult to estimate, however, and it may be that conditions of metamorphism in the cover rocks were already in existence at the time the Leatherwood pluton was emplaced. This would indicate a Laramide beginning for the generation of the $F_1$ folds. Although this latter suggestion would seem to be in conflict with the dual episodic character of the magmatism in the Basin and Range Province (Damon and Mauger, 1966), the conflict seems partially reconcilable. For instance, it may be argued that the histograms of Damon and Mauger showing magmatic pulses at the end of the Cretaceous and in the middle Tertiary reflect volcanism and hypabyssal pluton emplacement. Metamorphism at depth, however, may have continued between the two periods of magmatic activity. Also, until the age of the Leatherwood is substantiated, it would seem unwise not to give at least some consideration to the possibility that the pluton was emplaced early in the mid-Tertiary event which affected the Santa Catalina Range. The $F_1$ folds, then, would have been generated by emplacement in the Tertiary rather than by reactivation as suggested above.

Although the Leatherwood and Mount Bigelow centers of upwelling appear to have been active essentially synchronously, the Catalina Gneiss appears to have remained
active for a longer period and reached a higher elevation. During the time of the $F_1$ and $F_2$ fold generation, the area around Mount Lemmon must have lagged considerably to allow gliding toward the present summit. After activity in the Leatherwood decreased, the area surrounding Mount Lemmon probably continued to be uplifted by the Catalina Granite at depth.

During the late stages of metamorphism the Catalina Granite probably migrated from the site of generation and intruded the cover rocks along the northwest border of the area. The pluton seems to have intruded upward and slightly southeastward forming the $F_3$ system of folds in the cover to the southeast. The Leatherwood Quartz Diorite is also intruded by the granite (Peirce, 1958). Radiometric dating of a coarse-grained granite in Cargadero Canyon on the west flank of the Catalina Range yields a K-Ar age of $25.0 \pm 3$ m.y. (Damon and others, 1963). The granite in Cargadero Canyon has been identified as Catalina Granite by McCullough (1963) who showed the unit as continuous with the granite cropping out on Reef of Rock.

The activity in the Catalina Granite was probably accompanied by uplift in the Catalina Gneiss near Lookout Rock and less pronounced $F_3$ folds were formed in the cover rocks adjacent to the gneiss. $F_3$ folds were also generated in the area around Spencer Peak and Soldier Camp. At this
time, uplift by upwelling or upthrusting seems to have occurred along the entire southern margin of the study area. The flat thrust exposed south of Butterfly Peak and the Geesman fault were probably formed as a result of this uplift. The combined effects of the uplift in the south and the down faulting in the north may account for the structurally lower position of the Leatherwood and the northward inclination of the "buttress" of Apache Group and Cambrian rocks dipping toward the intrusion beneath Butterfly Peak.

The concept of the mantled gneiss dome (Eskola, 1949) and/or the migmatite dome (Haller, 1956) have been proposed to explain structures in other parts of the Santa Catalina Mountains (Pilkington, 1962; McCullough, 1963; Peterson, 1968). The centers of active upwelling with attendant gravity tectonics described in this report generally support either of these similar concepts. Some dissimilarities exist, but these differences do not present serious obstacles to the application of either model. Neither Haller nor Eskola have described gravity gliding as a process associated with mantled gneiss or migmatite domes, respectively. Yet this does not mean that gravity tectonics would not be attendant under the proper circumstances.

The fact that both concepts used here as models were described from orogenic belts where great thicknesses of sediments were present is perhaps a more serious nonparallelism of conditions. Deposition in the Catalina area has
probably never exceeded 17,000 feet at any time since the end of the older Precambrian. Anomalously high heat introduction into the area along fracture intersections would seem to adequately offset this comparatively thin accumulation of cover rock. Indeed, such an explanation would account for the rather localized centers of upwelling and the abrupt change in intensity of metamorphism areally.

The fracture directions which appear to have exerted important heat control in the area of investigation are approximately N 70° W, N 10° E to N 10° W, and N 25° to 35° E. The N 70° W (Texas lineament) direction seems to control the general trend of the Catalina Gneiss bordering the study area on the south and may determine the alignment of minor areas of upwelling within the gneiss. N 10° E to N 10° W are pronounced fracture directions that seem to have influenced local upwelling and the formation of F₃ folds in the Bear Wallow area. Prominent jointing in these directions is perpendicular to the trends of the F₂ folds in the Bear Wallow-Mount Bigelow area. On the basis of the orthogonal relationship, it was suggested earlier that the fractures were genetically related to the folds and represent transverse joints. The assertion that these directions are also important elements in the control of heat influx at first may seem to be a contradiction, since the gliding which generated the folds is largely confined to the cover rocks.
It must be considered, however, that the fold trends are roughly controlled by the orientation of the zone of upwelling. Thus the joint directions appear to be orthogonal to the long axes of the zones of the active upwelling as well as to the fold axes. It is not clear though, whether these north-northeast and north-northwest joint directions were originally generated by extension along actively elongating domal structures, or whether they are part of an older basement fracture pattern that became reactivated and reestablished in the younger structures. The writer prefers the latter explanation.

Some features within the area of investigation seem to favor one of the proposed concepts over the other, but no factor seems sufficiently definitive to eliminate either as a model. The emplacement of discordant plutons into the cover rocks—suprastructure of Haller—would seem to favor the concept of migmatite domes. On the other hand, the upwelling south of Mount Bigelow and the metasomatic and intrusive contacts common in the study area would be in fundamental agreement with either concept. The fact that the Scanlan Conglomerate in one place is in depositional contact with the Oracle Granite, and in another place, in metasomatic or intrusive contact with the Catalina Gneiss seems to suggest a mantled gneiss dome. However, the relationship of the Oracle Granite and the Catalina Gneiss is not established. Damon and others (1963, page 116) offered
convincing evidence that the Samaniego Granite (Catalina Granite) is Oracle Granite that has been reactivated in the Tertiary. McCullough (1963) also suggested that the Catalina Granite and the Oracle Granite are genetically related, and stated that the contact between the Catalina Granite and Catalina Gneiss is gradational. Recalling that Pilkington found the Apache Group in unconformable contact with banded augen gneiss which he suggests was originally a sedimentary rock, and Banerjee's suggestion that the Oracle Granite is metasomatized Pinal Schist, it seems reasonable to suggest that the early Precambrian basement beneath the Catalina Mountains was part of a migmatite terrain that became the site of sediment accumulation and was reactivated as a mantled gneiss dome in Cretaceous-Tertiary time.
LIST OF REFERENCES


LIST OF REFERENCES—Continued


LIST OF REFERENCES--Continued


Figure 2. GEOLOGIC MAP OF THE MT. BIGELOW–BEAR WALLOW AREA

EXPLANATION

ROCKS

TERTIARY

CRETACEOUS–TERTIARY(*)

CRETACEOUS (I)

STRUCTURE SYMBOLS

- Contact
- Fault
- Decollement
- Thrust Fault

SCALE 1:4,800

CONTOUR INTERVAL 80 FEET

GEOLOGY BY C. J. WAAG, 1967-68
Figure 23. LINEAR STRUCTURES OF THE MT. LEMMON AREA

EXPLANATION

STRUCTURE SYMBOLS
- Geodetic field work, line spacing consists of 1/2, 2, 5
- Fold axis location, angle number indicates number of readings
- Strike and dip of axial plane of fold
- Interpreted fold axes
- Long axis of deformed clast, 20 indicates number of readings
- Lineation, stretched mineral (sm) microcrenulation (me)
- Indicated direction of tectonic transport

SCALE 1:4,800
CONTOUR INTERVAL 60 FEET

ORIENTATION CATEGORIES
- 0-5%
- 5-10%
- 10-15%
- 15-20%
- > 20%

GEOLOGY BY C. J. WAAG
Figure 32. SKETCH OF THE RIDGE NORTHEAST OF TURKEY FLAT
FACING NORTHWEST

BY C. J. WAAG
Figure 31. SKETCH OF THE RIDGE NORTH OF MT. BIGELOW

FACING EAST