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San Bernardino County, California**

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San Jose State University, 1989

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ROCKS AND STRUCTURAL EVOLUTION
OF THE NORTHERN KILBECK HILLS,
SAN BERNARDINO COUNTY, CALIFORNIA

A Thesis

Presented to

The Faculty of the Department of Geology
San Jose State University

In Partial Fulfillment
of the Requirements for the Degree
Master of Science

By

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December, 1989

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ABSTRACT

The Kilbeck Hills, located in the Mojave Desert in southeastern San Bernardino County, California, lie in a region in which Proterozoic rocks of the North American craton, along with Paleozoic and Triassic strata and Jurassic igneous rocks, underwent intense ductile deformation and epidote-amphibolite to upper amphibolite facies metamorphism prior to emplacement of Late Cretaceous plutons. In the Kilbeck Hills the southwest end of one of the major structures in this region, a 45 km-long zone of northwest-dipping fold and thrust nappes, is exposed. In this zone, centered on the Old Woman Mountains, inverted Proterozoic basement and lower Paleozoic strata have been placed over upright, but folded, upper Paleozoic and Triassic(?) strata along a fault known informally as the Scanlon thrust.

The nappe complex in the Kilbeck Hills consists of six low-angle fault-bounded sheets. From structurally highest to lowest these are composed of: 1) inverted lower Paleozoic strata and their depositional basement of Proterozoic gneiss, plus Jurassic(?) granitoids; 2) inverted upper Paleozoic strata; 3) Triassic(?) strata and Jurassic(?) orthogneiss and orthoschist; 4) fault slivers of Paleozoic strata; 5) upright Paleozoic

strata; and 6) orthogneiss and orthoschist of Jurassic and Jurassic(?) age along with gneiss and schist of uncertain age. The fault between sheets 1 and 2 may be equivalent to the Scanlon thrust, but it is uncertain which of the low-angle faults in the Kilbeck Hills has accommodated the greatest amount of movement.

The earliest deformational features preserved in the area are metamorphic foliation, large-and small-scale, tight to isoclinal F1 folds, and a penetrative down-dip stretching lineation plunging northwest, which is subparallel to F1 fold axes. At a late stage in this deformational episode, ductile low-angle faults juxtaposed the various sheets of the nappe complex. Some of these faults were reactivated as brittle faults prior to the development of open to tight, northwest-vergent F2 folds, and open, northwest-plunging F3 folds.

A pluton of undeformed Late Cretaceous biotite monzogranite injects the gneisses and schists of sheet 6 at the base of the nappe zone. Its emplacement predated brittle faulting. A body of Late Cretaceous two-mica granite is exposed extensively in sheet 1, and dikes of similar composition cut the monzogranite pluton.

Mineral assemblages in pre-Late Cretaceous rocks of the area place these rocks in the lower temperature part of the amphibolite facies of regional metamorphism. A single episode of prograde metamorphism was largely

synchronous with the early, ductile deformation and produced diopside in siliceous dolomitic rocks and sillimanite in pelitic rocks of appropriate chemical composition. Field data, combined with published geobarometric and phase equilibrium studies, indicate that peak metamorphic temperatures were within the range of 500-650°C, at pressures of 3-7 kb.

INTRODUCTION

Cordilleran deformation in the southwestern United States ranges from Permian to Recent in age (Suppe, 1985). Tertiary and younger structures have been linked to strike-slip faulting and to varying rates, directions, and angles of subduction along the western margin of North America (Hamilton, 1969; Hamilton, 1986), but the specific tectonic controls on Mesozoic deformation are less certain. It is generally accepted that plate convergence was the dominant large-scale tectonic process in western North America, at least from mid-Jurassic through Cretaceous time. Also, it is known, in a general way, that the style of deformation was influenced by the nature of the basement, the thickness of the Paleozoic cover, and the position relative to the magmatic arc (Burchfiel and Davis, 1981, 1985). However, the specific tectonic processes involved in these style variations, their temporal evolution, and the geometry and kinematics of the resulting structures still are poorly understood.

The Mojave Desert region of southeastern California is an especially important area for understanding the Mesozoic tectonic evolution of the Cordillera. There, the Cordilleran fold and thrust belt and the southwest-trending edge of the North American craton intersect the southeast-trending Mesozoic magmatic arc (fig. 1).

Figure 1. Map of southwestern United States showing geologic boundaries mentioned in text. Numbered lines correspond to the following boundaries: 1) Western limit of Cretaceous plutonism in areas east of the San Andreas fault; 2) Eastern limit of Cretaceous plutonism; 3) Eastern limit of Cordilleran fold and thrust belt; 4) Approximate boundary between Paleozoic cratonal strata on east and miogeoclinal strata on west.

Pre-Cenozoic rocks of the region record a varied and complex Mesozoic history of sedimentation, magmatism, and deformation. Within the craton, in the east-central and southeastern Mojave Desert, there are a number of ranges in which pre-Cretaceous rocks underwent medium- to high-grade metamorphism and intense ductile deformation prior to emplacement of Late Cretaceous plutons. They provide samples of the middle crust as it existed at the end of Mesozoic time, overprinted locally by Tertiary extensional tectonics. One such area is centered on the Old Woman Mountains and the nearby Piute and Little Piute mountains and Kilbeck Hills (fig. 2,3). These ranges contain a 45 km-long zone of northwest-dipping late Mesozoic thrust and fold nappes, which deform rocks at least as young as mid-Jurassic and which are cut by Late Cretaceous plutons.

This report is the result of a structural mapping and petrographic investigation of the northern Kilbeck Hills, where the southwesternmost portion of the nappe zone is exposed. The Kilbeck Hills are especially significant for understanding the Mesozoic history of the region because a large number of the pre-Cenozoic stratigraphic and plutonic units there are exposed in close proximity. Detailed study of the stratigraphic, structural, and intrusive relationships in the Kilbeck Hills yields data on variations in Mesozoic structural styles in the Mojave region and on tectonic processes at mid-crustal levels in

Figure 2. Location map showing study area and nearby geographic features referred to in the text.

Figure 3. Index map of western U. S. and southern California showing the general location of the Kilbeck Hills in relation to physiographic provinces and geographic features mentioned in the text.

continental magmatic arcs.

A knowledge of the regional geology is necessary if the complex rocks and structures of the Kilbeck Hills are to be placed in context. In particular, an understanding of the geology of the Old Woman-Piute range is essential to understanding the observations and interpretations presented in this report. Therefore, the Proterozoic to early Tertiary geology of the Mojave Desert and the Old Woman-Piute range is described in some detail.

Location

The Kilbeck Hills are located in southeastern San Bernardino County, in the Cadiz Lake 15-minute quadrangle, $115^{\circ} 15'$ to $115^{\circ} 30'$ west longitude, and $34^{\circ} 15'$ to $34^{\circ} 30'$ north latitude (fig. 2). They lie approximately 80 km west of the Colorado River, and 6-8 km west of the southern Old Woman Mountains. The Kilbeck Hills are a low northern extension of the Iron Mountains, and form a drainage divide between Cadiz Lake to the west and Danby Lake to the southeast.

The Kilbeck Hills lie 25 km southeast of Cadiz, California, which is 5 km south of Highway 66 (fig. 2). A dirt road extends southeast from Cadiz, paralleling the tracks of the Atchison, Topeka, and Santa Fe Railroad. It skirts the north end of the Kilbeck Hills and continues to the southeast past Danby Lake, intersecting Highway 62 at

a point 8.5 km west of Rice, California. Except after heavy rains, the road from Cadiz is easily traversed by two-wheel-drive vehicles. A number of dirt roads, mostly leading to abandoned marble quarries or mining prospects, cross the study area. The condition of these roads ranges from good to impassable. No place in the study area is more than 1.2 km from a point accessible to a two-wheel-drive vehicle.

The area described in this report lies in the northern Kilbeck Hills and occupies an area of about 18 km². It consists of a series of hills and west- to northwest-trending ridges separated by low alluviated areas. Elevation in the study area ranges from 300 m to 500 m above sea level. Although relief is limited, many of the hills are quite steep, with slope angles locally approaching 40°. Vegetation is extremely sparse and exposure is excellent, especially in areas underlain by Paleozoic carbonate units.

Purpose and Methods

The goals of the study were: 1) to identify and map the various lithologic units and their major and minor structures; and 2) to work out the geologic history by determining, as far as possible, the relative ages of deformation, metamorphism, and plutonism.

Mapping was carried out during the Spring and Fall of

1984 and the Spring of 1985 on 1:6,000-scale enlargements of 1:40,000-scale aerial photographs. Field mapping was transferred by hand to a 1:5,000-scale enlargement of the U.S. Geological Survey Cadiz Lake Quadrangle topographic map (1:62,500 scale). All maps and photographs were supplied by the U.S. Geological Survey, Menlo Park, California. One hundred and five thin sections were prepared by the U. S. Geological Survey and by the author at San Jose State University. Mineral identifications and percentage estimations were made using a petrographic microscope.

Previous Work

Despite their accessibility and varied geology, the Kilbeck Hills have received little study. They were examined in reconnaissance work by Hazzard and Dosch (1937) and were mapped at a scale of 1:125,000 by Bassett and Kupfer (1964). The Kilbeck Hills also were mapped by geologists of the Land Department of the Southern Pacific Company in the period of 1958-1960. Their unpublished 1:24,000-scale mapping was used by Bishop (1963) in compiling the Needles 1° x 2° sheet of the geologic map of California. The northern Kilbeck Hills were mapped jointly by Evenson (1973) and Jones (1973) as parts of Master's thesis projects at the University of Southern California. In 1978, detailed reconnaissance mapping of

the Kilbeck Hills was carried out by Keith Howard and Paul Stone of the U. S. Geological Survey (Howard, unpublished data). Howard and Stone correlated the metasedimentary rocks of the area with the regional Paleozoic sequence (Stone and others, 1983) and were the first workers to appreciate the structural complexity of the northern Kilbeck Hills. The present study is an outgrowth of their work.

REGIONAL GEOLOGIC SETTING

The Mojave Desert occupies about 40,000 km² in southeastern California. It is bounded on the northwest by the Garlock Fault, on the southwest by the Transverse Ranges and the Salton Trough, and on the east by the Colorado River (fig. 3). It is a region characterized, in most places, by internal drainage. Aggrading alluviated basins separate ranges of hills and mountains of widely varying relief and areal extent. At least half of the area of the Mojave is covered by alluvium and playa deposits, with minor volcanic rocks of Quaternary age. Bedrock exposures are composed of, in descending order of abundance: 1) Mesozoic plutonic rocks; 2) Tertiary volcanic and nonmarine sedimentary rocks; 3) Proterozoic crystalline rocks; and 4) Paleozoic and Mesozoic strata.

All of the above rock types are present in the Old Woman-Piute range area, including the Kilbeck Hills and Little Piute Mountains (figs. 4,5). This area exposes a cross section of a portion of the middle crust as it existed at the end of Mesozoic time. The Kilbeck Hills expose a small part of the central area of this crustal cross section. Regional dip is toward the northwest, so that successively deeper structural levels are seen from north to south. The dominant tectonic imprint on the

Figure 5. Simplified geologic map of the Old Woman-Piute Mountains and northern Kilbeck Hills. Geology modified from C. Miller and others (1982).

Old Woman-Piute range is of late Mesozoic age. Proterozoic basement, Paleozoic-Triassic strata, and local Jurassic intrusive rocks have undergone ductile folding and thrusting, and greenschist-to amphibolite facies metamorphism, followed by intrusion and local migmatization of the country rocks by granitic plutons of Late Cretaceous age.

Proterozoic Basement Rocks

Proterozoic crystalline rocks crop out in many ranges in the eastern Mojave Desert, and are exposed extensively in the central and eastern Transverse Ranges (figs. 3,4). The oldest group of rocks includes amphibolites and a variety of metasedimentary rocks, which are intruded by granitoid gneisses. U-Pb and Rb-Sr ages on this early suite range from 1.6 to 1.8 Ga (Burchfiel and Davis, 1981). Rocks of similar age and lithology form a northeast-trending belt traceable from southeastern California to northern New Mexico and Colorado (Silver and others, 1977). In several ranges in the Mojave region the older Proterozoic rocks are intruded by potassic granites and syenites, all yielding radiometric ages of approximately 1.4 Ga (Anderson, 1983). They are interpreted as a southwestern extension of the continent-wide, middle to late Proterozoic anorogenic magmatic belt (Emslie, 1978).

Proterozoic metaplutonic and metasedimentary rocks are exposed through much of the Old Woman-Piute range. The most abundant unit is the Fenner Gneiss, a variably deformed augen gneiss of granodiorite to granite composition (Bender and Miller, 1987). Wooden and others (1988) reported U-Pb zircon data which indicate that the Fenner Gneiss is part of a major intrusive episode between 1683 and 1672 Ma. Also abundant locally are pelitic and psammitic metasedimentary rocks which are intruded by the Fenner Gneiss. Other Proterozoic rock types include amphibolite and various granitoid gneisses, including two-mica ± garnet granite. Diabase dikes, correlated by Fitzgibbon and Howard (1987) with a regional 1.1-1.2 Ga dike swarm, are present in the Piute Mountains. In places they cut deformed Fenner gneiss but are themselves undeformed (Howard and others, 1987). Xenoliths from Tertiary dikes in the Piute Mountains have granulite facies assemblages whose chemistry indicates high pressure metamorphism (Staude and Miller, 1987). These data suggest that Proterozoic rocks in the region underwent pre-1.1 to 1.2 Ga deformation and high-grade metamorphism.

Paleozoic Strata

Paleozoic sedimentary rocks, despite limited areas of exposure, provide valuable information about the

Phanerozoic history of the Mojave region. Paleozoic strata of the eastern and south-central Mojave Desert are a southwestward continuation of the cratonal and miogeoclinal facies belts of the Basin and Range province and the Colorado Plateau (figs. 1,3). Cratonal facies rocks can be traced southwestward to the edge of the North American craton in the San Bernardino Mountains (fig. 4). The fact that miogeoclinal rocks are not found outboard of cratonal rocks at the edge of the craton has led most workers (e.g., Burchfiel and Davis, 1981) to conclude that the Paleozoic-Early Triassic continental margin was truncated in Early Mesozoic time.

Paleozoic eugeoclinal rocks are present in the Goldstone-Lane Mountain area of the north-central Mojave (fig. 4), and in the El Paso Mountains north of the Garlock fault (Miller and Sutter, 1982). These deformed and metamorphosed strata, the depositional basement of which is nowhere exposed, contain coarse clastic and volcanic rocks which have no counterparts in coeval shelf and platform facies rocks to the east (Miller and Sutter, 1982). These rocks have been interpreted as being allochthonous with respect to Paleozoic rocks elsewhere in the Mojave, their emplacement postdating deposition of Permian strata and predating Permian-Triassic deformation and plutonism (Miller and Sutter, 1982).

A large area of the southeastern Mojave Desert lacks coherent fossiliferous sections of Paleozoic strata. Between the Marble and Ship mountains on the northwest and the Plomosa Mountains of Arizona on the southeast (fig. 4), all pre-Tertiary sedimentary rocks have undergone significant deformation and metamorphism. In early reconnaissance studies most of these metasedimentary rocks were interpreted as Precambrian in age. Marble, quartzite, and schist of the Kilbeck Hills were included with similar rocks of the Old Woman-Piute range in the Precambrian Essex Series by Hazzard and Dosch (1937). Miller (1944) assigned a Paleozoic age to metasedimentary rocks of the Big Maria, Little Maria, Riverside, and Palen mountains (fig. 4), but did not make any specific correlations with Paleozoic units in the region. That such correlations were possible at the formational level was recognized by Hamilton (1964). However, detailed regional correlations were not established until the late 1970's, when the stratigraphy of the Big Maria Mountains (Hamilton, 1982) and ranges of the Needles 1° x 2° quadrangle were worked out (Stone and others, 1983). These authors demonstrated that the metasedimentary sequences have a consistent succession of rock types from range to range, and correlated these sequences with the cratonal Paleozoic section of the Grand Canyon in Arizona (Stone and others, 1983).

Rocks now interpreted as Paleozoic are present at three structural levels in the Old Woman-Piute range area. In the Piute Mountains a north- to northwest-striking sequence of Cambrian to Mississippian strata, which has undergone intense ductile folding and thrust faulting (Fletcher and Karlstrom, 1987), is recognized (fig. 5). Metamorphic assemblages in Paleozoic pelitic rocks of the Piute Mountains range from staurolite to andalusite grade (C. Miller and others, 1982).

The second and most extensive area of Paleozoic rocks occurs in a belt, rarely more than 1 km wide, stretching southwest from the Little Piute Mountains to Scanlon Gulch and Meteor Gulch on the west side of the Old Woman Mountains (fig. 5). This belt exposes the inverted limb of a basement-cored nappe (C. Miller and others, 1982), which has been thrust over upright, but locally folded, middle and upper Paleozoic and Triassic strata. The thrust has been given the informal name "Scanlon thrust" by workers in the Old Woman-Piute range (Howard and others, 1987). Metamorphic assemblages in Paleozoic pelitic rocks of the belt range from andalusite grade in the Little Piute Mountains, to sillimanite-potassium feldspar grade in the Old Woman Mountains (C. Miller and others, 1982). Similarities of structure, rock types, and metamorphic grade suggest that the Paleozoic rocks of the Kilbeck Hills are a southwestward

continuation of this belt.

The third and structurally lowest group of Paleozoic rocks is located at the mouth of Brown's Wash in the southwestern Old Woman Mountains (fig. 5). There, marble now interpreted as Paleozoic occurs as slivers floating in migmatitic gneiss on the flanks of the range and in three isolated hills at the foot of the range (Jones, 1973). C. Miller and others (1982) interpreted the Paleozoic rocks of Brown's Wash as remnants of an infold of Paleozoic strata at a structural level below that of the main nappe complex exposed to the north at Scanlon Gulch.

Mesozoic Rocks

Mesozoic Strata

In the central and eastern Mojave Desert sandy limestones of the Moenkopi Formation locally rest paraconformably on Permian strata (Burchfiel and Davis, 1981). The lower part of the Moenkopi Formation is marine, but the upper part of the unit and all overlying strata are nonmarine. This sequence, capped by the Lower Jurassic Aztec Sandstone, is locally overlain unconformably by Jurassic volcanic rocks and, in the northeastern Mojave, by orogenic conglomerates of Jurassic and Cretaceous age (Burchfiel and Davis, 1981).

Regionally, lower Mesozoic strata show similarities

in rock types but differ widely in thickness. In the Big Maria Mountains (fig. 4), Hamilton (1982) described meta-conglomerates and greenish schists which lie with general structural concordance on Paleozoic rocks. The schists and conglomerates are overlain by Aztec-equivalent quartzites and Jurassic meta-ignimbrites. In the Riverside Mountains, Paleozoic strata are overlain by gypsiferous phyllite and by sericitic quartzite, correlated tentatively by Lyle (1982) with the Jurassic Aztec Sandstone. Two thick sections interpreted by Emerson (1982) as Mesozoic in age are present in the Little Maria Mountains. Directly overlying Paleozoic rocks is a 400-m-thick section of green quartzite, conglomerate, and meta-tuff. Also present is an overturned, tectonically-emplaced sequence containing 1,000 m of green, fine-grained schist. In the Arica Mountains, Baltz (1982) reported that the Kaibab Limestone is overlain by slightly metamorphosed sandstone, siltstone, and volcaniclastic rocks. Laminated calc-silicate schists of varying thickness overlie the metamorphosed Kaibab Limestone in the Old Woman Mountains and Little Piute Mountains. The schists have been correlated by Stone and others (1983) with the Triassic Moenkopi Formation.

Mesozoic Igneous Rocks

Mesozoic igneous rocks in the Mojave Desert are a part of the northwest-trending Cordilleran magmatic arc (fig. 1). Available geochronologic and petrologic data indicate that Mesozoic plutons occur in distinct spacial, temporal, and lithologic groups (John, 1981).

Triassic(?) Igneous Rocks. The oldest arc plutons probably are of Triassic age, although they may be as old as Permian. They have been identified only near the southwestern and northwestern boundaries of the Mojave Desert (John, 1981).

Igneous rocks of Middle and Late Triassic age are rare in the Mojave Desert. Volcanic rocks of this age have been identified in the Soda Mountains (Grose, 1959), at Old Dad Mountain (Dunne, 1977), and north of the Garlock Fault in the Inyo Mountains (Merriam, 1963) and the Panamint Range (Johnson, 1957).

Jurassic Igneous Rocks. Early Jurassic time was a period of quiescence in the Mojave Desert portion of the magmatic arc (Burchfiel and Davis, 1981). Continentally-derived sandstones equivalent to the Lower Jurassic Aztec Sandstone are recognized as far west as the Rodman Mountains (Miller and Carr, 1978) and as far south as the Palen Mountains (LeVeque, 1982) (fig. 4).

Arc activity resumed in Middle Jurassic and continued into Late Jurassic time. Rocks of this age include plutons and local hypabyssal and volcanic rocks. The Kilbeck Hills lie near the northeastern boundary of abundant Middle and Late Jurassic plutonism.

Rocks of the Jurassic magmatic arc are exposed extensively in ranges to the southwest, west, and northwest of the Kilbeck Hills. Howard and John (1984) described the Jurassic intrusive rocks of the Ship, Calumet, Sheep Hole and eastern Bullion mountains (fig. 4). They considered the most characteristic rock type of the area to be porphyritic sphene-hornblende-biotite quartz monzonite, monzogranite, or granodiorite, with lavender, gray, or cream-colored unzoned orthoclase phenocrysts, and clotted mafic minerals. D. Miller and others (1982) described the complex Jurassic igneous rocks of the Bristol Mountains. They reported that Jurassic plutonic rocks in the range intrude their own hypabyssal plutons and volcanic ejecta, and that hypabyssal Jurassic rocks are also present in the Bullion Mountains and Ship Mountains. From these relationships, and the fact that pluton contacts are discordant, D. Miller and others (1982) concluded that the Jurassic magmatic arc in the region is largely epizonal.

In general, Jurassic plutonic rocks show a

northeastward decrease in quartz content across the area from the Sheep Hole and Bullion mountains (fig. 4) to the Ship and Marble mountains (John, 1981). Rocks of the Ship Mountains and Marble Mountains are mostly syenites, monzonites, and quartz monzonites, whereas those of the Bullion Mountains and Sheep Hole Mountains are monzogranites and granodiorites with a greater tendency toward clotted mafic minerals and lavender potassium feldspar phenocrysts.

Cretaceous Plutonic Rocks. Cretaceous arc activity extended beyond the northeastward limits of Jurassic arc activity into southern Nevada and western Arizona (fig. 1). Virtually all of the Cretaceous igneous rocks of the Mojave region are plutonic. To the north and east of the Kilbeck Hills there are discrete plutons, whereas to the south and west composite batholiths developed (John, 1981). Late Cretaceous ages predominate over most of the region. Over large areas of the Mojave Desert and the adjacent Transverse Ranges, K-Ar ages of Proterozoic and earlier Mesozoic plutonic rocks were partially to completely reset during Late Cretaceous time (Miller and Morton, 1980).

Two distinct groups of late Cretaceous plutons are present in the Old Woman-Piute range area (fig. 5). The older group is exposed in three plutons with a total

area of 200 km², and is composed of metaluminous hornblende-biotite granodiorite (C. Miller and others, 1982). By far the largest of the plutons is the Old Woman pluton, a sheet-like body which concordantly intrudes the Paleozoic nappe zone from below in the Old Woman Mountains and Little Piute Mountains. The pluton is strongly foliated near its roof, where flattened mafic inclusions and schlieren parallel the contact. It is bounded above and below by layered, in part schistose, Precambrian(?) and (or) Jurassic(?) gneiss. Migmatites derived from Precambrian granitic rock occur beneath the layered gneiss. These are invaded by sheets and tongues, up to 100 m thick, of the Old Woman pluton (C. Miller and others, 1982). Foster and others (1987) reported that biotite ⁴⁰Ar/³⁹Ar data combined with U-Pb zircon data of J. E. Wright indicate an age of 71 Ma for the Old Woman pluton.

The younger group of five late Cretaceous plutons in the Old Woman-Piute range is composed of peraluminous two mica ± garnet granite, exposed over a total area of 160 km². In most places the two-mica granites intrude discordantly, but the Sweetwater Wash pluton of the northern Old Woman Mountains intrudes the top of the Paleozoic nappe complex concordantly and is moderately deformed. It cuts both the low-angle faults of the nappe zone and the underlying Old Woman pluton (C. Miller

and others, 1982). Concordant biotite, muscovite, and potassium feldspar $^{40}\text{Ar}/^{39}\text{Ar}$ ages reported by Foster and others (1987) indicate an emplacement age of 70-73 Ma for the two-mica granite bodies of the Old Woman-Piute range.

Rocks of the eastern edge of the late Cretaceous Cadiz Valley batholith underlie the southern Kilbeck Hills and extend southward to the Iron Mountains (Howard and others, 1982). This batholith is an elongate, northwest-trending plutonic complex which covers an area of at least $1,700 \text{ km}^2$ (John, 1981; D. Miller and others, 1982). It is composed of two intrusive phases. The older is variably foliated, hornblende-bearing, biotite-sphene granodiorite which intrudes northwest-striking, steeply-dipping wallrocks and contains inclusions of Jurassic intrusive rocks, Paleozoic and Triassic strata, and locally, Proterozoic gneiss. Younger plutons are variably porphyritic granodiorite to monzogranite containing 2-11% biotite and up to 3% muscovite. Rare, late stage garnet, two-mica aplite and pegmatite dikes cut the plutons. Biotite and muscovite K-Ar ages from the batholith range from 55-69 Ma (Armstrong and Suppe, 1973; Calzia and Morton, 1980).

Mesozoic Deformation

Although Mesozoic structures are widespread in the Mojave Desert, the Mesozoic history of the region is imperfectly understood. Dated structures range from Permo-Triassic to Late Cretaceous in age, suggesting that deformation continued, at least sporadically, throughout the Mesozoic. However, structures of a given age may be widely separated, and their relationships to one another and to the large-scale tectonics of the region often are obscure. Because of the scarcity of post-early Triassic-pre-Miocene strata in the region, Mesozoic deformation rarely can be dated by overlapping sedimentary sequences. Also, the age of the youngest deformed strata is of limited value in determining the timing of deformation in most places. Therefore, dating has relied primarily on the relationships of structures to Mesozoic plutons. Where intrusive rocks are absent, Mesozoic deformation often cannot be dated. Where plutons are abundant, however, they commonly have obliterated much of the evidence of the earlier geologic history.

Despite these difficulties, several generalizations can be made about Mesozoic deformation in the region:

- 1) overall, deformation appears to be dominated by east-vergent folds and thrust faults; 2) local Mesozoic high-angle faults suggest extension and/or strike-slip

movement during, or between episodes of thrust faulting (Burchfiel and Davis, 1981); 3) Mesozoic thrust faults in the Mojave Desert, especially those of the Old Woman-Piute range and the Kilbeck Hills, show greater involvement of basement rocks and a more ductile style of deformation than those to the north in southern Nevada and Utah; 4) the location of the eastern margin of deformation, which has a similar pattern to movements of the eastern margin of magmatic arc activity (Burchfiel and Davis, 1981), shifted eastward over time.

From the Las Vegas area northward, the Cordilleran thrust belt follows the hingeline between cratonic strata on the east, and miogeoclinal strata on the west (fig.1). The location of thrust faults apparently is determined by anisotropy in the westward-thickening sedimentary wedge, with decoupling of cover rocks from the basement. Southwestward, the Cordilleran fold and thrust belt intersects the Mesozoic magmatic arc, and the controls on thrust fault localization change. From Winter's Pass near the California-Nevada line to the New York Mountains (fig. 4) the thrusts retain their "thin-skinned" geometry but show less parallelism to upper plate strata and involve slices of basement rock, with upper and lower plates sharing a common mylonitic foliation (Burchfiel and Davis, 1985). Farther south, in the region from the Old Woman-Piute range to the Big

Maria Mountains, the style of deformation is strongly ductile, with basement-cored nappes, greenschist to amphibolite facies metamorphism, and often extreme attenuation of Paleozoic strata.

In their review of Mesozoic deformation in the Mojave region, Burchfiel and Davis (1981) proposed that the coincidence of the eastern limits of arc magmatism and east-vergent thrust faulting reflected westward underthrusting of the cold, rigid crust of the craton beneath the relatively warm, ductile magmatic arc. In this view, the arc-craton ductility contrast controlled the location of thrusting, which moved eastward as the eastern limit of the arc migrated over time. An alternative interpretation of late Mesozoic tectonism in the eastern Mojave has been proposed by Howard (1986). He noted that structural reconstructions indicate that the top of the Mesozoic batholith as a whole tapers downward to the northeast beneath a west-thinning wedge of cooler, more brittle cratonic crust at the edge of the Colorado Plateau. He suggested that the southwest vergence of some Mesozoic folds and thrusts, including a large anticlinorium in the Maria-Palen Mountains-McCoy basin area (fig. 4), could be explained by ductile northeastward underthrusting of the orogen beneath the overlying wedge of cratonic crust.

Early Tertiary Uplift

There are no sedimentary or volcanic rocks of known Paleocene or Eocene age in the Mojave Desert between the Garlock and San Andreas faults. Early Tertiary time was a period of prolonged uplift and erosion throughout the region. In most places the Mesozoic and Paleozoic volcanic and sedimentary cover was removed, so that Oligocene(?) and younger deposits rest directly on Mesozoic plutons and Precambrian basement rocks.

Radiometric data from the Old Woman-Piute range indicate that the range was uplifted in latest Cretaceous and early Tertiary time. Foster and others (1987) reported that two-mica granite plutons of the range, with emplacement ages of 70-73 Ma, contain potassium feldspars which were nearly closed to Ar loss (at 200-270°C) by 67 Ma. They also reported apatite fission track ages of 65-70 Ma for the Piute and northern Old Woman mountains and 30-50 Ma for the southern Old Woman Mountains. Foster and others (1987) concluded that the range was rapidly uplifted in latest Cretaceous time and that the Piute Mountains and northern Old Woman Mountains were within a few kilometers of the surface by 65 Ma.

LITHOLOGY AND STRATIGRAPHY

The northern Kilbeck Hills expose a variety of crystalline rocks, ranging from Proterozoic to Late Cretaceous in age (fig. 6). Age assignments and correlation of these rocks with units in nearby ranges are based on lithologic similarities and on depositional and intrusive relationships. The specific formation-level regional correlations of metamorphosed Paleozoic strata proposed by Stone and others (1983) are especially important in understanding the structural geometry of the Kilbeck Hills. Facing directions and excision or duplication of Paleozoic strata, along with fault-related structural features, indicate that the rocks of the Kilbeck Hills are arranged in a complex stack of thin fold and thrust nappes, similar in structural sequence to that in the Scanlon thrust zone of the Old Woman Mountains and Little Piute Mountains. These nappes in the Kilbeck Hills can be grouped into five low-angle fault-bounded rock "packages", or zones (fig. 7). Four of the five zones are themselves cut by one or more low-angle faults. The zones are identified by their distinctive rock types and (or) facing directions, and, in one case, by a lack of stratigraphic coherence. From structurally highest to lowest, these zones are:

Figure 6. Column showing relative ages of stratified and intrusive rocks in the northern Kilbeck Hills.

Age	Unit Name Dominant rock type (mafic mineral suite)
Late Cretaceous	Two-mica granite two-mica ± garnet granite
	Leucocratic monzogranite magnetite-biotite leuco-monzogranite
	Biotite monzogranite Biotite monzogranite to granodiorite
Jurassic and Jurassic(?)	Granite gneiss (biotite)
	Megacrystic monzogranite gneiss (biotite ± hornblende)
	Dark granodiorite gneiss (hornblende ± biotite)
Jurassic(?) and Triassic(?)	Diorite (hornblende)
	Gneiss, schist, and quartzite
Triassic(?)	Siliceous marble and calcareous quartzite
	Garnet-mica schist
Mesozoic	Migmatitic gneiss layered gneiss, augen gneiss, paraschist

Figure 6, cont.

Age	Unmetamorphosed Paleozoic equivalent	Unit name (rock types)
Permian	Kaibab Limestone	(dolomitic calcite marble)
	Coconino Sandstone	(micaceous quartzite)
	Hermit Shale	(calc-schist)
Permo- Pennsylvanian	Bird Spring Formation	(marble, calc- schist, quartzite)
Mississippian	Redwall Limestone	(calcite marble)
Devonian and (or) Cambrian		dolomite marble
Cambrian	Lower part of the Bonanza King Formation	(calcite marble)
	Bright Angel Shale	(mica schist)
	Tapeats Sandstone	(quartzite)
Proterozoic		syenogranite gneiss
		augen gneiss

Figure 7. Tectonic map of the northern Kilbeck Hills, showing fault-bounded lithostructural zones referred to in the text. Lithostructural zones shown on the map are listed below from structurally highest to lowest:

Inverted Paleozoic zone - Inverted Proterozoic basement (black) and Paleozoic metasedimentary strata (orange). Contains a major ductile low-angle fault separating inverted lower Paleozoic from inverted upper Paleozoic rocks. The zone is intruded by pre-tectonic Jurassic and Jurassic(?) granitoids and by post-tectonic Late Cretaceous two-mica granite.

Gneiss and Schist zone - Contains mostly fault-bounded strips of varying thickness and lateral continuity, composed of Triassic(?) metasedimentary strata and Jurassic(?) orthogneiss and orthoschist. Also contains rare fault slivers of Paleozoic carbonate strata.

Chaotic zone - Contains fault slivers of widely varying size, composed of Paleozoic metasedimentary strata and quartz-plagioclase-mica schist of unknown age. Locally present at the top of the zone are coherent, upright strips of upper Paleozoic strata.

Upright Paleozoic zone - A coherent upright slice of Paleozoic metasedimentary strata containing rocks of Cambro-Devonian through Permian age.

Migmatitic Gneiss zone - Augen gneiss layered gneiss, and granite gneiss, in part of Jurassic and Jurassic(?) age. Contains ductile-fault slivers of upright upper Paleozoic strata. The lower part of the zone is injected by Late Cretaceous monzogranite.

Intrusive rocks

Pre-tectonic Jurassic and Jurassic(?) intrusive rocks.

Post-tectonic Late Cretaceous plutons consisting of biotite monzogranite beneath the migmatitic gneiss zone, and two-mica granite intruded into the inverted Paleozoic zone.

1. Inverted Paleozoic Zone - This zone consists of inverted Paleozoic metasedimentary strata, overlain by their depositional basement of Proterozoic orthogneiss. The Proterozoic and Paleozoic rocks are intruded extensively by variably deformed Jurassic and Jurassic(?) plutonic rocks and by undeformed Late Cretaceous two-mica granite.
2. Gneiss and schist zone - Throughout its strike length this zone contains orthogneiss and orthoschist derived from Jurassic(?) protoliths, and local calcareous quartzite and paraschist. Along most of its length the zone also contains garnet-mica schist of Triassic(?) protolithic age, overlain at the southwestern end of the zone by siliceous marble and quartzite.
3. Chaotic zone - The chaotic zone consists of tectonic slivers from outcrop- to hillside-scale mostly made up of Paleozoic rocks. Locally, discontinuous slivers of upright, coherent, upper Paleozoic metasedimentary strata occur at the top of the zone.
4. Upright Paleozoic zone - Beneath the chaotic zone is a continuous strip of upright Paleozoic metasedimentary strata. Included in the zone are all

units from the Cambro-Devonian dolomite marble through the metamorphosed Kaibab Limestone.

5. Migmatitic gneiss zone - This zone consists of augen gneiss, granite gneiss, layered gneiss, orthoschist, and paraschist. Some of the rocks correspond to those of Jurassic(?) units in the overlying zones, but others are of uncertain age. Also present are tectonic inclusions of upright upper Paleozoic metasedimentary strata. Most of the zone is injected by biotite monzogranite of Late Cretaceous age.

Proterozoic Rocks

Augen Gneiss

Granodiorite to syenogranite augen gneiss of Proterozoic age is exposed extensively in the upper plate of the nappe complex. It forms the depositional basement of the Paleozoic metasedimentary sequence, and is intruded by Proterozoic, Jurassic, Jurassic(?), and Cretaceous plutonic rocks. The unit is correlative with the Fenner Gneiss of Hazzard and Dosch (1937) which is exposed to the north in the Old Woman-Piute range and in the Fenner Hills. No published radiometric ages for the gneiss are available from the Kilbeck Hills, but Wooden and others (1988) included the Fenner Gneiss in a major intrusive suite having U-Pb zircon ages of 1683-1672 Ma.

The dominant rock type of the unit is coarse- to medium-grained, dark brown-weathering, black and light gray biotite augen gneiss ranging from granodiorite to monzogranite in composition (fig. 8). White to pale gray microcline augen are up to 10 cm in longest dimension, and range from lenticular, to round, to blocky euhedral in shape. Deformational fabrics are variably developed in the rock. Well-foliated areas show coarse, wavy bands defined by lenticular augen and elongate quartz-feldspar aggregates, up to 1 cm thick, alternating with elongate, but discontinuous trains of biotite. Less deformed parts of the gneiss have blocky augen, often randomly oriented, with fabrics defined by irregular clots of oriented biotite. One thin section of the black and gray gneiss was examined. It contains tight, wavy aggregates of biotite separating large, irregularly-shaped crystals and smaller polygonal aggregates of felsic minerals. The rock consists of plagioclase (35%) of An 36 composition (determined by the "a"-normal method), microcline (28%), quartz (20%), and biotite (15%), with accessory magnetite, sphene, and apatite, and trace amounts of zircon and allanite.

Locally crosscutting the black and gray gneiss is a more leucocratic, coarse- to medium-grained granite gneiss containing pink, lenticular microcline augen up to 8 cm in length. The granite gneiss is pink and gray

Figure 8. Strongly deformed Proterozoic augen gneiss. Note lenticular microcline augen and minor F1b folds overturned to the left (southwest).

with thin streaks of black biotite, weathering to a light brown color. The rock is strongly foliated and was observed only in the more highly deformed portions of the unit. One sample of the rock was examined in thin section. It consists of perthitic microcline (40%), coarsely sutured quartz (25%), myrmekitic plagioclase (22%) of An 25 composition (determined by the "a"-normal method), and biotite (10%). Accessory muscovite (2%) is mostly secondary, but a few euhedral crystals intergrown with biotite may be primary. Trace minerals are magnetite, zircon, and apatite.

Monzogranite Gneiss

Monzogranite to syenogranite gneiss of Proterozoic age is exposed in the upper plate of the nappe complex. It cuts the Proterozoic augen gneiss and is overlain nonconformably by quartzite of the metamorphosed Tapeats Sandstone.

The unit is composed of pink and gray, light- to medium-brown-weathering, quartz-feldspar-biotite gneiss. Though variable in texture and fabric, the gneiss shows a narrow compositional range and may be recognized in outcrop by the presence of pink, commonly megacrystic microcline and abundant large quartz grains. Grain size ranges from medium to coarse. Megascopic deformational fabrics tend to vary with those of the augen gneiss

country rock. The least deformed gneisses show only a preferred orientation of biotite, which is evenly distributed through the rock. In the most highly deformed areas, the rock is a strongly foliated augen gneiss in which thin, discontinuous biotite aggregates wrap around lenticular potassium feldspar augen and trains of coarse quartz and feldspar.

The gneiss is of monzogranite to syenogranite composition. A sample examined in thin section consists of perthitic microcline (35%), commonly myrmekitic plagioclase of An 25 composition (determined by the "a"-normal method), quartz (25%), and biotite (11%). Accessory muscovite (3%) has a similar occurrence to that in the leucocratic variant of the augen gneiss unit. Most crystals are secondary, with a few larger primary(?) crystals intergrown with biotite. Trace minerals are magnetite, allanite, apatite, and zircon. Although this rock is similar in composition to the leucocratic granite gneiss of the augen gneiss unit, the two rocks are easily distinguished in the field. In general, the monzogranite gneiss is finer grained, less strongly deformed, and considerably more abundant than the older leucocratic augen gneiss.

Paleozoic Metasedimentary Rocks

The published Paleozoic lithostratigraphic cor-

relations of Stone and others (1983) were used in mapping the metasedimentary rocks of the study area. Paleozoic formational names are used here as a convenient means of identifying the age and stratigraphic position of an interval in a sequence of metasedimentary strata. Structural disruption and metamorphism of the rocks prevent any comparison of thickness or fossil content with undeformed and unmetamorphosed Paleozoic units in nearby ranges.

Table 1 shows a composite Paleozoic stratigraphic column for the Kilbeck Hills. Where both upper and lower stratigraphic contacts of a unit are preserved in the same outcrop, maximum and minimum tectonic thicknesses are given. Units for which no outcrop preserves both upper and lower stratigraphic contacts have only maximum exposed thicknesses listed. Minor differences exist between the unit identifications of Stone and others (1983) and those in this report. The Cambrian(?) white marble unit of Stone and others (1983) appears to be absent in the Kilbeck Hills. Outcrops near Chubbuck in section 10, T3N, R16E, mapped as Cambrian(?) white marble by Stone and others (1983, fig. 5), more recently have been interpreted as metamorphosed Mississippian Redwall Limestone (Brown, 1984; Stone and others, 1984). On the same hill, rocks mapped as metamorphosed Hermit Shale by Stone and others (1983, fig. 5) are here

Table 1.

Summary of lithologic units in metamorphosed
Paleozoic section, Kilbeck Hills

Age	Grand Canyon equivalent	Providence, Marble and Ship Mountains equivalent	Thickness (m) in study area
	Kaibab Limestone		19 (nature of upper contact uncertain)
P	Coconino Sandstone		16
	Hermit Shale		5
P	Supai Group	Bird Spring Formation	maximum = 157 (fault bounded) Complete sections as thin as 14 m
M	Redwall Limestone	Monte Cristo Limestone	6-19
D(?) C(?)		Upper part of Bonanza King Formation in part (?)	Maximum = 46, no complete section is present
	Muav Limestone	Lower part of Bonanza King Formation	11-47
C	Bright Angel Shale	Latham Shale	19-84
	Tapeats Sandstone	Tapeats Sandstone (Wood Canyon Formation + Zabriskie Quartzite)	56-90

interpreted as being Triassic(?) in age.

A complete, unfaulted Paleozoic sequence is nowhere exposed in the study area, but the lithologic succession can be pieced together by combining the partial sequences of the inverted and upright Paleozoic zones. At the top of the nappe complex, the inverted Paleozoic zone contains all Paleozoic units except the metamorphosed Mississippian Redwall Limestone (fig. 7, plate 1). It exposes a continuous inverted lower Paleozoic sequence including the metamorphosed equivalents of the Tapeats Sandstone, the Bright Angel Shale, the lower part of the Bonanza King Formation, and the Cambro-Devonian dolomite unit. Throughout the zone, this inverted sequence is underlain along a ductile low-angle fault by the metamorphosed Permo-Pennsylvanian Bird Spring Formation. Exposed locally beneath the metamorphosed Bird Spring Formation are inverted Permian strata of the metamorphosed Hermit Shale, Coconino Sandstone, and Kaibab Limestone.

The upright Paleozoic zone is composed of the Cambro-Devonian dolomite unit and the overlying metamorphosed Redwall Limestone, Bird Spring Formation, Hermit Shale, Coconino Sandstone, and Kaibab Limestone. The latter three units are preserved only locally along the ductile fault at the top of the upright Paleozoic zone.

The chaotic zone contains slivers of all Paleozoic units except the metamorphosed Bright Angel Shale and the cross-bedded quartzite subunit of the metamorphosed Tapeats Sandstone. The most abundant rock types in the zone are white calcite marble correlated with the metamorphosed Redwall Limestone, and tan dolomite marble correlated with the Cambro-Devonian dolomite unit. At three places, thin, coherent, upright sections containing Permian rocks of the metamorphosed Hermit Shale, Coconino Sandstone, and Kaibab Limestone are present at the top of the chaotic zone.

Upper Paleozoic rocks are present in the migmatitic gneiss zone as tectonic inclusions. The unit most commonly exposed is the metamorphosed Bird Spring Formation, overlain in places by the metamorphosed Hermit Shale, Coconino Sandstone, and Kaibab limestone. Wherever stratigraphically adjacent Paleozoic units are exposed in the gneiss, they are upright.

Metamorphosed Tapeats Sandstone

The base of the Paleozoic sequence comprises quartzite, minor conglomerate, and mica schist correlative with the Tapeats Sandstone of Cambrian age. With the exception of one small outcrop in the chaotic zone, the Tapeats is present only in the inverted Paleozoic zone. In the Marble Mountains, 26 km to the northwest, the

Tapeats Sandstone is divisible into two subunits that are equivalent to the Wood Canyon Formation and Zabriskie Quartzite of the southern Great Basin (Stewart, 1970). The same subunits are present in the Kilbeck Hills, and are described separately here.

Gray, Cross-Bedded Quartzite. Rocks equivalent to the Wood Canyon Formation are composed predominantly of brown-weathering, gray, cross-bedded quartzite. Cross-bedding is present in most outcrops and consists of tabular- to wedge-shaped sets 10-60 cm thick. The cross-bedded strata indicate younging to the southeast, confirming that the northwest-dipping rocks of the upper plate are inverted. Thickness of the "Wood Canyon" subunit ranges from 38 to 61 m.

Brown-weathering, gray quartzite with cross-strata defined by concentrations of dark minerals makes up approximately 90% of the subunit. Grain size, which may vary markedly from one cross-stratum to the next within a hand specimen, ranges from fine to coarse. Interlayered with the quartzite are fine-grained mica schists in layers up to 1 m thick. Interbeds of conglomeratic quartzite containing pebble- to cobble- and rare boulder-size clasts in a matrix-supported fabric occur in the lowermost 6 m of the subunit. The most common clast type is massive white quartz. Also present are clasts of augen gneiss and monzogranite gneiss.

One thin section of cross-bedded quartzite was examined. The rock consists of quartz (78%), microcline (12%), muscovite (5%), biotite (2%), plagioclase (1%), and magnetite (1%) with trace amounts of apatite, zircon, and sphene. Muscovite and magnetite are concentrated in 1- to 2-mm-thick laminae at the base of cross-strata.

The Tapeats Sandstone lies nonconformably on Proterozoic augen gneiss and monzogranite gneiss in the inverted Paleozoic zone. In most places, foliation in the basement gneisses parallels the contact and the layering in the overlying quartzite, but locally foliation in the gneiss is truncated at angles up to 30°. Up to 1.5 m of local relief is present on the erosion surface. Conglomeratic and non-conglomeratic quartzites lie directly on the basement rocks in some places, but elsewhere they lie on a coarse-grained, poorly-foliated, quartz-feldspar-muscovite-biotite schist. This rock is up to 3m in thickness, and grades downward into well foliated gneiss. Locally, discontinuous layers of the schist occur within the basal 2 to 3 m of the quartzite. Stone and others (1983) described quartz and feldspar augen schists at the base of the metamorphosed Tapeats Sandstone of the Big Maria Mountains and Old Woman Mountains that they interpreted as a fossil soil. The dominance of quartz and feldspar in the schists of the

Kilbeck Hills indicates that a more accurate term might be "meta-grus". The schist beds interlayered in the quartzites may be flood or mass movement deposits transported short distances from exposed highs on the erosion surface.

White quartzite. Rocks equivalent to the Zabriskie Quartzite consist of white, red-brown-weathering quartzite and minor interlayered mica schist. The quartzite is massive to thinly-layered, with layering defined by micaceous partings. Layer thickness ranges from 1 cm to 1 m. Interlayered schists range from a few millimeters to 0.5 m in thickness. Total thickness of the subunit ranges from 18 to 29 m. Where exposed, the contact with the underlying cross-bedded quartzites is gradational, with massive, gray "Wood Canyon" quartzites grading upward to clean, white Zabriskie-equivalent rocks over a thickness of 2 to 3 m.

The dominant rock type is pure, white, fine- to medium-grained quartzite. No feldspar is visible in hand specimen and micas are confined to parting surfaces. The only accessory mineral visible in hand specimen is magnetite in the form of subhedral to euhedral porphyroblasts up to 3 mm in size. Interlayered mica schists are fine- to medium-grained and consist of muscovite, biotite, quartz, and feldspar.

Metamorphosed Bright Angel Shale

A unit consisting of pelitic schist with minor marble and calcareous quartzite overlies the metamorphosed Tapeats Sandstone. The unit is correlated with the Cambrian Bright Angel Shale of the Grand Canyon region (Stone and others, 1983) and with the Cambrian Latham Shale of the Marble Mountains (Stewart, 1970). The Chambless Limestone and the Cadiz Formation, which overlie the Latham Shale in the Marble Mountains, apparently are missing in the Kilbeck Hills. The schists of the metamorphosed Bright Angel Shale range from 19 m to 88 m in tectonic thickness, and are present only in the upper plate of the nappe complex.

Most of the Bright Angel Shale (90-95%) is made up of fine- to medium-grained mica schist containing muscovite, biotite, quartz, and feldspar. Euhedral to subhedral magnetite porphyroblasts up to 4 mm in size are present locally.

Interlayered in the lower 5-8 m of the schists are yellow-brown, medium-grained, laminated calcite marble layers containing white mica and magnetite, and giving a slightly fetid odor when broken. The marble layers may be as much as 0.4 m thick, but are mostly 7-15 cm in thickness and make up only 1-2% of the unit. The fact that the marble layers occur near the base of the schist

makes it unlikely that they could be correlative with the Chambless Limestone of the Marble Mountains.

Brown-weathering, green, laminated, calcareous quartzite makes up 2-3% of the unit. The quartzite is fine- to medium-grained and occurs in layers 6-20 cm in thickness. The layers are present throughout the unit. They probably are equivalent to quartzite and silty sandstone beds interlayered in the Latham Shale of the Marble Mountains (Stewart, 1970).

Three thin sections of the mica schist were examined. Average mineral percentages are: muscovite (35%), quartz (30%), biotite (20%), plagioclase (8%), potassium feldspar (3%), chlorite (3%), and magnetite (1%). Foliation in the rock is defined by preferred orientation of micas, by a variably-developed shape fabric in quartz and plagioclase, and by thin, elongate aggregates of quartz. Micas and felsic minerals are segregated into discrete domains to a much greater degree than in micaceous schists of other units in the study area.

One thin section of a typical calcareous quartzite was examined. This rock contains quartz (40%), microcline (30%), hornblende (10%), diopside (10%), and plagioclase (10%) with trace amounts of sphene and magnetite. Foliation is defined by alignment of hornblende and by a weak shape fabric in felsic minerals.

Lithologic layering is due to alternating laminae of quartz + feldspar + hornblende, and quartz + feldspar + diopside + minor hornblende.

Metamorphosed lower part of the Bonanza King Formation

Sharply overlying the metamorphosed Bright Angel Shale is a distinctive unit consisting of banded calcite marble which is equivalent to the lower part of the Cambrian Bonanza King Formation of the Ship, Marble, and Providence mountains (Palmer and Hazzard, 1956; Stone and others, 1983). The unit also is considered by Stone and others (1983) to be broadly correlative with the Cambrian Muav Limestone of the western Grand Canyon.

The metamorphosed lower part of the Bonanza King Formation in the study area is exposed extensively in the inverted Paleozoic zone and is present as tectonic slivers in the chaotic zone. In the inverted Paleozoic zone an apparently complete section of the unit, ranging up to 47 m in thickness is present. At the top of the unit, banded calcite marble is interlayered with the overlying dolomite marble over an interval up to 25 m thick. Although this may be a gradational stratigraphic contact, the widespread presence of tight to isoclinal folds within the unit suggests that the interlayering of rock types may be a product of intense folding. Regionally, Stone and other (1983) indicate that the

metamorphosed lower part of the Bonanza King Formation is overlain sharply by dolomite marble. For mapping purposes the upper stratigraphic contact of the unit in the inverted Paleozoic zone was placed at the first appearance of interlayered dolomite marble.

The metamorphosed lower part of the Bonanza King Formation consists primarily of medium-grained, laminated calcite marble (fig. 9). On fresh surfaces laminae are white and blue-gray. White laminae weather to a very pale gray or yellow-gray, whereas blue-gray laminae weather to darker shades of gray. Laminae, commonly wavy and discontinuous, range in thickness from less than 1 mm to over 1 cm.

Interlayered with the laminated marble are massive, tan-weathering, medium- to coarse-grained calcite marble layers. These massive intervals range up to 1 m in thickness and, like the laminae in the banded rocks, are discontinuous. They are recessive-weathering and friable in contrast to the resistant laminated layers. The abundance of massive layers is variable, but they seldom make up more than 20% of the unit.

One thin section of the laminated marble was analyzed. The rock is composed of calcite (94%), dolomite (2%), colorless Mg-chlorite (2%), epidote (1%), and magnetite (0.5%). Foliation defined by a variable shape fabric in calcite and dolomite and by preferred

Figure 9. Calcite marble of the metamorphosed lower part of the Bonanza King Formation showing characteristic weathered surface colors and discontinuous banding.

orientatation of Mg-chlorite crystals. Carbonate minerals show no evidence of bent twin lamellae, but Mg-chlorite crystals commonly are bent and unrecrystallized. Color banding in the rock is caused by concentration of Mg-chlorite and magnetite in the darker laminae.

Dolomite Marble

Tan-weathering dolomite marble gradationally overlies the metamorphosed lower part of the Bonanza King Formation in the inverted Paleozoic zone, and is overlain sharply by the metamorphosed Redwall Limestone in the upright Paleozoic zone. A complete section of the dolomite marble unit is nowhere preserved, but structural thicknesses range up to 46 m.

Precise correlation of the dolomite marble unit is uncertain. Stone and others (1983) reported that the equivalent stratigraphic position in the western Grand Canyon is occupied by an unnamed dolomite unit of Late Cambrian age and the overlying Temple Butte Limestone of Devonian age (McKee, 1969). More locally, the equivalent interval is occupied in the Ship, Marble, and Providence mountains by the Middle Cambrian upper part of the Bonanza King Formation, the Late Cambrian Nopah Formation, and the Devonian Sultan Limestone (Palmer and Hazzard, 1956). Stone and others (1983) indicated that the dolomite marble of their metamorphosed sections is

at least partially equivalent to these units. The presence in the Kilbeck Hills of what appears to be a gradational stratigraphic contact with the underlying banded calcite marble suggests that the Middle Cambrian upper part of the Bonanza King Formation is the most likely equivalent of at least the lower part of the dolomite marble.

Most of the unit consists of white to blue, massive to laminated, medium- to coarse-grained dolomite marble. White marble weathers to a light tan color, whereas the blue marble weathers to a pale blue-brown. In general, the blue marble is layered or laminated on scales as small as 5 mm. White marble tends to be more massive with layering defined by joint-like parting surfaces at intervals ranging from 1 cm up to half a meter or more (fig. 10).

In hand specimen both the white and the blue rocks consist predominantly of dolomite with variable but minor amounts of calcite. Phlogopite commonly is visible on foliation surfaces; tremolite and magnetite are accessory minerals. Metamorphic foliation is defined by a variable shape fabric in dolomite and by alignment of tremolite and phlogopite.

In thin section, S1 foliation is defined by a variable but generally strong shape fabric in dolomite and preferred orientation of phlogopite. Dolomite crystals

Figure 10. Tan-weathering, white Cambro-Devonian dolomite marble overlain by banded marble of the metamorphosed lower part of the Bonanza King Formation. Note the uneven development of layering in the dolomite.

(96%) are up to 3 mm in size and are entirely anhedral with few 120° triple junctions. Anhedral calcite crystals (3%) are up to 0.7 mm in size and mostly occupy interstices between dolomite crystals. Phlogopite crystals (1%) are up to 0.5 mm in size and occur as elongate to stubby, subhedral to anhedral grains often showing square terminations. Much of the phlogopite is disseminated, but some occurs in loose throughgoing clusters, probably reflecting slightly more argillaceous layers in the protolith. Subhedral blades of tremolite (1%) are up to 6 mm in length and lie parallel to S1 foliation. Magnetite is a trace constituent, occurring as equant euhedral to subhedral crystals and as anhedral interstitial crystals elongate parallel to S1.

Locally, two varieties of dolomitic calcite marble occur near the top of the unit in the upright Paleozoic zone. The first is a light-gray, scaly-weathering, fine-grained marble with layering on a scale of 1-5 cm. The second is a buff-colored, medium- to fine-grained, irregularly laminated rock containing abundant phlogopite and variable amounts of tremolite. Where exposed, these rock types are intensely folded. As a result, their exact thickness and stratigraphic positions in the unit are uncertain.

Metamorphosed Redwall Limestone

A distinctive white, chert-banded calcite marble sharply overlies the dolomite marble unit. The white marble is interpreted by Stone and others (1983) as the metamorphosed equivalent of a regionally extensive Mississippian carbonate sequence including the Redwall Limestone of the western Grand Canyon and the Monte Cristo Limestone of the Providence Mountains.

The unit is present as tectonic slivers in the chaotic zone and, in at least one place, in the migmatitic gneiss zone. It is the only Paleozoic unit exposed throughout the strike length of the upright Paleozoic zone; both the upper and lower contacts are preserved in several places. These sections range from 6 m to 19 m in thickness. In the inverted Paleozoic zone, the Redwall-equivalent marble is absent. There, the inverted Cambro-Devonian dolomite marble unit lies directly on the metamorphosed Permo-Pennsylvanian Bird Spring Formation along a ductile, low-angle fault.

This unit consists of white, medium- to coarse-grained calcite marble containing thin bands of meta-chert (fig. 11). Layering in the marble is defined by micaceous partings at intervals of 1 to 10 cm. Weathered surfaces range in color from white and light gray to pale red and yellow. In general, red and yellow colors are confined to the upper and lowermost 1 to 2 m

Figure 11. Metamorphosed Redwall Limestone of the up-right Paleozoic zone showing the metachert interval near the middle of the unit. Key-ring in lower center provides scale.

of the unit. Thin (3-10 mm) quartz-rich layers, especially common near the top of the unit, weather to a light- to medium-brown color. Elongate, dark gray carbonaceous(?) streaks up to 1 cm in thickness are present locally.

Dark brown-weathering metachert bands occur singly and in groups. They are composed of fine-grained quartz and range in thickness from 1 to 3 cm. Groups of closely spaced chert bands range from 15 cm to 1 m thick. A prominent group occurs near the bottom, roughly one-fifth of the way above the base of the unit. Another, thinner group occurs near the middle of the unit. A few individual metachert bands occur in the upper half of the unit.

In thin section the marble is seen to consist almost entirely of calcite and quartz. Calcite crystals (90%) are up to 3 mm in size, with the coarsest occurring in quartz-poor layers. Calcite crystals are anhedral with a moderately well-developed shape fabric and no crystallographic fabric. Deformation twinning is common. Quartz crystals (10%) are anhedral, up to 0.3 mm in size, and have a weak to fair shape fabric. Disseminated quartz is round to oval in shape, whereas rare quartz aggregates show polygonal textures. Stubby, disseminated crystals of muscovite, up to 0.2 mm in size, are a trace constituent.

Metamorphosed Bird Spring Formation

A heterogeneous unit composed of calcite marble, sandy marble, calc-silicate rocks, and quartzite gradationally overlies the metamorphosed Redwall Limestone. The unit is interpreted by Stone and others (1983) as being equivalent to the Bird Spring Formation of Pennsylvanian and Early Permian age. Its internal stratigraphy in the Kilbeck Hills is obscured by structural complexities. In general, the lower part of the unit is dominated by calcite marble and the upper part by calc-silicate-bearing marble and quartzite. However, the wide thickness variations shown by the unit and the limited area of exposure make even these generalizations tenuous.

A "complete" section of the unit is present only in the southwest corner of section 16 and adjacent parts of sections 20 and 21, T3N, R16E (see plate 1). In section 16 the unit is 14 m thick and consists of 1.2 m of thinly layered calcite marble transitional with the underlying Mississippian marble, overlain by 12.8 m of interlayered calc-silicate marble, calcite marble, calc-silicate schist and calcareous quartzite. In section 10, T3N, R16E, an inverted and fault-bounded section of the unit is 157 m thick. It consists of 54 m of interlayered calc-silicate marble, calc-silicate rock, and quartzite overlain by 103 m of calcite marble with minor

calc-silicate rock, quartzite, and dolomite marble. It is uncertain whether these thickness differences are due primarily to variable ductile attenuation of the unit or to unrecognized isoclinal folding or ductile, low-angle faulting.

Regionally, the Bird Spring Formation is part of a northwest-thickening Pennsylvanian and Early Permian sequence which grades from dominantly quartzitic in the southeast to dominantly calcareous in the northwest. In the Palen, Arica, Big and Little Maria, and Riverside mountains to the southeast, the Mississippian marble is overlain by the Supai Group which is composed of poorly bedded calcareous quartzite, calc-silicate rocks, and subordinate calcite and dolomite marble. Stone and others (1983) reported that the Supai Group typically ranges from 50 to 500 m in tectonic thickness.

Seventy kilometers north of the Kilbeck Hills, the Pennsylvanian and Lower Permian Bird Spring Formation of the Providence Mountains rests unconformably on the Mississippian Monte Cristo Limestone. The Bird Spring Formation is nearly 900 m thick and consists of limestone and sandy limestone, in part cherty and dolomitic, with minor sandstone and shale (Hazzard, 1954). In the Ship Mountains, 16 km northwest of the Kilbeck Hills, nonmetamorphosed Pennsylvanian and Lower Permian carbonates and minor quartzite that are equivalent to the Bird

Spring Formation are at least 750 m thick with neither the base nor the top exposed. Marble, quartzite and calc-silicate rocks of the Old Woman-Piute range are interpreted as equivalent to the Pennsylvanian-Early Permian rocks of the Ship Mountains (Stone and others, 1983). In the Little Piute Mountains, Bird Spring-equivalent rocks are about 75 m in tectonic thickness and consist of a lower subunit of variably sandy calcite marble, a middle subunit of quartzite, calc-silicate rocks, and variably sandy and siliceous marble, and an upper subunit of sparingly sandy calcite marble and minor quartzite.

Calcite marble is the most abundant rock type in the metamorphosed Bird Spring Formation of the study area. It is white to pale tan in color and may weather white or, more commonly, pale pinkish-gray or light brown. It is coarse- to fine-grained with foliation defined by alignment of elongate calcite crystals in the coarser-grained rocks. Exposures of calcite marble may be up to 25 m in thickness, but more commonly they consist of layers 1 cm to 0.5 m thick interlayered with sandy marble, calc-silicate marble, or quartzite.

Sandy marble occurs interlayered with calcite marble and quartzite. The rock is medium- to fine-grained and has laminations and layers up to 15 cm thick defined by differences in quartz content. Quartz

crystals weather dark brown and stand out in relief against the softer calcite which weathers buff, pale yellow, or pale red. Quartz-rich rocks may show a nearly solid dark brown color, whereas quartz-poor rocks have only disseminated specks of dark brown quartz against a paler calcite groundmass.

In thin section, foliation in the sandy marble is defined primarily by elongate quartz aggregates, and by a variably-developed calcite shape fabric. No crystallographic fabric is visible in calcite or quartz. Calcite crystals (75-80%) are 0.5-1 mm in size and show irregular to polygonal grain boundaries. Undulatory extinction and deformation twinning are common. Quartz crystals (15-20%) are 0.2 mm or less in size and occur as variably elongate aggregates of irregularly-shaped crystals and as disseminated grains. Undulatory extinction is ubiquitous in the quartz. Accessory minerals include potassium feldspar (1%) and ragged, often bent phlogopite (0.5%), plus trace amounts of zircon, apatite, and epidote.

Calc-silicate marble is a more highly metamorphosed equivalent of the sandy marble. The most common type is white, pale-brown-weathering wollastonite marble, slabs of which make a distinctive ringing sound when struck with a hammer. Also abundant locally is pale green to nearly white, diopside-bearing marble which may contain

thin layers rich in pale brown idocrase. Both rock types occur as layers up to 1 m thick, but typically they are interlayered on a scale of less than 20 cm both with each other and with quartzite and calcite marble.

In thin section the wollastonite marble is seen to contain widely differing proportions of calcite and wollastonite with minor feldspars, diopside, and quartz. Calcite (30-65%) is up to 1 mm in size, showing a variable shape fabric and no crystallographic fabric. Calcite-calcite grain boundaries range from irregular to polygonal. Subhedral wollastonite (30-60%) is up to 0.5 mm in size and shows a variable, but generally strong alignment with the calcite shape fabric. Accessory minerals include potassium feldspar (8%), plagioclase (4%), diopside (0.5-2%), and quartz (0-2%).

Three different varieties of calcareous quartzite occur in the unit. Most common is brown-weathering, green, diopside-bearing quartzite. A second type, abundant locally, is dark brown-weathering, white, wollastonite-bearing quartzite, containing quartz, calcite, and wollastonite in mutual contact (fig. 12). In thin section this rock consists of quartz (45%), wollastonite (35%), calcite (6%), plagioclase (5%), diopside (5%), and potassium feldspar (4%). The least common variety is dark-brown-weathering, orange and pale-brown, laminated calcareous quartzite consisting of quartz (35%),

Figure 12. Wollastonite-bearing calcareous quartzite from the metamorphosed Bird Spring Formation. Note that calcite (C), quartz (Q), and wollastonite (W) are in mutual contact in the center of the photograph (crossed nicols, x 100).

calcite (30%), idocrase (15%), garnet (10%), and plagioclase (8%), with minor phlogopite, potassium feldspar, and diopside.

Vitreous quartzite is a minor, but locally important, component of the unit. It is dark brown-weathering and medium-grained, and appears in hand specimen to be virtually 100% quartz. In most of the study area, vitreous quartzite layers are less than 15 cm thick; however, in the large Bird Spring exposure of section 10, T3N, R16E, several vitreous quartzite layers approach 1 m in thickness.

Thinly-laminated calc-silicate schists are a locally abundant component of the metamorphosed Bird Spring Formation. Their stratigraphic position in the unit is uncertain. The schists are brown-weathering, green, and light brown rocks, which tend to weather recessively. They are quite similar in appearance and mineralogy to calc-schists of the metamorphosed Hermit Shale. In general, the schists of the metamorphosed Bird Spring Formation tend to be less thinly and regularly laminated than those of the Hermit Shale, but in some cases the two cannot readily be distinguished.

Thin section study of calc-silicate schist reveals that lamination in the rock is due to alternation of wollastonite-rich and microcline-diopside-rich domains (fig. 13). Wollastonite (48%) occurs as small, elongate

Figure 13. Wollastonite-bearing calc-silicate schist from the metamorphosed Bird Spring Formation, showing alternating laminae rich in polygonal potassium feldspar and fine-grained wollastonite. Anhedral diopside is present in all layers (crossed nicols, x 40).

crystals with a fair preferred orientation parallel to layering. It occurs as inclusions in microcline and garnet. Microcline (28%) occurs primarily as polygonal crystals in elongate aggregates, along with diopside and minor plagioclase and calcite. Diopside (15%) occurs as small anhedral crystals and as larger, locally skeletal crystals containing patches of microcline. Plagioclase (6%) shows undulatory extinction and deformation twinning. It occurs mostly in microcline aggregates. Calcite (4%) forms patches of small anhedral crystals in wollastonite-rich layers and occurs as polygonal grains in feldspar aggregates. Pale grey-green garnet (4%) forms large skeletal patches commonly elongate in the foliation. It contains inclusions of all other minerals in the rock.

Metamorphosed Hermit Shale

Sharply overlying the metamorphosed Bird Spring Formation are green calc-silicate schists interpreted by Stone and others (1983) as metamorphosed equivalents of the Permian Hermit Shale. Though missing in the Providence and Plomosa mountains, Hermit-equivalent rocks are present in nearly all of the metamorphosed Upper Paleozoic sections studied by Stone and others (1983). In the Big Maria Mountains, Hamilton (1982) described the Hermit Shale as fine-grained, pale green, quartzitic,

locally wollastonite-bearing, calc-silicate schist, with tectonic thicknesses ranging from 0.7 to 50 m. In the Riverside Mountains Lyle (1982) interpreted greenish-white, porcelaneous, chloritic schists as metamorphosed Hermit Shale. The schists overlie cliff-forming quartzite and calc-silicate rock of the metamorphosed Supai Group and have a maximum thickness of only 5 m. Fine-grained, green-gray quartz phyllite in the Arica Mountains, correlated by Baltz (1982) with Hermit Shale, ranges up to 90 m in thickness. Hermit-equivalent rocks also are present in the Little Piute Mountains, where they consist of weakly metamorphosed purple and red calcareous mudstone, siltstone, and fine-grained sandstone (Stone and others, 1983). In the Kilbeck Hills, complete sections of the metamorphosed Hermit Shale range up to 19 m in thickness, but most are less than 10 m thick.

The metamorphosed Hermit Shale consists of green, fine- to medium-grained, laminated calc-silicate schist. It tends to be a slope-forming unit, and good, continuous exposures are rare. The rock weathers to a variety of brown colors ranging from pale red-brown to a deep chocolate-brown. The schist is laminated to thinly-layered, with light-brown to pale-green layers alternating with dark green layers. The scale of layering ranges from 2 mm to 1 cm.

In thin section the dark green layers are seen to be composed of diopside + potassium feldspar + plagioclase, whereas light brown layers contain wollastonite with minor potassium feldspar, diopside and calcite. Wollastonite (40%) is up to 0.6 mm in length and shows a fair alignment with compositional layering. Diopside (20%) is up to 3 mm in size and occurs as anhedral, in part skeletal, crystals which include feldspars and wollastonite. Elongate diopside crystals have their long axes parallel to compositional layering. Microcline (20%) is up to 1 mm in size, but most occurs as fine polygonal crystals in elongate feldspar aggregates. Larger, irregularly shaped crystals are intergrown with diopside. Many microcline crystals show weakly developed grid twinning. Plagioclase is present as anhedral, altered grains up to 0.3 mm in diameter, a few of which have well developed polysynthetic twinning. Neither potassium feldspar nor plagioclase have a crystallographic fabric. Calcite (5%) along with iron oxides is present as fine interstitial material in wollastonite-rich layers.

Metamorphosed Coconino Sandstone

A thin sequence of micaceous quartzite sharply overlies the metamorphosed Hermit Shale. It has been correlated with the Coconino Sandstone of Permian age

(Stone and others, 1983). Although the rock types in the unit are rather uniform throughout the region, the tectonic thickness is not. In the Big Maria Mountains, Hamilton (1982) reported thicknesses as ranging from 80 m to as little as 1 m. Emerson (1982) described a single 10-m-thick exposure of metamorphosed Coconino Sandstone in the Little Maria Mountains. In the Arica Mountains, Baltz (1982) noted thicknesses ranging from 27 m to over 170 m in rocks correlative with the Coconino Sandstone. In the Kilbeck Hills the metamorphosed Coconino Sandstone has a maximum tectonic thickness of 16 m, but in most exposures it is less than 5 m thick.

The unit consists entirely of strongly foliated and lineated, fine-grained micaceous quartzite. Fresh surfaces are light gray to white, whereas weathered surfaces are light gray, gray-brown, or dark-brown. Weathered foliation surfaces generally contain enough muscovite to give the rock a silvery sheen. In many outcrops the rock has two orthogonal sets of closely spaced joints perpendicular to S₁ foliation. These cause this unit to weather into small slabs and chips.

In thin section foliation is defined by a strong preferred orientation of muscovite and a variable tendency toward elongation of quartz crystals. Quartz crystals make up 75 to 90% of the rock, are typically 0.1-0.3 mm in diameter and show no crystallographic

fabric. Polygonal textures are well developed locally (fig. 14). Plagioclase crystals (5-10%) are anhedral, up to 0.4 mm in length, and highly altered to sericite and calcite. Muscovite crystals (5-8%) are up to 0.4 mm in length and occur as slender disseminated grains showing ragged terminations. Calcite (0-10%) is present as irregularly shaped interstitial crystals up to 0.4 mm in longest dimension. Potassium feldspar, magnetite, and zircon are present in trace amounts.

Metamorphosed Kaibab Limestone

The metamorphosed Coconino Sandstone is overlain by up to 9 m of varicolored calcite and dolomite marble. The unit is interpreted by Stone and others (1983) as broadly correlative with the combined Kaibab Limestone and Toroweap Formation of the western Grand Canyon. Stone and others (1983) described the "typical" metamorphosed Kaibab Limestone of their study area as ranging from 30 to 350 m in tectonic thickness, with a lower one-half of buff dolomitic marble and an upper one-half of thick-bedded to massive gray calcite marble and white, in part chert-banded, calcite marble. A thin zone of red schistose marble typically is present at the base of the unit. Local variations noted by Stone and others (1983) include the absence of dolomitic marble in the Little Piute Mountains and the presence of anhydrite

Figure 14. Micaceous quartzite from the metamorphosed Coconino Sandstone, with S1 foliation (from lower right to upper left) defined by oriented muscovite crystals. Note the well-developed polygonal textures shown by quartz crystals (crossed nicols, x 40).

schist in the Big Maria, Little Maria, Riverside, and Palen mountains.

The metamorphosed Kaibab Limestone of the Kilbeck Hills consists of blue and white, fine- to medium-grained, variably dolomitic calcite marble. Weathered surfaces may be buff, pale gray, light brown, or blue (fig. 15). Layering may occur on a variety of scales. At the largest scale, 1- to 3 m-thick intervals are dominated by one color. At the smallest scale are intervals from 1 to 20 cm thick which are laminated on scales of 1 to 3 mm. Color bands in the unit may resemble superficially those in the metamorphosed lower part of the Bonanza King Formation. In general, however, layering in the Kaibab tends to be thicker, more variable in thickness and color, more continuous, and less wavy in appearance.

The most abundant rock type in the unit is tremolite-bearing dolomitic calcite marble. In thin section the rock is seen to consist of up to 96% calcite and 2-8% dolomite and has a foliation defined by alignment of tremolite and chlorite and aligned domains of larger versus smaller calcite crystals. Calcite and dolomite range up to 2.5 mm in size and show a rough elongation in the foliation. Most calcite-calcite grain boundaries are highly irregular, although some 120° triple

Figure 15. Dolomitic calcite marble of the metamorphosed Kaibab Limestone. This view illustrates the lithologic variability as well as the varied scale of color banding in the unit. More dolomitic rocks such as those above the hammer weather to a tan color and appear similar to rocks of the Cambro-Devonian dolomite marble unit.

junctions occur in finer-grained domains. Bent twin lamellae are common, especially in larger grains. Tremolite (1-8%) is up to 2 mm in size, and occurs either as single, subhedral grains or as sheaf-like aggregates of skeletal crystals (fig. 16). Ragged, commonly bent crystals of pleochroic Fe-chlorite (trace to 2%) and colorless Mg-chlorite (trace to 6%) occur as individual crystals and as loose aggregates oriented parallel to foliation. Fe-chlorite also appears in thin veins parallel to foliation. Trace amounts of phlogopite also are present.

Mesozoic Rocks

Mesozoic rocks in the study area include (1) two metasedimentary units assigned a Triassic(?) protolith age, (2) a unit of mixed igneous and sedimentary protoliths assigned a Triassic(?) and Jurassic(?) age, (3) one intrusive unit assigned a Jurassic protolith age, (4) three intrusive units assigned a Jurassic(?) protolith age, (5) three intrusive units assigned a Late Cretaceous age, and (6) a unit of mixed igneous and sedimentary protoliths and Late Cretaceous intrusive rocks assigned a Mesozoic age.

Triassic(?) units include garnet-mica schist and siliceous marble and calcareous quartzite. Also assigned a Triassic(?) age are laminated calcareous

Figure 16. Tremolite-bearing dolomitic calcite marble of the metamorphosed Kaibab Limestone, with S1 foliation defined by the orientation of tremolite (Tr), colorless Mg-chlorite (Mg-ch), and pleochroic Fe-chlorite (Fe-ch) (crossed nicols, x 40).

quartzites of the Jurassic(?) and Triassic(?) gneiss, schist, and quartzite unit. The relative ages of Triassic(?) rocks in the study area are unknown. The laminated calcareous quartzite occurs primarily at or near the base of the gneiss and schist zone, commonly in contact with metamorphosed Kaibab Limestone of the underlying chaotic zone; however, the nature of the contact between the two rocks is uncertain and the quartzite was not assigned to any specific regional Triassic or Jurassic metasedimentary unit.

Jurassic and Jurassic(?) rocks of the study area consist of a complex suite of plutonic rocks of dioritic to granitic composition. Based on published descriptions of Jurassic arc rocks in the region (John, 1981; Howard and John, 1984), most of the Jurassic and Jurassic(?) rocks in the Kilbeck Hills appear to be closer compositionally to the quartz-rich rocks of the Sheep Hole and Bullion mountains than to the quartz-poor rocks of the Ship and Marble mountains.

An undeformed biotite monzogranite pluton in the southern part of the study area has been correlated by K. Howard (personal communication, 1984) with the granodiorite of the Late Cretaceous Old Woman pluton. Cutting the monzogranite is a two-mica granite pluton correlated by Howard with Late Cretaceous two-mica granite plutons of the Old Woman-Piute range.

Garnet-Mica Schist

Garnet-bearing mica schist is a major component of the gneiss and schist zone over most of the zone's strike length. The schist is present as a continuous, partially fault-bounded, strip in sections 16 and 17, but it is absent in the gneiss and schist zone of section 10, T3N, R16E. The thickness of the garnet-mica schist in the gneiss and schist zone ranges from 3 m to 54 m.

The stratigraphic position of the schist is uncertain, but available data point to a Mesozoic, probably Triassic age. This age assignment is based on the rock types of the unit and its occurrence only in the gneiss and schist zone, which is bounded by inward facing Paleozoic sequences. Mica schists of Triassic age are uncommon in the Mojave region. The nearest occurrence to the Kilbeck Hills is in the Little Maria Mountains. There, biotite-muscovite-(garnet)-epidote schists were included by Shklanka (1963) in the upper part of the Permian(?) Maria Formation of Miller (1944). Stone and others (1983) have reinterpreted the upper part of this unit as Triassic(?) in age.

A Proterozoic or Paleozoic age for the garnet-mica schist unit seems improbable. No metasedimentary rocks are known to be associated with the Proterozoic gneisses of the study area, although staurolite-bearing mica

schists of probable Proterozoic age are present in the Little Piute Mountains to the northeast. Correlation with the metamorphosed Bright Angel Shale appears unlikely because the Bright Angel schists lack both garnet and sillimanite, and in most places they have a higher muscovite/biotite ratio than the garnet mica schist.

The unit consists of fine- to medium-grained mica schist, which in most places contains porphyroblasts of garnet and (or) magnetite up to 4 mm in diameter. Weathered surfaces range from silver-gray to gray-brown to a nearly black "desert varnish" color. The S1 foliation is the only planar structural element present. The L1 lineation, commonly paralleling hinge lines of F1 microfolds, is visible in nearly all outcrops. F2 crenulations are often present as well, their hinge lines forming the L2 lineation on S1 surfaces.

The schist contains the mineral assemblage quartz + biotite + muscovite + plagioclase + magnetite + garnet ± microcline ± sillimanite. Biotite/muscovite ratios range from 3/2 to 2/3. Microcline is present in most samples, but usually it is less abundant than plagioclase.

Ten thin sections of the schist were examined. Average mineral percentages are: quartz (40%), biotite (28%), muscovite (25%), plagioclase (6%), Kspar (4%), garnet (1%), magnetite (1%). Pale pink, isotropic

garnets are up to 3 mm in diameter and are rounded to angular and equant to elongate parallel to S1 foliation. Many garnet crystals are highly fractured, with chlorite alteration along fractures. Linear inclusion trails of quartz occur in a few garnet crystals. These commonly parallel the external S1 foliation indicating that garnet growth at least partly postdates S1 development. Rare sillimanite-bearing samples are relatively rich in muscovite and contain few garnets and no microcline (fig. 17). Fibrolitic sillimanite crystals are up to 0.5 mm long and occur in lenticular domains which are rich in biotite and contain very little muscovite. Biotite and quartz crystals within the sillimanite-bearing domains are slightly finer grained. The long axes of the domains and of sillimanite crystals are oriented parallel to the L1 lineation.

Also present in the unit are layers of light-brown-weathering, laminated calcareous quartzite up to 15 cm in thickness. Quartzite layers occupy no consistent stratigraphic position and make up less than 5% of the unit. The rock contains the assemblage quartz + calcite + microcline + diopside + plagioclase + garnet + biotite.

Siliceous Marble and Calcareous Quartzite

A unit composed of interlayered siliceous marble

Figure 17. Sillimanite-bearing schist from the Triassic(?) garnet-mica schist unit. This view, perpendicular to S1 and L1, shows characteristic clustering of fibrolitic sillimanite crystals, which are seen in cross-section (plane polarized light, x 100).

and calcareous arkosic quartzite overlies the Triassic(?) garnet-mica schist in the gneiss and schist zone. This unit has a maximum thickness of 38 m and is present in sections 16 and 17, T3N, R16E. Structurally above the unit is a ductile, low-angle fault which progressively cuts out the siliceous marble and calcareous quartzite along strike toward the northeast, bringing the underlying garnet-mica schist directly against the overlying upper slice of the Jurassic(?) and Triassic(?) gneiss, schist, and quartzite unit. Along its lower contact, the marble and quartzite are interlayered with the garnet-mica schist over an interval of 2-4 m (fig. 18). The interlayering of rock types implies that the contact is depositional, but the relative age of the units is unknown.

Lithologically, the unit resembles parts of the metamorphosed Bird Spring Formation, especially the structurally lowest part of the thick Bird Spring exposure in the inverted Paleozoic zone of section 10, T3N, R16E. Nevertheless, the unit is assigned a Triassic(?) age because it lies in apparent stratigraphic contact with the underlying garnet-mica schist which is interpreted here as Triassic(?) in age.

The unit consists of white to pale-green, medium- to coarse-grained siliceous marble and brown-weathering, green, fine-grained calcareous, arkosic quartzite. The

Figure 18. Contact between Triassic(?) garnet mica-schist and Triassic(?) siliceous marble and calcareous quartzite. Fine-scale interlayering of rock types at the contact suggests that it is depositional. The relative age of the two units is uncertain.

two rock types are interlayered over a wide range of scales from 2 mm to 10 m. Small-scale layering in both rock types often is discontinuous and quartzite layers commonly are boudinaged (fig. 19). The most characteristic rock of the unit is a "knotted" marble, with thin, discontinuous layers and boudins of pale brown-weathering diopside-feldspar rock in a white or pale green, siliceous marble matrix. The marble tends to be punky-weathering and form slopes, whereas quartzite-rich intervals form steeper, cliff-like outcrops.

The marble is dominated by layers of contrasting mineralogy. The most obvious feature of the layering is the segregation of calcite and quartz from microcline. Although this segregation is not complete, calcite and quartz tend to occur with plagioclase, diopside, and scapolite, whereas microcline occurs in layers with diopside and only minor amounts of other minerals. The brown-weathering "knots" in the marble are composed of coarse diopside, with calcite and minor feldspars. All minerals show a good to excellent shape fabric but no crystallographic fabric. Examination of three thin sections of the rock yielded the following mineral percentages: calcite (25-30%), quartz (20-25%), diopside (15-20%), microcline (15-20%), plagioclase (5-8%), and scapolite (3-6%). Trace minerals are tremolite and sphene.

Figure 19. Triassic(?) siliceous marble and calcareous quartzite. The thin, discontinuous, dark brown layers are diopside-rich intervals in the marble. Calcareous quartzite layers are a lighter brown color and more continuous. Note the boudinage of the thicker calcareous quartzite layers.

Quartzite in the unit is laminated on a millimeter to centimeter scale. Thin section study shows domains of quartz and calcite alternating with domains rich in microcline. The lamination is accentuated by the presence of abundant biotite in some of the microcline-rich laminae. Foliation is defined by preferred orientation of biotite and by shape fabrics in quartz and diopside and, to a lesser extent, feldspars. Mineral percentages in the quartzite (based on one thin section) are: microcline (35%), quartz (32%), diopside (15%), calcite (8%), plagioclase (7%), biotite (2%), and sphene (0.5%). Trace amounts of magnetite also are present.

Gneiss, Schist, and Quartzite

A heterogeneous unit of gneiss, schist, and quartzite with a maximum structural thickness of 48 m occupies a large part of the gneiss and schist zone. All of its contacts with other units and most of the internal contacts between different rock types are interpreted as ductile, low-angle faults. Two separate fault-bounded slices of the unit are exposed in sections 16, 17, and 20, T3N, R16E. The upper slice consists entirely of metagneous rocks, whereas the lower slice contains both metagneous and metasedimentary rocks. Calcareous quartzites form a discontinuous band up to 6 m thick at

or near the structural base of the lower slice (fig. 20). This band of metasedimentary rocks also occurs at the base of the gneiss and schist zone of section 10, T3N, R16E, suggesting that the exposures there are a northeastward continuation of the lower slice of the gneiss, schist, and quartzite unit of section 16, T3N, R16E.

The metaigneous component of the unit is interpreted as Jurassic(?) in age based on its similarity to Jurassic(?) orthogneisses that cut Paleozoic rocks. The metasedimentary component of the unit is assigned a Triassic(?) age because no comparable rocks are known in the Paleozoic section and no metasedimentary rocks of Precambrian or post-Triassic age are known to be present in the study area.

Rocks within the unit that are interpreted as metaigneous are fine- to medium-grained orthogneisses and orthoschists containing two feldspars + quartz + biotite ± hornblende. They range from monzogranite to granodiorite in composition. In terms of fabric, the rocks range from foliated, weakly-lineated, compositionally homogeneous gneiss to strongly-lineated orthoschists containing elongate feldspar augen and quartz aggregates.

In thin section the foliation in the homogeneous gneiss is defined by aligned, disseminated biotite, by

Figure 20. Laminated calcareous quartzite of the Jurassic(?) and Triassic(?) gneiss, schist, and quartzite unit. The rocks lie at the base of the gneiss and schist zone in section 10, T3N, R16E, and overlie upright chaotic zone strata of the metamorphosed Kaibab Limestone. The nature of the contact is uncertain.

a shape fabric in quartz and feldspars, and by rare trains of sutured quartz crystals. Although felsic minerals are inequant, they are not strongly elongated parallel to the L1 lineation. Monomineralic aggregates of quartz and feldspars, especially plagioclase, show well-developed polygonal textures.

Orthoschists show a more pronounced segregation of felsic from mafic minerals and a stronger shape fabric in felsic minerals. Elongate trains of quartz and microcline up to 3 cm long and 2 mm thick alternate with domains containing biotite + feldspars + quartz ± hornblende. The foliation appears weaker in thin sections cut perpendicular to the L1 lineation. Crystals and aggregates of felsic minerals are less elongate and mica-rich domains are less continuous in this direction.

Most of the metagneous rocks of the unit have undergone some degree of metasomatic alteration. The most widespread indicator of metasomatism is epidote, present in discrete, cross-cutting and concordant veins, and as discontinuous, foliation-parallel trains of crystals. Calcite and garnet also are present in minor amounts locally. A sample of epidote-bearing biotite-hornblende-monzogranite gneiss was examined in thin section. The rock consists of plagioclase (28%), microcline (28%), quartz (20%), hornblende (10%), epidote (8%), biotite (5%), and magnetite (1%). Epidote-rich

layers contain little magnetite and virtually no biotite.

In other rocks, a greater degree of metasomatism is indicated by calc-silicate-bearing layers up to 2 cm thick which contain scapolite + epidote + diopside + quartz. These layers are common near contacts with carbonates of the underlying chaotic zone. In most cases the calc-silicate layers make up a small percentage of what is clearly a metaigneous rock. Locally, however, the calc-silicate layers make up 50% of the rock, and the protolith is less certain. In such places the rocks are banded and consist of black layers of potassium feldspar + hornblende alternating with green layers of scapolite + diopside + epidote. Layers of quartz monzodiorite composition containing two feldspars, quartz, biotite, and hornblende also are present. Thin section study of the rock gives the following overall mineral composition: scapolite (35%), potassium feldspar (25%), diopside (15%), hornblende (10%), epidote (10%), plagioclase (2%), and sphene (1%). Trace amounts of biotite, zircon, apatite, quartz, and calcite also are present.

Less abundant green and pink, irregularly-layered rocks containing two feldspars, quartz, and minor hornblende, along with abundant garnet and diopside and lesser amounts of epidote and calcite are seen only

along contacts with marble of the chaotic zone.

Metasedimentary rocks in the unit include laminated calcareous quartzites and quartz-plagioclase-mica schists. Calcareous quartzites are green and white, brown-weathering, fine- to medium-grained rocks, that are laminated on a scale of 1-2 mm. Thin section study reveals that lamination is due to elongate quartz aggregates or layers of quartz + plagioclase alternating with layers dominated by calcite + garnet + epidote or by plagioclase + minor epidote and hornblende. Diopside is present throughout the rock except in monomineralic quartz aggregates. Overall mineral percentages in the rock are: quartz (30%), calcite (20%), plagioclase (15%), diopside (10%), epidote (10%), garnet (10%), hornblende (4%), and microcline (2%) with trace amounts of sphene.

Gray, medium-grained, quartz-plagioclase-mica schists are interlayered locally with orthoschist and orthogneiss. Similar schists which were mapped as fault slivers of the gneiss, schist and quartzite unit also occur in the chaotic zone. The schists contain discontinuous, millimeter-scale laminations defined by alternating quartz-plagioclase and mica-rich domains. The rock tends to cleave along mica-rich domains; cleavage surfaces have a silvery sheen. Surfaces at a high angle to foliation weather light- to dark-brown.

In thin section, foliation in the schist is defined by variably-developed shape fabrics in quartz and plagioclase and by preferred orientation of micas. Quartz (55%) is anhedral, elongate, and up to 1.5 mm in length. Quartz-quartz boundaries are irregular but not sutured. Well-defined, elongate aggregates of pure quartz are not present. Plagioclase (25%) is anhedral, less elongate than quartz, and evenly distributed over most of the thin section. Polysynthetic twinning is fairly common and most does not appear to be bent or wedge-shaped. Muscovite (10%) and biotite (9%) are up to 2 mm in length and are subhedral with ragged terminations. Despite the appearance of mica-rich domains in hand specimen, thin section study shows that the micas are quite evenly distributed throughout the rock.

Diorite

A hornblende diorite pluton of Jurassic(?) age is exposed over an area of approximately 1.3 km² in the northern part of the study area. The diorite forms high, rounded hills and weathers to a dark-brown to dark-blue-gray color. The diorite is in contact only with alluvium and with Jurassic granitoids. External contacts with the dark granodiorite gneiss and the undivided Jurassic(?) gneiss (see plate 1) are poorly exposed, but they appear to be straight and abrupt, giving

no clue to the relative age of the units. The diorite, however, is cut by dikes of Jurassic dark granodiorite gneiss and Jurassic(?) granite gneiss. These relationships indicate that the diorite is the oldest Mesozoic plutonic rock so far recognized in the study area.

The dominant rock type is gray and black, medium-grained, hornblende diorite. In hand specimen the rock appears to consist entirely of plagioclase and hornblende. Megascopic fabrics generally are poorly developed. The only fabric visible in most places is a weak alignment of hornblende crystals and hornblende aggregates. They range from clusters of two or three crystals to multi-crystal clots up to 6 cm in diameter. Locally, however, the diorite shows a strong alignment of hornblende. As the pluton was not mapped or sampled in detail, the origin and distribution of fabrics in the diorite are uncertain. One sample from an outcrop 5 m above the contact with dark granodiorite gneiss was examined in thin section. Plagioclase (60%) is subhedral to anhedral and commonly is lath-shaped. Deformation twinning, undulatory extinction, and throughgoing fractures are common. The "a"-normal method gives a composition of An 45. Hornblende (38%) is subhedral to anhedral. Many crystals are altered and recrystallized. Larger grains commonly are a mass of randomly oriented blades and hexagonal cross sections of recrystallized

hornblende. Trace constituents include: potassium feldspar, biotite, sphene, apatite, and magnetite.

The diorite is cut by dikes of several different types of gneissic granitoids. In all cases observed, fabrics in the dikes are parallel to the weaker fabrics in the diorite and subparallel to dike margins. Most abundant are dikes of light gray, medium- to fine-grained Jurassic(?) granite gneiss. Also present are dikes of dark Jurassic granodiorite gneiss. A third type, not observed elsewhere in the study area, is a white and black gneiss containing white to pale-gray feldspar, minor quartz, and 15 to 20 percent of mostly disseminated, strongly aligned hornblende.

An area of dark-green and black, altered diorite occurs in the southwestern portion of the pluton. Plagioclase crystals here have undergone nearly complete alteration to fine, anhedral epidote with lesser amounts of albite and sericite. Hornblende crystals contain numerous red, oxidized patches, and hornblende clusters contain interstitial aggregates of magnetite and sphene.

Dark Granodiorite Gneiss

Dark, biotite-hornblende granodiorite gneiss is exposed at all structural levels in the northern Kilbeck Hills. It cuts Proterozoic and Paleozoic rocks as well as the Jurassic(?) diorite pluton at the north end of

the study area. The gneiss has yielded a U-Pb zircon age of 160 m.y. (J. E. Wright, per. comm., 1985). This date is interpreted as a crystallization age. The granodiorite gneiss, the most abundant rock type in most of the areas mapped as undivided Jurassic(?) gneiss (plate 1), is exposed extensively in the inverted Paleozoic zone. It occurs locally at lower levels of the nappe complex, but none of the outcrops is large enough to be mapped at a scale of 1:5,000. The gneiss forms elongate, sill-like bodies of widely varying thickness that are oriented sub-parallel to S1 foliation in Proterozoic crystalline rocks and to S0 layering and S1 foliation in Paleozoic strata. It is uncertain to what extent the gneiss bodies have retained their original, pre-D1 intrusive shapes and orientations.

The unit consists of dark gray to nearly black, medium- to fine-grained gneiss ranging from granodiorite and quartz monzodiorite, to monzogranite and quartz monzonite in composition. In hand specimen, the distinguishing characteristics of the unit are dark color, lack of potassium feldspar megacrysts, and the predominance of hornblende over biotite. Irregularly-shaped plagioclase megacrysts are present locally. These white, blocky crystals are up to 8 mm in diameter and are seldom abundant. The unit was not studied in enough detail to determine whether or not the megacrysts are

associated with a particular rock type.

Gneissic (S1) fabrics are variable in the unit, but generally they are strongly developed. The weakest fabrics occur in quartz-poor rocks, especially those intruding marble. Stronger fabrics are present in the quartz- and biotite-rich gneisses. The rocks commonly are schistose where hornblende and biotite are subequal. Linear fabrics also are well, but variably, developed. The L1 lineation is defined primarily by aligned hornblende crystals. Strong S1 and L1 fabrics do not always go together. Hornblende may be strongly aligned in poorly foliated hornblende-rich rocks but poorly aligned in biotite-rich schistose rocks.

In thin section the S1 foliation is seen to be defined primarily by alignment of hornblende and biotite, and by a variably-developed shape fabric in quartz and feldspar. Polygonal textures are common in aggregates of felsic minerals; crystallographic fabrics are absent.

Megacrystic Monzogranite Gneiss

Gray, in part megacrystic, monzogranite to granodiorite gneiss is exposed extensively in the inverted Paleozoic zone and comprises an uncertain but possibly large portion of the migmatitic gneiss zone. It cross-cuts the dark granodiorite gneiss and is cut by the

granite gneiss. The unit probably is correlative with Jurassic porphyritic granodiorite, monzogranite, and quartz monzonite mapped in ranges west of the Kilbeck Hills by Howard and John (1984). Rocks of the unit contain biotite $\pm$ hornblende in widely varying proportions and commonly contain potassium feldspar megacrysts. Like the dark granodiorite gneiss and the granite gneiss, the unit occurs as sill-like bodies oriented parallel to S1 foliation in older gneissic units and to S1 foliation and S0 layering in Paleozoic metasedimentary rocks.

The compositional limits of the unit are poorly defined. Many megacrystic rocks contain hornblende, but generally more biotite. Rocks lacking megacrysts and hornblende may be impossible to distinguish from the granite gneiss unit although the latter tends to be finer-grained and to have a slightly lower color index (5-15, versus 10-25 for the megacrystic gneiss). Non-megacrystic rocks with subequal amounts of hornblende and biotite are generally distinct from the dark granodiorite gneiss by virtue of lower maximum color index (25 vs 40). The field criterion adopted is that non-megacrystic rocks are assigned to this unit if they contain hornblende subordinate to biotite.

This unit consists of gray, light to dark brown-weathering, medium- to fine-grained hornblende-biotite

gneiss of monzogranite to granodiorite composition. About half of the exposed area of the unit contains white, pale gray, or pale pink, rounded to blocky perthitic microcline augen. The rock weathers to irregular, thin slabs that are elongate parallel to foliation. Aside from discontinuous millimeter-scale aggregates rich in felsic or mafic minerals, the gneiss is homogeneous with no compositional banding visible in hand specimen. The S1 foliation is well developed throughout the unit, but the L1 lineation ranges from strong to weak. In places the only linear fabric element is the preferred orientation of elongate microcline augen.

In thin section, foliation is defined by alignment of biotite, by a variable shape fabric in felsic minerals, and by elongation of feldspar and quartz aggregates. A typical sample of megacrystic gneiss consists of plagioclase (34%) of composition An 28-30 (determined by the "a"-normal method), perthitic microcline (34%), quartz (20%), and biotite (10%) with trace amounts of magnetite, sphene, zircon, allanite, and secondary muscovite.

Granite Gneiss

Light gray granite gneiss crosscuts all other foliated rocks and occurs in the inverted Paleozoic zone,

the gneiss and schist zone, and locally the migmatitic gneiss zone. Like the dark granodiorite gneiss and the megacrystic monzogranite gneiss, it is exposed most extensively in the inverted Paleozoic zone where it forms elongate bodies of varying size oriented sub-parallel to the S1 foliation. Small dikes of the granite gneiss may be oriented at up to 30° to S1, whereas similar bodies of megacrystic monzogranite gneiss rarely are oriented at more than 5° to S1. After the dark granodiorite gneiss, the granite gneiss is the most abundant component in areas mapped as undivided Jurassic(?) gneiss. The minimum age of the unit is uncertain, but its similarity to the Jurassic and Jurassic(?) leucocratic granite gneiss of the Calumet and Sheephole mountains, as described by Howard and John (1984), suggests a Jurassic(?) age.

This unit consists of pale gray, light- to medium-brown-weathering, biotite-bearing gneiss ranging from monzogranite to syenogranite in composition. The gneiss is fine- to medium-grained and is a true "granite gneiss" lacking any compositional layering except for discontinuous millimeter-scale quartz and feldspar aggregates. In outcrop, the distinguishing characteristics of the unit are the absence of feldspar megacrysts and hornblende, and a relatively leucocratic composition with a maximum color index of 10 to 15.

In thin section the foliation is defined by alignment of biotite and by elongate aggregates of felsic minerals. Shape fabrics are best developed in quartz and least developed in plagioclase. The mineral composition of a typical sample is: microcline (35%), plagioclase (33%) of composition An 27 (determined by the "a"-normal method), quartz (20%), biotite (10%), magnetite (1%), and sphene (0.5%) with trace amounts of apatite and zircon.

Migmatitic Gneiss

Migmatitic granitoid gneiss along with minor parashist and amphibolite occupies the lowest structural level in the nappe complex. The unit forms low, rounded hills which weather to a dark gray-brown color. The gneisses are overlain by rocks of the upright Paleozoic zone along a brittle, low-angle fault, and are intruded by a large body of monzogranite of Late Cretaceous age. The gneisses thin abruptly toward the southwest from a maximum exposed thickness of 250 m as the monzogranite pluton cuts upward to the overlying upright Paleozoic zone.

The monzogranite has migmatized the country rock gneisses through a zone many tens of meters thick. Near the contact, blocks of gneiss up to 5 m long are engulfed in the monzogranite. Some of the smaller gneiss

xenoliths have been partially digested and are preserved as layered, biotite-rich schlieren (fig. 21). Foliation in the xenoliths and the in-place country rock is everywhere parallel.

The ages of the various rock types within the unit are not well constrained. Some of the gneisses clearly are correlative with three of the Jurassic and Jurassic(?) granitoids present elsewhere in the study area. Amphibolites within the unit may be metamorphosed equivalents of Jurassic(?) diorite exposed in the northern part of the study area. Paraschists are similar to quartz-plagioclase-mica schists of the Jurassic(?) and Triassic(?) gneiss, schist, and quartzite unit. Therefore these rocks may be Triassic(?) in age. Much of the unit, especially the granitic to dioritic layered gneisses, has no obvious counterpart elsewhere in the study area and is of unknown age (fig. 22).

The most abundant rock type is gray, fine- to medium-grained, schistose, biotite monzogranite augen gneiss. The rock weathers dark brown and splits into slabs 1-2 cm thick parallel to S1 foliation. In hand specimen the gneiss shows an irregular lamination on a scale of 1-3 mm with discontinuous trains of quartz and feldspars alternating with biotite-rich laminae. Pink potassium feldspar augen typically are 2-3 mm thick and up to 1 cm in length parallel to L1. In thin section,

Figure 21. Late Cretaceous biotite monzogranite intruding thinly layered gneiss of the Mesozoic migmatitic gneiss unit. The smaller xenoliths to the right of center have been partially digested. Note the variable size and lack of rotation of the xenoliths.

Figure 22. Layered gneiss of the Mesozoic migmatitic gneiss unit. The vertical and lateral lithologic variability is characteristic of much of the unit.

the S1 foliation is defined by aligned biotite and by elongation of felsic minerals, especially quartz and microcline, parallel to L1. Trains of felsic minerals form the discontinuous lamination seen in hand specimen.

In thin section the rock contains 37% plagioclase of An 28 composition (determined by the "a"-normal method). Plagioclase crystals commonly are myrmekitic, have polygonal textures in aggregates and show a weakly developed shape fabric. Microcline (33%) occurs as perthitic augen and as small anhedral crystals. Augen contain inclusions of quartz, plagioclase, and biotite. Quartz (20%) occurs as disseminated grains and in elongate aggregates. Quartz-quartz grain boundaries in aggregates tend to be sutured rather than polygonal. Biotite (10%) occurs as disseminated grains and in elongate aggregates. Magnetite (1%) is anhedral to subhedral, commonly forming trains parallel to S1 foliation. Trace constituents are apatite, sphene, zircon, and allanite.

The second most abundant component of the unit is dark gray to white, fine- to medium-grained irregularly-banded gneiss. Layering occurs at two different scales; mafic-felsic laminations 1-4 mm thick, and bands of differing rock types 5 cm to 3 m thick. Both laminae and lithologic bands range from continuous (at outcrop scale) to discontinuous. The most common lithologic

types are: 1) biotite monzogranite with or without lenticular microcline augen; 2) biotite leuco-syenogranite to monzogranite, generally containing lenticular microcline augen; and 3) hornblende-biotite granodiorite containing small, rounded to lenticular augen of sodic andesine.

Another important rock type, abundant near the base of the unit, is fine- to medium-grained, gray and white to gray and pink, biotite monzogranite gneiss, which shows a wispy, irregular compositional banding. Mafic-rich layers contain biotite + plagioclase + quartz + minor microcline. Felsic-rich layers of variable thickness and continuity are rich in microcline and contain thin, discontinuous selvages of slightly coarser biotite. A few of the felsic layers also contain pink, rounded and fractured garnets up to 1 mm in diameter.

Lenticular pods of black amphibolite up to 5 m in length and 2 m in thickness (fig. 23) are present locally. This rock contains dark olive-green hornblende (50%) and plagioclase (40%) of An 43 composition (determined by the "a"-normal method). Plagioclase commonly shows undulatory extinction and deformation twinning. Quartz (6%) and magnetite (4%) are present as interstitial accessories. Magnetite also occurs with sphene as a common alteration product of hornblende. Foliation, defined by alignment of hornblende, by a

Figure 23. Amphibolite pod in the Mesozoic migmatitic gneiss unit. The black amphibolite pod (seen just below center) is wrapped around by S1 foliation in the enclosing gneiss and is not cut by the biotite monzogranite.

rough alignment of plagioclase and quartz clusters, and by a variable shape fabric in plagioclase and quartz, is moderately well developed.

Roughly 10% of the unit can be correlated directly with the Jurassic and Jurassic(?) granitoids of the study area. Dark hornblende-biotite granodiorite and biotite-hornblende quartz monzodiorite layers in the unit appear nearly identical to rocks of the Jurassic dark granodiorite gneiss unit. Rare megacrystic biotite monzogranite gneiss layers are very similar to rocks of the megacrystic monzogranite gneiss unit. Local layers of biotite granite gneiss which are identical to rocks of the granite gneiss unit crosscut other rock types at shallow angles. Many of the Jurassic and Jurassic(?) rocks of the unit appear more highly deformed than their equivalents at higher structural levels. Fine lithologic banding is more pronounced due to greater mafic-felsic segregation and to greater elongation of quartz and feldspar aggregates. Microcline and plagioclase augen are smaller, more numerous, and more highly fractured than those in the megacrystic monzogranite gneiss and the dark granodiorite gneiss of the inverted Paleozoic zone.

Medium-grained quartz-plagioclase-mica schists comprise 10-15% of the unit. The schists are speckled with silver muscovite crystals on foliation surfaces and

weather gray-brown. Except for the presence of muscovite, hand specimens appear very similar to the schistose metagneiss of the unit. The schists are interlayered in the gneisses forming strips traceable for hundreds of meters parallel to the strike of S1 foliation. The boundaries of the schist layers are not sharp; they grade with decreasing amounts of muscovite and increasing amounts of potassium feldspar into schistose biotite monzogranite to granodiorite gneiss. At places, the schists contain continuous interlayered, gradational 1- 3-cm-thick domains in which mineral percentages vary. One type of layer has quartz (45%), plagioclase (30%), muscovite (15%), and biotite (10%). A second type of layer is slightly finer grained and contains plagioclase (45%), quartz (25%), biotite (20%), muscovite (9%), and magnetite (1%).

Biotite Monzogranite

A body of biotite monzogranite to granodiorite occupies the southern portion of the study area, intruding the migmatitic gneiss zone and the upright Paleozoic zone. Rocks of the Late Cretaceous Cadiz Valley batholith lie south of the pluton in the central and southern Kilbeck Hills. Contact relationships of the unit with the Cadiz Valley batholith were not investigated in this study. The pluton has been correlated by K. Howard

(personal communication, 1984) with the Late Cretaceous Old Woman pluton, which occupies a similar structural position at the base of the nappe complex in the Old Woman Mountains and Little Piute Mountains.

The pluton intrudes rocks of the migmatitic gneiss zone both concordantly and discordantly. Along strike to the southwest, the pluton cuts upward into the upright Paleozoic zone with only thin strips of gneiss remaining at the base of the Paleozoic rocks. The contact with the carbonate rocks of the upright Paleozoic zone is sharp with none of the lit-par-lit concordant intrusion of monzogranite seen in the migmatitic gneiss zone. Dikes, sills, and small irregularly-shaped bodies of similar rocks discordantly intrude higher levels of the nappe complex. The pluton itself is cut by dikes of leucocratic monzogranite and two-mica ± garnet granite. No obvious contact aureole is apparent in either the migmatitic gneiss zone or the upright Paleozoic zone.

For at least 300-400 m below its upper contact, the pluton shows a well-developed planar alignment of biotite and locally a weaker alignment of feldspars. The biotite fabric generally is parallel to the intrusive contact and to S1 foliation in country rock xenoliths. Above the contact, the biotite fabric is parallel to the margins of dikes and sills derived from the pluton.

Although the pluton was not sampled or mapped in detail, field and limited petrographic study indicate that the pluton is relatively homogeneous. It consists of light-gray, medium- to fine-grained, biotite monzogranite to granodiorite. The rocks are equigranular in most places, but salmon-colored potassium feldspar phenocrysts up to 15 mm in diameter are present locally near the roof of the pluton. Potassium feldspar is everywhere subordinate to plagioclase and may be salmon-colored or white. Biotite is the only mafic mineral observed. Sphene which is an abundant, characteristic accessory mineral in many places, especially in biotite-rich rocks, is a variable component of the unit.

In thin section, plagioclase (45%) is subhedral to anhedral and commonly myrmekitic. The Michel-Levy method gives a composition of An 25, whereas the "a"-normal method gives a composition of An 27. Plagioclase crystals show some bending and offset of polysynthetic twin lamellae along fractures, but most twinning is not deformational. A few crystals show a vague normal zoning pattern. Potassium feldspar (25%) is subhedral to anhedral and perthitic with a few crystals showing a diffuse grid extinction pattern. Quartz (20%) is anhedral and interstitial with irregular to sutured quartz-quartz boundaries. Biotite (9%) occurs as ragged stubby crystals, commonly in discontinuous trains. Many

crystals are altered to chlorite and magnetite. Trace constituents include sphene, magnetite, apatite, zircon, and allanite.

Irregular, discontinuous compositional layering is present locally in the uppermost 10-12 m of the pluton. It parallels the biotite fabric and consists of two types of layers: 1) "normal" monzogranite to granodiorite layers with variable biotite content; and 2) medium- to coarse-grained layers rich in quartz and salmon-colored potassium feldspar. The layers are 5 mm to 3 cm thick and are separated by wispy 1- to 5-mm-thick biotite-rich zones. Coarse pegmatite occurs as pods within the layering and as crosscutting dikes extending into the country rock. One noteworthy pegmatite type in the migmatitic gneiss zone, almost certainly derived from the pluton, contains 80-85% pink potassium feldspar with quartz and minor micas.

Leucocratic Monzogranite

Undeformed leucocratic monzogranite occurring as dikes and as irregularly-shaped bodies up to 2,500 m² in area intrudes all the litho-structural zones in the study area. It cuts rocks as young as the Late Cretaceous biotite monzogranite, but its age relative to the Late Cretaceous two-mica granite unit is uncertain.

This unit consists of pale gray, buff- to brown-

weathering, medium grained leucocratic biotite-magnetite granite. The unit weathers to relatively smooth, rounded outcrops in contrast to the biotite monzogranite which weathers to rougher, grus-covered outcrops. In hand specimen, the rock contains abundant quartz, pale pink, commonly lath-shaped potassium feldspar, and magnetite crystals up to 3 mm in size.

Two thin sections of the leucocratic monzogranite were analyzed. Microcline (40%) is stubby to lath-shaped and subhedral to anhedral. Most crystals are perthitic with wide, discontinuous exsolution lamellae. Quartz (30%) is anhedral with ubiquitous undulatory extinction and common subgrain development. Plagioclase (29%) is anhedral to subhedral and commonly is myrmekitic. The Michel-Levy method gives a plagioclase composition of An 26. Magnetite (0.5%) is anhedral to subhedral and commonly occurs in clusters with minor sphene and chlorite. Accessory minerals are biotite and chlorite, which occur separately and together. Commonly they are associated in loose clusters of randomly-oriented crystals. Trace constituents include sphene, zircon, and rare apatite.

Two-Mica Granite

A two-mica granite pluton intrudes the inverted Paleozoic zone in the western portion of the study area.

The granite cuts Proterozoic, Jurassic, and Jurassic(?) granitoid gneiss. Dikes of two-mica ± garnet granite occur locally at all structural levels and cut the Late Cretaceous biotite monzogranite body at the base of the nappe complex. The two-mica granite pluton has been correlated by K. Howard (per. comm., 1984) with Late Cretaceous two-mica granite plutons of the Old Woman-Piute range.

The pluton intrudes its gneissic country rocks both concordantly and discordantly to S1 foliation. In most outcrops, the granite has at least a weak preferred orientation of biotite, usually parallel to S1 foliation in wallrocks and xenoliths. Locally, the fabric is strongly developed with long axes of feldspar crystals and trains of anhedral quartz paralleling the long axes of biotite crystals.

The most abundant rock type in the unit is white and pink, pale brown-weathering, medium- to fine-grained monzogranite, containing biotite + muscovite ± garnet. Biotite tends to occur in small clusters and trains and seldom makes up more than 5% of the rock. Muscovite content is variable, but it is everywhere subordinate to biotite. Garnets ranging up to 1 mm in diameter are deep red and euhedral to subhedral. Potassium feldspar is reddish-pink and commonly is altered to a red-brown color.

In thin section, a non-garnetiferous sample of the rock contains plagioclase (40%) of composition An 24 (determined by the "a"-normal method), microcline (35%), quartz (20%), biotite (4%), and muscovite (0.5%) with trace amounts of magnetite, zircon, and apatite. Most polysynthetic twin lamellae in plagioclase are straight and parallel, but in a few grains they are bent and wedge-shaped. Quartz crystals show undulatory extinction and many have well-developed subgrains.

The second most abundant rock type is a two-mica + garnet granite containing pale-gray potassium feldspar. The rock differs from the white and pink granite in having slightly more muscovite and more abundant and larger garnets (up to 3 mm in diameter). Intrusive relationships of the two rock types are unknown.

A third rock type that locally cuts the other two is a pale pinkish-gray, fine-grained leucogranite containing pale-pink potassium feldspar, subequal amounts of biotite and muscovite, and as much as 1% garnet. It occurs as dikes and as irregularly-shaped bodies several hundred square meters in area. Many of the two-mica, garnet-bearing dikes at lower structural levels in the nappe complex are identical to the leucogranite.

STRUCTURAL GEOLOGY

The northern Kilbeck Hills have a complex structural history recorded by superimposed megascopic structures and microscopic fabric elements. Overall, the area consists of a stack of thrust and fold nappes composed of L-S tectonites derived from rocks of Proterozoic through Jurassic and Jurassic(?) age. An early set of ductile structures is associated with nappe emplacement and metamorphism. The nappe complex later was intruded by plutons of Late Cretaceous age, which themselves are affected by later episodes of faulting and folding. The observed structural elements can be grouped into five deformational events:

D1. Nappe-forming deformation. This complex event (or events) produced metamorphic foliation (S1) with local preservation of sedimentary bedding (S0). Within this predominantly northwest-dipping foliation is a penetrative down-dip mineral lineation (L1). Open to isoclinal folds (F1) fold the S0 layering and, in part, the S1 foliation. In most places paralleling, but locally cutting, the S1 foliation are ductile, low-angle faults that have juxtaposed the five nappe units. Other ductile, low-angle faults are known or inferred in all of the exposed nappes. At some unknown stage in this

early deformation, large-scale folding overturned the upper plate of the nappe complex (the inverted Paleozoic zone). As the timing and significance of this folding are uncertain, it is not assigned to a separate fold generation but is discussed along with the F1 fold set.

- D2. Brittle, low-angle faulting. Brittle, low-angle faults marked by thin zones of gouge and (or) breccia occur over much of the study area. These faults formed in part through reactivation of ductile faults. Available evidence indicates that the brittle, low-angle faults postdate intrusion of Late Cretaceous plutons.
- D3. F2 northwest-vergent folding. Microscopic to megascopic, northwest-vergent folds deform the early ductile structures and the brittle, low-angle faults. Hinges of F2 microfolds locally produce a lineation (L2) on S1 foliation surfaces.
- D4. F3 folding. Mesoscopic northwest-plunging folds locally refold F2 and earlier structures. Megascopic folds of this generation may account for the large-scale S0 and S1 strike changes within the study area (plate 1).
- D5. High-angle faulting. Mostly northeast- to northwest-trending high-angle faults cut all other structures with the possible exception of the F3 folds.

Pre-D1 Structures

S0 Compositional Layering

Compositional layering derived from sedimentary bedding is present in all Paleozoic and Triassic(?) meta-sedimentary units. Except in the hinge zones of tight to isoclinal F1 folds, the S0 layering parallels the S1 foliation fabric. Although the compositional layering is inherited from depositional lithostratigraphy, in most places it cannot be equated with sedimentary bedding. Unequivocal sedimentary bedding is recognized only in the lower, cross-bedded portion of the metamorphosed Tapeats Sandstone. Cross-bedding in the Tapeats is well-preserved and is everywhere inverted, making it unlikely that any transposition of bedding has occurred through isoclinal folding. In all other Paleozoic metasedimentary units, variable lithologic successions and thicknesses indicate that bedding has been modified tectonically.

The extent to which S0 layering has been transposed into parallelism with S1 foliation is variable. In much of the chaotic zone and in the metamorphosed Bonanza King Formation of the inverted Paleozoic zone, transposition of S0 clearly has occurred through intense F1 folding. Elsewhere, Paleozoic units generally show little evidence of pervasive transposition. In the metamorphosed Bird Spring Formation of the upright Paleozoic zone, F1 folds are scarce and a consistent lithologic sequence of

calcite marble overlain by sandy marble and quartzite is present. In the underlying metamorphosed Redwall Limestone, two groups of metachert bands, although varying in thickness and containing rare hand specimen-scale isoclinal folds, occupy consistent "stratigraphic" positions over much of the unit's strike length. In these cases S0 apparently was deformed primarily by layer parallel shear resulting in ductile attenuation of the strata with relatively limited development of F1 folds and at least local preservation of the units' gross lithostratigraphy.

D1 Structures

S1 Metamorphic Foliation

Metamorphic foliation is the dominant planar fabric element in Proterozoic through Jurassic and Jurassic(?) rocks of the study area (fig. 24). Its degree of development varies from strong and penetrative in micaceous rocks and marble to very weak in most of the Jurassic(?) dioritic intrusive rocks. In metasedimentary rocks, S1 coincides with S0 layering except in the hinge zones of tight to isoclinal F1 folds where S1 cuts across the folded S0 surfaces.

At the microscopic level, S1 primarily is defined by alignment of micas and amphiboles and by shape fabrics in quartz, feldspars, and carbonates. In hand specimen,

Figure 24. Pole density diagram of S1 foliation from the northern Kilbeck Hills (lower hemisphere plot, 173 attitudes). Contour interval: 1%, 3%, 6%, 9%, 12% per 1% area.

foliation is defined by planar alignment of micas, by carbonate and felsic mineral shape fabrics, and, in meta-plutonic rocks, by aligned feldspar augen and lenticular aggregates of quartz and feldspar. In parts of the Mesozoic migmatitic gneiss unit, mafic-felsic compositional banding parallels S1 mineral fabrics and serves to define the foliation at outcrop scale.

Tectonite fabrics in the Jurassic(?) diorite pluton at the north end of the study area are variable. Locally the pluton shows no megascopic evidence of deformation, but in many places a preferred orientation of hornblende is seen. Dikes of the dark Jurassic granodiorite gneiss and the Jurassic(?) granite gneiss have a strongly developed foliation which parallels the weaker fabrics in the dioritic country rock. Rare dioritic dike rocks at lower structural levels also show variable tectonite fabrics. Where they intrude Proterozoic gneiss, the dikes are foliated, but where they intrude Paleozoic marble they commonly retain their igneous textures.

L1 Mineral Lineation

A penetrative, northwest- to west-trending, down-dip mineral lineation is seen on S1 foliation surfaces in most rocks of Proterozoic through Jurassic(?) and Jurassic age (fig. 25). The lineation is best developed in pelitic schists, in orthoschists of Jurassic(?) age, and

Figure 25. Pole density diagram of L1 lineation from the northern Kilbeck Hills (lower hemisphere plot, 102 points). Contour interval: 1%, 4%, 8%, 10%, 12% per 1% area.

in the micaceous quartzite of the metamorphosed Coconino Sandstone. Calcite and dolomite marble are weakly lineated except where they contain abundant prismatic calc-silicate minerals such as tremolite or wollastonite. In general, the lineation is defined by: 1) alignment of tabular and prismatic minerals such as micas, amphiboles, and wollastonite; 2) elongation of anhedral felsic and carbonate minerals; and 3) preferred orientation of elongate feldspar augen and aggregates of felsic minerals. In tightly folded outcrops of the metamorphosed Bonanza King and Bird Spring formations, which contain abundant marker layers, the lineation is defined by alignment of F1 fold hinges (fig. 26).

The L1 lineation is interpreted as a stretching or elongation lineation. In schistose rocks of the study area, boudins in aplite and pegmatite veins lying at a low angle to S1 have long axes orthogonal to L1. In strongly lineated orthoschists, felsic mineral aggregates have the shape of variably-flattened prolate ellipsoids with the long axes of the ellipsoids parallel to L1. These data suggest that L1 records the direction of maximum extension in the D1 deformation.

F1 Folds

Open to isoclinal, northwest- to west-trending folds are seen at all structural levels in the study area (fig.

Figure 26. L1 lineation in the metmorphosed Bird Spring Formation. The lineation is most obvious in isoclinally folded dark-brown-weathering quartzite layers in the wollastonite-bearing marble.

Figure 27. Pole density diagram of F1 fold axes in the northern Kilbeck Hills (lower hemisphere plot, 94 points). Contour intervals: 1%, 4%, 7%, 10% per 1% area.

27). They are present in all pre-Cretaceous units except the Jurassic(?) diorite pluton. The folds, most obvious at hand specimen to outcrop scale, have hinge lines subparallel to the L1 lineation. All F1 folds are similar in style.

Folds of the F1 set can be distinguished from F2 and F3 folds on the basis of style, orientation, and vergence (table 2). In addition, the F1 folds can be divided into two subsets on the basis of vergence and overprinting relationships. Folds of the earlier subset (F1a) are tight to isoclinal, fold S0 layering, and have S1 as an axial planar foliation. Folds of this subset, which are most common in Paleozoic carbonate strata (fig. 28), do not fold mappable stratigraphic or ductile fault contacts. Folds of the later subset (F1b) are open to tight in profile and similar in style with a strong preferred vergence toward the southwest (fig. 29). In most rocks, the F1b folds affect both S0 layering and S1 foliation, but in Paleozoic carbonate strata the F1b folds locally have S1 as an axial planar foliation. Folds of this subset, which may be disharmonic where they fold fault contacts within the chaotic zone, deform both ductile faults and stratigraphic contacts.

The presence in the study area of inward-facing Paleozoic sequences separated by Triassic(?) strata indicates that large-scale, basement-involved folding has

Table 2. Sequence and characteristics of folds in the northern Kilbeck Hills

	<u>F1 FOLDS</u>	<u>F2 FOLDS</u>	<u>F3 FOLDS</u>
Trend of fold axes and lineation	N. 10° W. to S. 85° W.	North to N. 80° E.	N. 15° W. to N. 70° W.
Linear structures associated	L1	L2	None
Preferred sense of vergence	Southwest	Northwest	None
Scale	Micro-to-meso-scale	Micro-to-mega-scale	Meso-to-mega-scale
Character of folds	Tight to isoclinal	Open to tight	Open to gentle
Relationships to S-surfaces	S1 axial planar to Fla. F1b folds S1 in most places	Folds S1	Folds S1
Relationship to metamorphism	Synchronous with amphibolite grade metamorphism (M1)	Post-M1. Either pre-or syn-retrograde metamorphism	Post metamorphic
Distribution	Pervasive, but best developed in carbonate strata	Sporadic overall, but locally well-developed	Sporadic

Figure 28. Isoclinal Fl₁ folds in metachert bands of the metamorphosed Redwall Limestone. Pencil at upper right provides scale.

Figure 29. Cambro-Devonian dolomite marble and banded marble of the metamorphosed lower part of the Bonanza King Formation deformed by tight and isoclinal F1b folds.

occurred. Although the exact orientation and age of the large-scale folds are uncertain, they are correlated with the Fla fold set because: 1) the involvement of Proterozoic basement rocks in the folds and their tight or isoclinal shape indicate that the folds are ductile in style, similar to the Fla subset; and 2) the larger folds must predate asymmetric Flb folds because the latter have the same sense of vergence in both upright and inverted Paleozoic sequences.

Penetrative extensional lineations in metamorphic tectonites have been interpreted to parallel the direction of shear during ductile deformation (Escher and Watterson, 1974). The presence in the northern Kilbeck Hills of multiple ductile thrust faults implies that the D1 deformation in the area involved at least a component of simple shear. In simple shear deformation, linear fabrics such as fold axes become progressively reoriented toward the x- or maximum extension-axis of the finite strain ellipsoid as the x-axis rotates toward the shear direction with increasing finite strain (Bryant and Reed, 1969; Escher and Watterson, 1974). Folds which are reoriented in this manner will become tight, and ultimately isoclinal in shape at high shear strains. The parallel orientation of the L1 lineation and the F1 fold axes in the study area suggests that the F1 fold axes were rotated into parallelism with L1 during the D1

deformation.

The time of development of the F1b folds is uncertain. They are a product of ductile deformation, but most of them developed after regeneration of S1 foliation had ceased. The folds could be entirely unrelated to the D1 deformation, resulting from later top-to-the-southwest simple shear. Alternatively, they might have been produced by a slight reorientation of the shear direction late in the D1 deformation resulting in overturning of late-developing F1 folds.

Ductile, Low-Angle Faults

Ductile, low-angle faults bound and occur within most of the litho-structural zones of the nappe complex. The faults fit the definition of tectonic slides as described by Hutton (1979); they developed under metamorphic conditions, they are associated with large-scale folding, and they were produced during an early phase of the deformational history. Unlike many slides described from the Caledonides of northwest Europe (e.g., Hutton, 1983), the faults in the Kilbeck Hills show no intensification of folding or foliation development as the fault plane is approached.

At most places in the study area, ductile faults parallel S0 layering and S1 foliation with no brecciation or gouge development along contacts. The faults may be

recognized by one or more of the following criteria:

1. Truncation of structures - S0 layering and S1 foliation locally are truncated at shallow angles by ductile faults. At one place in the chaotic zone, a large, outcrop-scale F1 isoclinal fold of S0 having S1 as axial planar foliation is cut by a ductile fault (fig. 30).

2. Missing units - Where faults cut Paleozoic strata they may be recognized if one or more units is missing from the stratigraphic sequence. This criterion must be used with caution in identifying major faults. A missing unit along a contact does not necessarily imply relative movement across the fault. The unit may have been thinned to zero thickness by ductile attenuation produced by large shear strains.

3. Truncation of units along strike - Lithologic units may thin gradually and pinch out along strike, sometimes appearing and disappearing repeatedly. Commonly the truncation of rock types along faults is not accompanied by visible truncation of S0 layering and S1 foliation at the fault contact.

4. Fault slivers - Slivers of anomalous rocks not present in hanging wall or footwall blocks locally are present along fault contacts. The most common rock types in the slivers are Paleozoic carbonate strata, deformed Jurassic(?) intrusive rocks, and calc-silicate or

Figure 30. Isoclinal F1a fold in metamorphosed Red-wall Limestone of the chaotic zone. The fold is truncated above by a ductile fault.

micaceous schists of uncertain age (fig. 31).

In general, low-angle faults at higher levels in the nappe complex have the most consistently ductile style. The fault within the inverted Paleozoic zone is almost everywhere parallel to S0 layering and S1 foliation. It can be recognized by the absence of the metamorphosed Mississippian Redwall Limestone along the contact, by the local presence of slivers of layered calc-silicate rocks and by rare low-angle truncation of S0 and S1.

The faults that bound the gneiss and schist zone are concordant with S0 and S1 and are recognized by the gradual cutting-out of rock types along strike. The two mapped faults within the gneiss and schist zone place rocks of the Jurassic(?) and Triassic(?) gneiss, schist, and quartzite unit against the Triassic(?) garnet-mica schist and siliceous marble and calcareous quartzite units. Both contacts are straight and concordant with S1 and show no intrusive relationships. A single fault sliver of isoclinally folded Paleozoic marble occurs along the lower contact between the Jurassic(?) and Triassic(?) gneiss, schist, and quartzite unit and the Triassic(?) garnet-mica schist unit.

The faults within the chaotic zone are more variable in style. The zone consists of elongate tectonic slivers, of varying sizes, composed mostly of Paleozoic metasedimentary strata. The most abundant rock types in the zone

Figure 31. Fault sliver of calc-silicate schist separating the overlying Cambro-Devonian dolomite marble from the underlying metamorphosed Bird Spring Formation in the inverted Paleozoic zone.

are white marble of the metamorphosed Redwall Limestone, tan dolomite of the Cambro-Devonian dolomite marble unit, and banded calcite marble of the metamorphosed lower part of the Bonanza King Formation. Most of the faults within the chaotic zone are concordant, but low-angle truncation of S0 and S1 is much more common than at higher levels in the nappe complex. Commonly such truncation occurs only at the ends of fault slivers (fig. 32). Ductile faults of the chaotic zone cut Fla isoclinal folds but are folded by southwest-vergent Flb folds.

The fault at the base of the chaotic zone is ductile in character with local brittle reactivation (fig. 33). Along its ductile portions, the fault generally parallels lithologic layering and S1 foliation at the outcrop scale; however, units at the top of the underlying upright Paleozoic zone undergo significant changes in thickness over distances of 100 m or less. Despite the parallelism of the fault to S1 foliation, these thickness changes probably are due to excision of strata by the fault rather than to abrupt variations in ductile attenuation of the upright Paleozoic zone strata.

The large number of ductile low-angle faults in the Kilbeck Hills makes it difficult to correlate a single fault in the study area with the Scanlon thrust of the Old Woman Mountains and Little Piute Mountains. In the latter ranges, the Scanlon thrust places inverted

Figure 32. Dark micaceous schist and tan Cambro-Devonian dolomite marble truncated beneath a ductile fault in the chaotic zone. The fault is deformed by a northwest-vergent F2 fold which plunges northeast, away from the viewer.

Figure 33. Brittle faulting along the chaotic zone-upright Paleozoic zone contact. Shallow-dipping marble of the chaotic zone at top and left is faulted against steeply-dipping rocks of the metamorphosed Bird Spring Formation at lower right.

Proterozoic or Cambrian rocks as young as the metamorphosed lower part of the Bonanza King Formation over upright strata of Cambro-Devonian to Triassic age. The footwall strata commonly are tightly folded so that the upper plate locally rests on inverted Permo-Pennsylvanian to Triassic rocks. In the Kilbeck Hills, the fault most likely correlative to the Scanlon thrust is the one which places Cambro-Devonian dolomite marble over metamorphosed Bird Spring Formation strata in the inverted Paleozoic zone. Only one Paleozoic unit is missing along this fault, however, in contrast to the three or more units which typically are cut out by the Scanlon thrust. This could imply that the Scanlon thrust dies out along strike toward the southwest.

Alternatively, if the Scanlon thrust cuts up section (down structurally) toward the southwest to include rocks as young as Permian in the upper plate, it could be correlative with the ductile fault separating the inverted Paleozoic zone from the Triassic(?) and Jurassic(?) rocks of the underlying gneiss and schist zone. In either case, it appears that movement in the Kilbeck Hills was not concentrated on a single fault but was distributed between a number of low-angle faults.

The concentration of low-angle faults in the chaotic zone and the juxtaposition of multiple fault slivers in it indicate that the zone experienced a different style

of faulting from the rest of the nappe complex and that faulting was concentrated there at some point in the deformation history. The zone apparently results from movement on a series of closely-spaced ductile faults which cut across stratigraphic contacts. The presence of upright Permian strata at the top of the zone implies that the faults cut through the upright Paleozoic zone.

The chaotic zone apparently developed late in the D1 deformation episode. The relative frequency of low-angle truncation of S0 and S1 in the zone suggests that ductile faulting continued beyond the time when increased strain rate and (or) decreased temperatures began to prevent recrystallization and S1 foliation development from keeping pace with fault movement. If this interpretation is correct, the chaotic zone may have formed in an area in which fault movement was concentrated in the late stages of D1 deformation, after isoclinal F1a folding, but prior to the development of the asymmetric F1b folds.

D2 Deformation

Brittle, Low-Angle Faults

Brittle, low-angle faults overprint the ductile D1 structures in the chaotic and upright Paleozoic zones. Although the brittle structures often are difficult to identify on weathered surfaces, they are obvious on the fresh exposures in marble quarries. They consist of

irregular, discontinuous, low-angle faults and fractures, the former marked by narrow zones of gouge and breccia up to 30 cm thick. These features locally occur along older ductile faults such as the fault between the chaotic and upright Paleozoic zones (fig. 33).

The only mappable fault that shows a brittle deformation style throughout its length is the one separating the upright Paleozoic zone from the underlying migmatitic gneiss zone. In many places this contact is marked by up to 2 m of gouge and brecciated rock (fig. 34) with a red, iron oxide stain.

D3 Deformation

F2 Folds

Northeast-plunging F2 folds are present over much of the study area (fig. 35). They occur as crenulations and megascopic folds having amplitudes of 1 mm to 30 m (fig. 36). They fold the F1 fold set, the L1 lineation, and both ductile and brittle low-angle faults. F2 folds also fold the intrusive contact between the migmatitic gneiss zone and the Late Cretaceous monzogranite pluton.

Macro-scale F2 folds are concentric to similar in style and gentle to tight in shape. Asymmetric F2 folds are almost invariably northwest-vergent. No evidence was found for an axial planar foliation associated with the

Figure 34. Brittle, low-angle fault at the base of the upright Paleozoic zone. The fault places shallow-dipping rocks of the metamorphosed Bird Spring Formation over metamorphosed Hermit Shale which dips moderately to the right. The rocks beneath the fault are part of a ductile fault sliver in the Mesozoic migmatitic gneiss unit.

Figure 36. View toward the northeast of northwest-vergent F2 folds in the inverted Paleozoic zone.

F2 folds in micaceous rocks. Locally, carbonate strata have a spaced cleavage in the hinge zones of F2 folds. No offset of layering is seen across these cleavage surfaces.

Micro-scale F2 folds occur only in micaceous rocks, being most prominent in the Triassic(?) garnet-mica schist unit (fig. 37). They consist of crenulations of the S1 foliation with an amplitude rarely exceeding 3 mm. The crenulations commonly occur in conjugate sets separated by angles of 60-70°. S1 foliation usually bisects the obtuse angle between the sets. In all cases observed in thin section, micas are bent by the crenulations and have not undergone subsequent recrystallization. The local presence of chlorite in the crenulation hinge zones suggests that the F2 folds developed during retrograde metamorphism.

L2 Lineation

A discontinuous crenulation lineation, L2, is seen in pelitic schists and locally in Jurassic(?) ortho-schist. The lineation is formed by alignment of F2 micro-fold hinges in the S1 foliation. Like the F2 folds, the L2 lineation generally is northeast-trending except where reoriented by F3 folds. L2 typically is oriented at 45-70° from the L1 mineral lineation on S1 foliation surfaces. Locally, especially in the Triassic(?)

Figure 37. F2 microfolds in the Triassic(?) garnet-mica schist. Note the bending of mica crystals around fold hinges (crossed nicols, 1 x 40).

garnet-mica schist unit, L2 is more obvious in outcrop than is L1.

D4 Deformation

F3 Folds

Northwest-plunging, gentle to open F3 folds are developed sporadically over much of the study area (fig. 38). They fold all other structures with the possible exception of the brittle, high-angle faults. Hand-specimen-scale F3 folds are very rare; most examples have wavelengths of at least 3-4 m. All of the F3 folds observed in outcrop are concentric in style. At the largest scale, the F3 folds appear to be responsible for the large-scale strike changes in Paleozoic and Triassic(?) strata. However, earlier structures, such as the L1 lineation and the F1 folds, show little or no systematic reorientation from one limb to the other of these large F3(?) folds.

D5 Deformation

High-Angle Faults

Brittle, high-angle faults are the youngest structures recognized in the study area. Most of the faults are northwest- to northeast-striking and have dip angles of at least 80°. They are recognized by offset or truncation of layering in stratified rocks and (or) by

Figure 38. Orientation diagram of F3 fold axes from the northern Kilbeck Hills (lower hemisphere plot, 7 points).

straight, abrupt contacts between intrusive rock units. None of the faults in the study area cuts Quaternary alluvium.

The most extensively exposed high-angle fault cuts rocks as young as Jurassic(?) in the northwestern part of section 16, T3N, R16E. A down-to-the-west sense of movement on this fault is inferred from offset of northwest-dipping Paleozoic strata of the inverted Paleozoic zone in section 17, T3N, R16E.

Microtextures and Microfabrics

Microscopic fabrics in the study area are defined by the elongation of aggregates and individual crystals of felsic and carbonate minerals. Quartz shows the most consistent and strongly-developed shape fabrics. Quartz crystals in Jurassic(?) and Proterozoic metaplutonic rocks may be up to 1.5 cm long with length:width ratios of 10:1 or greater in sections cut parallel to L1. Carbonate minerals and smaller feldspar crystals usually have at least a weak shape fabric paralleling S1, but only rarely do they show the same degree of elongation as quartz. Feldspar megacrysts may be rounded, blocky, or elongate. Rounded and blocky microcline augen frequently are fractured, with finer-grained tails of quartz and feldspars. Elongate augen also may be broken, with the segments often separated by straight, quartz-filled

fractures. None of the thin sections examined in this study have any recognizable crystallographic fabrics in felsic or carbonate minerals.

Polygonal textures are very common in metamorphic rocks of the northern Kilbeck Hills. They occur most frequently in plagioclase feldspar, but are common also in quartz, potassium feldspar, and carbonate minerals. Often, polygonal textures occur in elongate monomineralic aggregates of quartz or feldspar which appear to have recrystallized from a single elongate grain. Individual polygonal grains are themselves slightly elongate in many cases, with grain long axes parallel to the L1 lineation.

Micro-scale strain features are present in all rocks of the study area. Undulatory extinction is ubiquitous in quartz and common in feldspars. Subgrains and deformation bands commonly are seen in quartz crystals. Deformation twinning is very common in carbonate minerals and is well-developed locally in plagioclase, especially in diorites and amphibolites. Undulatory extinction, subgrains, and deformation bands occur in both elongate and equant, polygonal quartz grains.

The data presented above suggest the following three stage microstructural history:

- 1) The strong shape fabrics in grains and aggregates of felsic and carbonate minerals indicate that pre-Cretaceous rocks underwent an

early episode of intense strain probably dominated by crystal-plastic deformation at the grain scale. The fact that elongation is greatest parallel to the L1 lineation implies that shape fabrics developed during the D1 deformation.

- 2) The widespread development of polygonal textures and the apparent lack of crystallographic fabrics in felsic and carbonate minerals imply that D1 textures and fabrics were modified by static recrystallization. Such recrystallization may have resulted in grain growth in finer grained aggregates originally produced by dynamic recrystallization of highly strained larger grains. Static recrystallization may have been driven by heat introduced by Late Cretaceous plutons.
- 3) The presence of undulatory extinction, subgrains, and deformation bands in both elongate and polygonal quartz grains indicates that the rocks underwent post-recrystallization strain. The timing of this grain-scale strain is uncertain, but it must postdate the emplacement of Late Cretaceous plutons because the plutons commonly show subgrain development in quartz,

and limited deformation twinning in plagioclase feldspar. In older rocks, strain in elongate, unrecrystallized grains may result from superposition of D1 and post-plutonic deformation features.

METAMORPHISM

Rocks of the northern Kilbeck Hills record a single phase of prograde metamorphism. Petrographic study of mineralogy and fabric relationships indicates that peak metamorphic conditions were synchronous with D1 deformation and produced lower amphibolite grade mineral assemblages. No conclusive evidence was found in this study for relict, pre-D1 metamorphic minerals. The amphibolite grade assemblages locally have undergone retrograde greenschist facies metamorphism. Regional geochronological data imply that the syn-D1 metamorphism is of Late Cretaceous age (Foster and others, 1988). The age of retrograde metamorphism is uncertain.

Summary of Metamorphic Mineral Assemblages

Lower amphibolite grade metamorphism is indicated by: 1) the local occurrence of sillimanite + muscovite in the Triassic(?) garnet-mica schist unit, 2) the widespread occurrence of diopside without forsterite in the metamorphosed Bird Spring Formation and Hermit Shale, and 3) the occurrence of hornblende + calcic andesine in rare amphibolite pods in the migmatitic gneiss unit.

Metamorphic assemblages are summarized below. In each assemblage, minerals are listed in order of decreasing abundance. Assemblages are listed in order of

decreasing abundance within mapped units. The word "metamorphosed" is omitted from the names of Paleozoic units. Abbreviations of common minerals include:
 cc=calcite, Kspar=potassium feldspar, hbl=hornblende, plag=plagioclase feldspar, qz=quartz, woll=wollastonite.

Tapeats Sandstone (Cambrian)

qz + Kspar + plag + biotite + muscovite

Bright Angel Shale (Cambrian)

muscovite + biotite + qz + plag + Kspar

Bonanza King Formation - lower part (Cambrian)

cc + Mg-chlorite + epidote

Dolomite marble (Cambrian and (or) Devonian)

dolomite + cc + phlogopite + tremolite + Mg-chlorite

Redwall Limestone (Mississippian)

cc + qz

Bird Spring Formation (Permo-Pennsylvanian)

woll + cc + diopside ± qz ± Kspar ± plag

cc + qz + phlogopite + Kspar ± dolomite

cc + diopside + qz + idocrase + Kspar + plag

woll + Kspar + diopside + plag + garnet + epidote

+ cc

qz + cc + garnet + idocrase + plag + diopside

Hermit Shale (Permian)

woll + diopside + Kspar + qz ± cc ± plag

Coconino Sandstone (Permian)

qz + muscovite + cc + Kspar + plag

Kaibab Limestone (Permian)

cc + dolomite + tremolite + Mg-chlorite + phlogo-
pite + Kspar + qz

cc + dolomite + tremolite + Mg-chlorite + Fe-chlorite
+ qz

Garnet-Mica Schist (Triassic(?))

biotite + muscovite + qz + plag + garnet ± Kspar ±
sillimanite ± chlorite

qz + cc + Kspar + diopside + plag + garnet + biotite
± chlorite

Siliceous Marble and Calcareous Quartzite (Triassic(?))

cc + qz + diopside + Kspar + plag + scapolite
± tremolite

qz + Kspar + diopside + cc + plag + biotite

Gneiss, Schist, and Quartzite (Triassic(?) and Juras-
sic(?))

plag + Kspar + qz + hbl + epidote + biotite
± diopside

Kspar + qz + hbl + plag + diopside + scapolite
+ epidote

scapolite + Kspar + diopside + hbl + epidote
+ sphene ± plag ± cc

qz + cc + plag + diopside + epidote + garnet + hbl
+ Kspar

qz + plag + biotite + muscovite ± chlorite

Migmatitic Gneiss (Mesozoic)

qz + plag + biotite + muscovite

plag + Kspar + qz + biotite + garnet

hbl + plag (An 44) + qz + magnetite

Age of Metamorphism

Geochronological studies by Foster and others (1987, 1988) in the Old Woman-Piute range indicate that regional metamorphism slightly predated Late Cretaceous plutonism. They reported $^{40}\text{Ar}/^{39}\text{Ar}$ release spectra on hornblende from Proterozoic rocks. Their data show that partial to complete outgassing of argon occurred at 80-85 Ma at temperatures near to or greater than 500° C. These ages contrast with hornblende $^{40}\text{Ar}/^{39}\text{Ar}$ ages of 73-75 Ma from the Old Woman granodiorite. Biotite and potassium feldspar from the same pluton give $^{40}\text{Ar}/^{39}\text{Ar}$ ages of 71 and 69 Ma, respectively (Foster and others, 1988). The hornblende ages are interpreted as crystallization ages by Foster and coauthors, whereas the biotite-potassium feldspar ages are interpreted as cooling ages recording the time at which temperatures in the pluton decreased through the interval 300-250° C.

Microfabric data from the Kilbeck Hills indicate clearly that peak metamorphic conditions were largely synchronous with D1 deformation. In the Triassic(?)

garnet-mica schist unit, fibrolitic sillimanite crystals show a strong preferred orientation subparallel to the L1 lineation. Garnet porphyroblasts in the same unit contain straight to curved inclusion trails of quartz that parallel the external S1 fabric at garnet grain boundaries. Included quartz crystals are finer-grained than those outside the garnets; this implies that the garnets grew during coarsening of the quartz and micas which define S1 but after the initial development of the S1 fabric. Rare S-shaped inclusion trails imply that the garnets and (or) the external S1 fabric rotated during garnet crystallization.

Available evidence suggests that diopside is a syn- to post-D1 mineral. It occurs in a variety of sizes from small, interstitial grains to porphyroblasts up to 6 mm in diameter. At all sizes, the crystals are equant to slightly elongate parallel to S1, and are anhedral to subhedral. Some of the larger diopside crystals in calc-silicate schists of the metamorphosed Bird Spring Formation have their long axes and c crystallographic axes oriented at a high angle to S1 foliation. The extent and significance of this preferred orientation are uncertain, but the crystals appear to have grown in their present orientation rather than having been rotated into it because they lack fine-grained "tails" and show no evidence of fracture or internal strain. In calcareous

schists and quartzites from several units, diopside locally is skeletal with inclusions of equant, commonly polygonal, calcite, quartz, or feldspars. These relationships imply that diopside crystallization and grain growth outlasted D1 deformation and may have continued after the recrystallization of D1 microfabrics.

Significance of Wollastonite

Wollastonite is abundant locally in the metamorphosed Bird Spring Formation and in the metamorphosed Hermit Shale. It is present: 1) in all of the tectonic inclusions of Upper Paleozoic rocks in the migmatitic gneiss zone; 2) in the metamorphosed Bird Spring Formation and Hermit Shale in the upright Paleozoic zone in the northwest corner of section 21, T 3N, R16E; and 3) in the thick Bird Spring sequence of section 10, T3N, R16E. Elsewhere, the metamorphosed Bird Spring Formation and Hermit Shale, as well as the metamorphosed Redwall Limestone and Cocconino Sandstone, contain quartz and calcite with no evidence of reaction.

In hand specimen and in thin section, wollastonite shows a preferred orientation with the b crystallographic axis subparallel to the L1 lineation. Very locally, however, massive wollastonite layers are present in the metamorphosed Bird Spring Formation. These layers contain up to 70% wollastonite which occurs in radiating clusters

along with interstitial calcite and small sub-circular clusters of diopside crystals. Thin section study of the metamorphosed Bird Spring Formation reveals variations in the extent to which quartz and calcite reacted to form wollastonite. Layers containing wollastonite and one reactant locally alternate with layers which have all three minerals in mutual contact (fig. 12). These observations suggest that either temperature or the fluid phase composition changed before the reaction could proceed to completion in all layers.

Two interpretations have been proposed for the regional development of wollastonite in Permo-Pennsylvanian Supai Group-Bird Spring Formation rocks. Hamilton (1982) interpreted massive wollastonite beds in the Supai Group of the Big Maria Mountains as a product of Jurassic contact metamorphism. In contrast, Hoisch (1987) interpreted wollastonite in the Big Maria Mountains as a product of syntectonic Late Cretaceous regional metamorphism at mid-crustal depths. Hoisch (1987) estimated pressures to be approximately 3.0 ± 1 kb and concluded from calcite-dolomite and two feldspar thermometry that regional metamorphic temperature ranged from 430-590° C. Only extremely low partial pressures of CO_2 would allow the quartz + calcite = wollastonite + CO_2 reaction to proceed under these pressure-temperature conditions. Hoisch postulated that large fluxes of hot, H_2O -rich

fluids raised temperatures as much as 300° above those predicted by the steady-state geotherm and allowed the wollastonite-forming reaction to proceed by continuously removing CO₂ from the system. Although these large fluid fluxes are recorded by reactions only in the Supai Group, Hoisch (1987) concluded from the presence of wollastonite in structurally isolated Supai exposures that the fluids must have moved through adjacent units as well. He cited the presence of quartz-filled, crosscutting fractures in adjacent units as possible evidence for passage of metamorphic fluids.

Available field data on wollastonite-bearing rocks of the Kilbeck Hills do not provide unequivocal support for the model of Hoisch (1987) or that of Hamilton (1982). The only wollastonite-bearing rocks that actually are cut by Jurassic(?) metaplutonic rocks are in the thick section of the metamorphosed Bird Spring Formation exposed in section 10, T3N, R16E. Wollastonite-bearing slivers of Upper Paleozoic rocks in the migmatitic gneiss zone locally are in tectonic contact with orthogneiss derived from Jurassic(?) granitoids, but the relative positions of the two rock types in Jurassic time are unknown. The local development of greisen and skarn in Paleozoic carbonates near contacts with Jurassic and Jurassic(?) intrusive rocks and the common saussuritization of plagioclase in the intrusive rocks themselves suggest the possibility

that the H₂O-rich fluid flux envisioned by Hoisch (1987) could have been associated with emplacement of Jurassic intrusive rocks. Whatever its origin, the fluid flow produced wollastonite only in the Bird Spring Formation and the Hermit Shale. In the northwest corner of section 21, T3N, R16E, wollastonite-bearing rocks of the metamorphosed Bird Spring Formation and Hermit Shale rest positionally on metamorphosed Redwall Limestone containing calcite + quartz and lacking wollastonite.

Microtextures in the metamorphosed Bird Spring Formation seem to support a syn-D1, Late Cretaceous age for wollastonite development. The common parallelism of wollastonite crystals to the L1 lineation is permissive of either a pre- or syn-D1 age, but the presence of radiating clusters of wollastonite in several thick layers of high wollastonite content suggests that it is in part a post-D1 mineral. In view of the intense D1 deformation undergone by Paleozoic carbonate strata, it seems unlikely that any competency contrasts brought about by relatively minor differences in mineral percentages could allow radiating clusters of pre-tectonic wollastonite to avoid the D1 deformation in certain layers. A more reasonable interpretation is that wollastonite development locally outlasted Late Cretaceous D1 deformation.

Metamorphic Temperature and Pressure

Calcium silicate mineral assemblages and the presence of sillimanite can provide loose constraints on peak metamorphic temperatures in the study area. The exact temperature values inferred from phase equilibrium experiments will vary with pressure and, in the case of devolatilization reactions, with fluid phase composition. Pressure of crystallization of the amphibolite grade metamorphic assemblages may be estimated using available geochemical data on depth of emplacement of Late Cretaceous plutons combined with field observations in the study area and in the Old Woman Mountains.

Geobarometry and petrologic studies on depth of emplacement of peraluminous granitoids allow general estimates of crystallization pressure of Late Cretaceous plutons in the study area. Young and Wooden (1988) used Al-hornblende, phengite-white mica, and plagioclase-garnet-muscovite-biotite barometers on Late Cretaceous granodiorite and two-mica granite plutons of the Old Woman Mountains. Their estimates of crystallization pressure ranged from 5 kb to 7 kb. Zen (1988) summarized available data on the depth of emplacement of muscovite-bearing granites and concluded that abundant fluorine, as reflected by the presence of fluorite and (or) topaz, is the only component that could allow primary plutonic muscovite to crystallize at pressures under 3 kb. No fluorite or

topaz was observed in two thin sections of two-mica granite from the study area and neither mineral was reported by C. Miller and others (1982) as an accessory in two-mica granites of the Old Woman-Piute range.

Field observations and metamorphic mineral assemblages indicate that the Old Woman Mountains experienced higher metamorphic temperatures than did the Kilbeck Hills. C. Miller and others (1982) reported the presence of sillimanite + potassium feldspar in the metamorphosed Bright Angel Shale along much of the Scanlon thrust zone in the Old Woman Mountains. The author has observed forsterite in metamorphosed Kaibab Limestone at Scanlon Gulch in the western Old Woman Mountains. Grain sizes in micas and wollastonite at Scanlon Gulch are coarser than any observed in the Kilbeck Hills. These data indicate that much of the Scanlon thrust zone in the Old Woman Mountains was metamorphosed at upper amphibolite grade conditions. Although the temperature difference between the two areas may be due in part to a lower late Cretaceous geothermal gradient in the Kilbeck Hills, part also may be a result of lower metamorphic pressures there. However, pressure differences were not great enough to produce a difference in intrusive styles of Late Cretaceous plutons in the two areas. Nowhere did the plutons produce a contact aureole in the country rocks (C. Miller and others, 1982). The plutons which intrude the Scanlon

thrust are at least in part concordant bodies, although they generally have sharp contacts at the hand specimen scale. According to the classification of Buddington (1959) and Hyndman (1985), these characteristics indicate that the plutons are deep level mesozonal intrusions.

The data presented above suggest that late Cretaceous lithostatic pressure in the study area was at least 3 kb but probably no greater than 7 kb. Thus, a range of 3-7 kb is used in deducing minimum and maximum metamorphic temperatures from P-T and T-X phase diagrams.

The presence of sillimanite in the Triassic(?) garnet-mica schist provides a minimum temperature value at any given pressure. The polymorphic transformations of andalusite or kyanite to sillimanite are independent of fluid phase composition. Figure 39 shows the experimentally determined P-T stability relations of the Al_2SiO_5 polymorphs, as presented by Holdaway (1971). The absolute temperature minimum for sillimanite is approximately 500° C at 3.8 kb. Temperatures for sillimanite stability increase with increasing and decreasing pressure from the invariant point, reaching values of approximately 560° C at both 3 kb and 5 kb pressure and approximately 660° C at 7 kb pressure. At 4 kb pressure near the invariant point, the kyanite-sillimanite phase boundary lies at approximately 510° C.

Mineral assemblages in siliceous dolomitic marbles of

Figure 39. P-T diagram showing phase relations of Al_2SiO_5 polymorphs (after Holdaway, 1971). Dashed lines indicate minimum temperatures for sillimanite stability at 3, 4, and 5 kb.

the metamorphosed Kaibab Limestone, the metamorphosed Bird Spring Formation, and the Cambro-Devonian dolomite unit ideally should provide close constraints on maximum and minimum temperatures of regional metamorphism. Because most prograde reactions in these rocks involve release of volatile constituents, they commonly are portrayed on diagrams plotting temperature versus fluid phase composition. A schematic T-X CO₂ diagram showing phase assemblages and reaction boundaries in the system CaO-MgO-SiO₂-H₂O-CO₂ is shown in figure 40. . .

Two factors complicate temperature estimates based on this system. The first stems from disagreement over the positions of several of the relevant reaction boundaries. Eggert and Kerrick (1981) reported that two internally consistent data sets give different positions for reactions 6, 8, and 9. Experiments of Metz (1976) and Kase and Metz (1980) using synthetic starting materials yield a 4 kb T-X CO₂ value for invariant point C of approximately 670° C at 73% CO₂ in the fluid phase. In contrast, Eggert and Kerrick (1981) cited unpublished experimental data of G. K. Jacobs who used natural minerals which gave a 4 kb T-X CO₂ value for invariant point C of approximately 625° C at 91% CO₂ in the fluid phase.

The second complicating factor results from the presence of wollastonite in the metamorphosed Bird Spring Formation if we assume that wollastonite was a product of

Figure 40. T-X CO₂ diagram at 3 kb showing selected reaction boundaries and invariant points in the siliceous dolomite system. Abbreviations: cc=calcite; dol=dolomite; diop=diopside; fo=forsterite; qz=quartz; trem=tremolite. Numbered reaction boundaries are:

- (1) $5 \text{ talc} + 6 \text{ cc} + 4 \text{ qz} = 3 \text{ trem} + 6 \text{ CO}_2 + 2 \text{ H}_2\text{O}$
- (2) $2 \text{ talc} + 3 \text{ cc} = \text{trem} + \text{dol} + \text{CO}_2 + \text{H}_2\text{O}$
- (3) $5 \text{ dol} + 8 \text{ qz} + \text{H}_2\text{O} = \text{trem} + 3 \text{ cc} + 7 \text{ CO}_2$
- (4) $\text{trem} + 3 \text{ cc} + 2 \text{ qz} = 5 \text{ diop} + 3 \text{ CO}_2 + \text{H}_2\text{O}$
- (5) $\text{dol} + 2 \text{ qz} = \text{diop} + 2 \text{ CO}_2$
- (6) $\text{trem} + 3 \text{ cc} = \text{dol} + 4 \text{ diop} + \text{CO}_2 + \text{H}_2\text{O}$
- (7) $\text{cc} + \text{qtz} = \text{woll} + \text{CO}_2$
- (8) $\text{trem} + 11 \text{ dol} = 8 \text{ fo} + 13 \text{ cc} + 9 \text{ CO}_2 + \text{H}_2\text{O}$
- (9) $3 \text{ dol} + \text{diop} = 4 \text{ cc} + 2 \text{ fo} + 2 \text{ CO}_2$

Data sources: Reaction (7) derived from Winkler (1979, fig. 9-7) based on experimental data of Greenwood (1967) and Harker and Tuttle (1956). Reaction (8) based on experimental data of Metz (1976). All other reaction boundaries interpolated from 2 kbar and 4 kbar data of Eggert and Kerrick (1981, fig. 8).

syn-D1 regional metamorphism and that all units reached the same peak metamorphic temperatures. For wollastonite to form at pressures of 3-7 kb requires either very low X_{CO_2} in the fluid phase or temperatures more appropriate to upper amphibolite or granulite grade than to lower amphibolite grade metamorphism. Mineral assemblages in other units indicate that such high temperatures were not reached during the syn-D1 metamorphism. If a CO_2 -poor fluid phase was present in the Bird Spring Formation, was it also present in other units? The fact that wollastonite-bearing rocks of the metamorphosed Bird Spring Formation locally are in stratigraphic contact with metamorphosed Redwall Limestone in which calcite and quartz did not react implies that fluid compositions during the syn-D1 metamorphism were not uniform. The uncertainty of fluid phase composition diminishes the value of the siliceous dolomitic marbles in determining metamorphic temperatures from the T-X diagrams.

Despite these difficulties, several general constraints can be placed on temperatures and fluid phase compositions based on metamorphic assemblages in the siliceous dolomitic marbles of the study area (fig. 40). In the Cambro-Devonian dolomite marble and the metamorphosed Kaibab Limestone, the presence of tremolite indicates that fluid phase composition was to the left of invariant point B and tremolite without talc, quartz, diopside, or

forsterite, indicates that peak metamorphic temperatures in all units were above those of reactions 2 and 3 but below those of reactions 6 and 8. Invariant point C is an overall temperature maximum.

In the metamorphosed Bird Spring Formation, the assemblage diopside + calcite + quartz without tremolite indicates that temperatures were above those of reaction 4. In addition, the presence of wollastonite implies that the maximum possible temperature and $X \text{CO}_2$ lie at the intersection point of reactions 7 and 8. At 3 kb fluid-pressure this intersection point gives approximate maxima of 575°C and $0.15 X \text{CO}_2$. Extrapolation of the reaction curves of Eggert and Kerrick (1981, p. 1046) suggests that at 6 kb the intersection point would occur at temperatures between 600°C and 650°C and at $X \text{CO}_2$ values of less than 0.1. Thus 650°C would be an overall 6 kb temperature maximum if fluid phase composition were the same in wollastonite-bearing calcite marble and forsterite-free siliceous dolomitic marble. If fluid phase compositions were not the same in the two rock types the overall temperature maximum would again lie at invariant point C.

Retrograde Metamorphism

Retrograde metamorphism is recognizable only in pelitic rocks of the Triassic(?) garnet-mica schist and the metamorphosed Bright Angel Shale. It is indicated by

chlorite crystals up to 1 mm in diameter which replace both garnet and biotite. In the garnet-mica schist, chlorite occurs along fractures in garnet porphyroblasts. It also occurs locally as undeformed subhedral plates oriented with their (001) cleavage planes at a high angle to S1 foliation. In both units, aggregates of undeformed anhedral chlorite crystals locally replace biotite in the hinge zones of F2 microfolds.

The above relationships indicate that chlorite in the pelitic rocks grew after the D1 deformation and after the development of prograde biotite and garnet. The occurrence of undeformed chlorite in the hinge zones of F2 microfolds indicates that retrograde chlorite developed late in or after the D3 deformation.

GEOLOGIC HISTORY

Intrusion of granitic plutonic rocks correlative with the mid-Proterozoic Fenner gneiss is the earliest event recorded in the Kilbeck Hills. These coarsely megacrystic granodiorites and granites, the wallrocks of which are not preserved in the study area, were intruded by younger Proterozoic granites. None of the Proterozoic rocks retain any clear evidence of Proterozoic deformation or metamorphism. Prior to the Cambrian they were uplifted and eroded and were then covered by a carbonate and clastic Paleozoic and Triassic(?) sedimentary sequence of transitional cratonal-miogeoclinal character.

The Proterozoic, Paleozoic, and Triassic(?) rocks were first cut by dikes, sills, and plutons of dioritic to granitic composition. One member of this suite was intruded in early Late Jurassic time and others are similar compositionally to dated Jurassic plutonic rocks exposed in nearby ranges. The depth of emplacement of Jurassic and Jurassic(?) intrusive rocks is uncertain.

By Late Cretaceous time the rocks of the Kilbeck Hills, as part of the Old Woman-Piute structural block, were at mid-crustal levels. All of the pre-Cretaceous rocks of the study area underwent intense ductile shearing at lower amphibolite grade metamorphic conditions, resulting in attenuation of the Paleozoic units to a small

fraction of their initial stratigraphic thickness. Large-scale folding led to the development of at least one basement-cored fold nappe and hinge lines of smaller-scale open to isoclinal folds were rotated progressively toward the penetrative down-dip stretching lineation. Late in this deformational event (D1) a complex of ductile low-angle faults developed. During the last stages of ductile faulting, movement was concentrated in a narrow zone (the chaotic zone) in the upright Paleozoic rocks beneath the inverted limb of the fold nappe. The ductile faults were folded by southwest-vergent folds which may have been a product of a separate deformation event or a result of reorientation of the D1 shear direction near the end of D1 deformation.

After most if not all of the D1 deformation had ceased, but while the Jurassic and older rocks were still at mid-crustal levels, Late Cretaceous monzogranite and two-mica granite plutons were intruded. Heat from the plutons caused extensive recrystallization in the metamorphic country rocks producing widespread polygonal textures and obliterating D1 crystallographic fabrics in felsic and carbonate minerals. After the intrusion of the monzogranite and two-mica granite plutons, the rocks were cut by brittle, low-angle faults.

After low-angle faulting had ceased, the rocks were deformed by northwest-vergent F2 folds and then by

northwest-plunging F3 folds. The youngest recorded deformation produced mostly northwest- to northeast-striking, high-angle faults, the largest of which had a down-to-the-west sense of offset. None of the faults cuts Quaternary alluvium.

At some point in latest Cretaceous and (or) earliest Tertiary time, the Kilbeck Hills were uplifted along with the Old Woman-Piute range. $^{40}\text{Ar}/^{39}\text{Ar}$ geochronological studies indicate that the northern Old Woman Mountains, which were at about the same structural level as the Kilbeck Hills, were within a few kilometers of the surface by 65 Ma (Foster and others, 1988). D2 brittle, low-angle faults or D5 high-angle faults may have been a product of this uplift event. In the latter case, brittle low-angle faults as well as F2 and probably F3 folds must all have been produced between the time of emplacement of Late Cretaceous plutons and the beginning of latest Cretaceous-earliest Tertiary uplift.

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U·M·I

PLATE 1

GEOLOGIC MAP OF THE NORTHE SAN BERNARDINO COUNTY,

by

Eric D. Horrington

1989

NORTHERN KILBECK HILLS

DUNTY, CALIFORNIA

Lorrington

39

SCALE 1 : 4975

0.

0.

CONTO

DATUM

1200

4975

_____ 0.5 miles

_____ 1 kilometer

CONTOUR INTERVAL : 40 ft

DATUM IS MEAN SEA LEVEL

1989

Approx. Mean

Declination, 1985

CONTOUR

DATUM IS

CONTOUR INTERVAL : 40 ft

DATUM IS MEAN SEA LEVEL

INTERVAL : 40 ft

MEAN SEA LEVEL

8

Jm

Chubbuck

BM 1014

Topographic base
from U.S. Geol. Survey
1 : 62,500
Cadiz Lake ; 1956
Quadrangle

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MAP SYMBOLS

Stratigraphic and intrusive contacts

High-angle fault; bar and ball on downthrown side

Thrust fault — teeth on upper plate;

dashed where inferred;

dotted where concealed;

queried where contact between stratigraphically adjacent metasedimentary units is interpreted as a low-angle fault

31

Anticline or antiform

42

Syncline or synform

10

Overturned anticline

29

Asymmetric fold — showing sense of asymmetry

14

Isoclinal fold

70

Strike and dip of overturned beds — shown in cross-bedded portions of the metamorphosed Tapeats Sandstone

48

Strike and dip of foliation

DESCRIPTION OF MAP UNITS

ALLUVIUM (QUATERNARY)
WINDBLOWN SAND (QUATERNARY)
TWO-MICA GRANITE (LATE CRETACEOUS)
LEUCOCRATIC MONZOGANITE (LATE CRETACEOUS)
BIOTITE MONZOGANITE (LATE CRETACEOUS)
MONZOGANITE GNEISS (JURASSIC(?))
PORPHYRITIC MONZOGANITE AND GRANODIORITE GNEISS (JURASSIC(?))
DARK GRANODIORITE GNEISS (JURASSIC)
DIORITE; ad = altered diorite (JURASSIC(?))
GNEISS (JURASSIC AND JURASSIC(?)) — Consists of the undivided monzogranite gneiss (Jmgn), porphyritic monzogranite and granodiorite gneiss (Jpgn), and dark granodiorite gneiss (Jggn)
QUARTZOFELDSPATHIC, CALC-SILICATE, AND MICACEOUS GNEISS AND SCHIST (JURASSIC(?) AND TRIASSIC(?))
GNEISS (JURASSIC, JURASSIC(?), AND PROTEROZOIC) — Consists of the undivided monzogranite gneiss (Jmgn).

WINDBLOWN SAND (QUATERNARY)
TWO-MICA GRANITE (LATE CRETACEOUS)
LEUCOCRATIC MONZOGRAHITE (LATE CRETACEOUS)
BIOTITE MONZOGRAHITE (LATE CRETACEOUS)
MONZOGRAHITE GNEISS (JURASSIC(?))
PORPHYRITIC MONZOGRAHITE AND GRANODIORITE GNEISS (JURASSIC(?))
DARK GRANODIORITE GNEISS (JURASSIC)
DIORITE; ad = altered diorite (JURASSIC(?))
GNEISS (JURASSIC AND JURASSIC(?)) — Consists of the undivided monzogranite gneiss (Jmgn), porphyritic monzogranite and granodiorite gneiss (Jpgn), and dark granodiorite gneiss (Jggn)
QUARTZOZOFELDSPATHIC, CALC-SILICATE, AND MICACEOUS GNEISS AND SCHIST (JURASSIC(?) AND TRIASSIC(?))
GNEISS (JURASSIC, JURASSIC(?), AND PROTEROZOIC) — Consists of the undivided monzogranite gneiss (Jmgn), porphyritic monzogranite and granodiorite gneiss (Jpgn), dark granodiorite gneiss (Jggn), and augen gneiss (Pagn)
SILICEOUS MARBLE AND CALCAREOUS QUARTZITE (TRIASSIC(?))
GARNET-MICA SCHIST (TRIASSIC (?))
MIGMATITIC GNEISS (MESOZOIC) — Consists of the undivided monzogranite gneiss (Jmgn), porphyritic monzogranite and granodiorite gneiss (Jpgn), and dark granodiorite gneiss (Jggn), as well as layered granitoid gneiss, amphibolite, and micaceous schist of uncertain age
METAMORPHOSED KAIBAB LIMESTONE (PERMIAN)
METAMORPHOSED COCONINO SANDSTONE (PERMIAN)

SILICEOUS MARBLE AND CALCAREOUS QUARTZITE
(TRIASSIC(?))

GARNET-MICA SCHIST.(TRIASSIC (?))

MIGMATITIC GNEISS (MESOZOIC) — Consists of the undivided monzogranite gneiss (Jmgn), porphyritic monzogranite and granodiorite gneiss (Jpgn), and dark granodiorite gneiss (Jgggn), as well as layered granitoid gneiss, amphibolite, and micaceous schist of uncertain age

METAMORPHOSED KAIBAB LIMESTONE (PERMIAN)

METAMORPHOSED COCONINO SANDSTONE (PERMIAN)

METAMORPHOSED HERMIT SHALE (PERMIAN)

METAMORPHOSED KAIBAB LIMESTONE, COCONINO SANDSTONE,
AND HERMIT SHALE, UNDIVIDED (PERMIAN)

METAMORPHOSED BIRD SPRING FORMATION (PENNSYLVANIAN
AND PERMIAN)

METAMORPHOSED REDWALL LIMESTONE (MISSISSIPPIAN)

DOLOMITE MARBLE (CAMBRIAN AND (OR) DEVONIAN)

METAMORPHOSED BONANZA KING FORMATION (LOWER PART)
(CAMBRIAN)

METAMORPHOSED BRIGHT ANGEL SHALE (CAMBRIAN)

METAMORPHOSED TAPEATS SANDSTONE (CAMBRIAN) —
Divided into:
White quartzite - equivalent to the Zabriskie Quartzite of
the Death Valley region

Feldspathic, cross-bedded quartzite - equivalent to the Wood Canyon
Formation of the Death Valley region

PINK MONZOGRANITE GNEISS (PROTEROZOIC)

AUGEN GNEISS (PROTEROZOIC)

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U·M·I

CORRELATION OF MAP UNITS

RELATION OF MAP UNITS

} JURASSIC

} JURASSIC (?)

} JURASSIC (?)
AND
TRIASSIC (?)

} JURASSIC (?)
AND
PROTEROZOIC

} TRIASSIC (?)

} MESOZOIC

} PERMIAN

} Lower Permian
and Pennsylvanian

} PERMIAN
AND
PENNSYLVANIAN

} MISSISSIPPIAN

} CAMBRIAN
AND (OR)
DEVONIAN

} PALEOZOIC

JPgn

pu

Tgms

M/gn

PK

Pc

Ph

PPb

Mr

DCm

Ebk

Eba

Etwq

Etfq

Egn

Egn

Pkch

} CARBONIFEROUS AND PROTEROZOIC

} TRIASSIC

} PERMIAN

} Lower Permian and Pennsylvanian

} PERMIAN AND PENNSYLVANIAN

} MISSISSIPPIAN

} CAMBRIAN AND ORDOVICIAN

} CAMBRIAN

} TRIASSIC (?)

} PERMIAN

} PERMIAN
AND
PENNSYLVANIAN

} MISSISSIPPIAN

} CAMBRIAN
AND (OR)
DEVONIAN

} CAMBRIAN

} Lower Permian
and Pennsylvanian

PALEOZOIC

PROTEROZOIC

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A

elevation (ft) 1800

scale 1 : 4975

horizontal scale = vertical scale

0.5
miles

scale 1 : 4975

horizontal scale = vertical scale

0.5
miles

0 0.5 1
kilometers

PLATE 4

A'

see PLATE 2 for explanation of symbols

A'

see PLATE 2 for explanation of symbols