I. GEOLOGY OF THE WILDROSE AREA, PANAMINT RANGE, CALIFORNIA

II. GEOCHRONOLOGIC STUDIES IN THE DEATH VALLEY-MOJAVE DESERT REGION, CALIFORNIA

Thesis by

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ABSTRACT

The bedrock in the Wildrose area is predominantly a sequence of metamorphosed sedimentary and possible volcanic rocks that is more than 15,000 feet thick. All the metamorphic rocks are Precambrian in age on stratigraphic evidence. The rocks are divided into three age groups, early Precambrian, later Precambrian, and Precambrian (?), that are separated by unconformities. The early Precambrian Panamint metamorphic complex has been divided into two series that are separated by a profound unconformity. Small stocks and dikes of granitic rocks, which are presumably Cretaceous in age, intrude the metamorphic rocks. Tertiary sedimentary rocks and Quaternary alluvium are also present in the area.

The structural features of the area are ascribed to four periods of deformation. The large folds in the early Precambrian rocks and associated minor folds were produced during the first two periods of deformation of early Precambrian age. Low angle faults are evidence of the third deformation which is considered to be Cretaceous in age. The fourth period of deformation produced high angle faults. Movement along the high angle faults began in late Tertiary time and has continued to the present.

All the Precambrian rocks of the area have been affected by middle or lower middle grade metamorphism. Even though structural evidence suggests that metamorphism must have accompanied the earlier periods of deformation, the present mineral assemblages reflect the third period of deformation which, on the basis of mineral age measurements, was Cretaceous in age. Narrow contact metamorphic aureoles surround the larger masses of granite.

Radioactive ages were measured in five areas in the Death Valley-Mojave Desert region. The rocks are Precambrian in age on stratigraphic evidence in three areas, the Wildrose area of the Panamint Range, the Mountain Pass district, and the Marble Mountains. The rocks in Joshua Tree National Monument and the Kilbeck Hills are presumably Precambrian, but no stratigraphic evidence of their age has been found. The geochronologic studies in the five areas indicate three distinct groups of K-Ar and Rb-Sr ages. Ages of approximately 1650 million years are obtained on metamorphic rocks and associated pegmatites in the Mountain Pass district. Younger igneous rocks at Mountain Pass and in the Marble Mountains have age patterns that are interpreted to indicate intrusion in the 1350 to 1450 million year interval. Metamorphic rocks in the Wildrose area, Joshua Tree National Monument, and the Kilbeck Hills have ages in the 75 to 85 million year interval which are interpreted as the age of regional metamorphism related to the Cretaceous Nevadan orogeny. An age of 73 million years was obtained on a post-metamorphic granite.

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Frontispiece - The central Panamint Range looking east. Wildrose Canyon on the left and Tuber Canyon on the right. Telescope Peak (11, 049 feet) on the right skyline. The eastern wall of the Wildrose graben forms the middle foreground. The dark areas are underlain by amphibolite and the light areas by micaceous quartzite of the lower metamorphic series of the Panamint complex.

PART I: GEOLOGY OF THE WILDROSE AREA, PANAMINT RANGE, CALIFORNIA

INTRODUCTION

Object of the investigation

The primary objective of this investigation is to understand and to possibly correlate the early Precambrian rocks in a part of southern California. Most workers have characterized this complex part of the section as "undivided igneous and metamorphic rocks." Lithologic correlation in such terranes is fraught with uncertainties, especially in a region like southern California where the much younger deposits of basins and valleys separate the metamorphic rocks that crop out in the adjoining mountain ranges. However, two methods of study might be combined to yield a basis for satisfactory correlation between certain areas: (1) careful delineation of the stratigraphic sequence in a representative section of Precambrian rocks, using classical stratigraphic methods, and (2) measurement of the absolute ages of minerals separated from these metamorphic rocks. Correlations between lithologically similar terranes may be made by comparing rock types or sequences of rock types. However, if a sequence of geologic "events" can be established by measuring absolute ages in a "standard section," it may be possible to correlate sections of different lithology which show the same sequence of mineral ages.

The Wildrose area in the central Panamint Range was selected for a test study because it is underlain by a very thick and diverse section of known Precambrian age. In addition it was hoped that this section has not been affected by the Mesozoic orogeny associated with intrusion of the Sierra Nevada batholith to the west, so that the absolute ages determined for rocks of the Wildrose area would be Precambrian and perhaps would reflect more than one event in Precambrian history. These latter objectives of the study were not entirely realized, as will be discussed farther on. However, the general geology and the stratigraphic relationships of the Wildrose area were studied in detail, and the results of this work are presented first.

General features of the area

During the period December 1958 to May 1960, approximately 60 days was spent in mapping and studying the geology of the Wildrose area. As herein defined this area is bounded approximately by parallels 36°12'30" and 36°17'30" north latitude and meridians 117°04'00" and 117°14'00" west longitude. It lies about 165 miles northeast of Los Angeles (Figure 1), in the central Panamint Range, San Bernardino County, California. The area comprises 52 square miles, of which about one-half lies within the Telescope Peak quadrangle and the remainder within the Emigrant Canyon quadrangle. Approximately 85 percent of the ground lies within the Death Valley National Monument.

Primary access to the area is by county highway from Trona, California, about 40 miles southwest of Wildrose Station, in lower Wildrose Canyon. This highway, which crosses the northwestern part of the map area, provides the main access to the Death Valley region from Los Angeles.

There are no towns in the area, but permanent residences are maintained at two places. Wildrose Station is

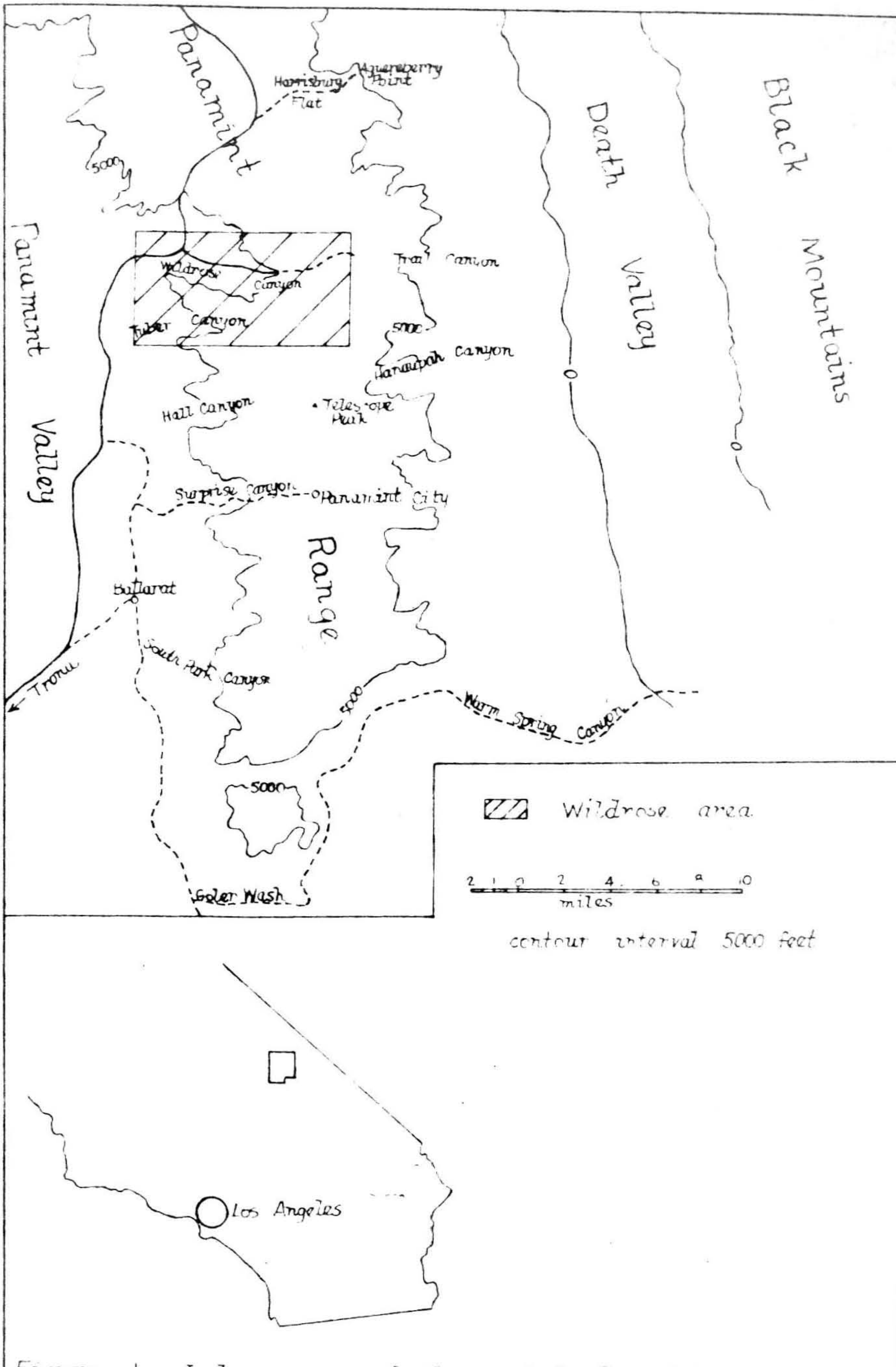

Figure 1. Index map of the central Panamint Range

operated as a concession from the National Park Service and is open year-round. A park ranger is stationed in Wildrose Canyon at the former summer headquarters of Death Valley National Monument.

The central Panamint Range typically has a steep western face gashed by deep, narrow canyons that commonly have springs in their upper portions. Maximum local relief in the Wildrose area is about 4,000 feet. The minimum elevation is approximately 2,600 feet in the southwest corner of the map area (see Plate 1), and the maximum elevations are along the east side of the area at Wildrose Peak (9,064 feet), Rogers Peak (9,994 feet), and Bennett Peak (9,980 feet). East-trending ridges and narrow, steep-walled canyons dominate the southern half of the area. The northern half has wider alluvium-filled valleys, such as Wildrose and A Canyons. These broad, nearly enclosed basins are considered to be remnants of an older erosion surface of low or moderate relief (Maxson, 1950). Harrisburg Flat (several miles north of the Wildrose area), Wildrose Canyon, and Middle and South Parks in the southern Panamint Range are related to this erosion surface, the present elevation of which has resulted from faulting of basin-range type along the west front of the range.

Vegetation in the Wildrose area is sparse but varied, and three floral zones, evidently controlled by elevation, can be recognized. Below an altitude of approximately 6,400 feet small brush is dominant. Mesquite, desert holly, Titastrophia, creosote bush, and cacti (chiefly prickly pear), are the most common plant types. Willow, wild rose, and cottonwood are concentrated around springs in Wildrose and Tuber Canyons. Above 6,400 feet and below 8,200 feet is a zone of piñon pine and juniper. Limber pine is the dominant conifer above an altitude of 8,200 feet.

Methods of investigation

Field data were plotted on a base map compiled from portions of the U. S. Geological Survey Telescope Peak (1952) and Emigrant Canyon (1952) 15-minute quadrangles (1:62,500) that were enlarged to a scale of 1:12,000. The names of some geographic features were adopted from D. E. White (1940). Some of the geology was mapped directly on the topographic base, but most was mapped on 1:48,000 scale aerial photographs (1948) enlarged to a scale of 1:12,000. Information was transferred to the topographic base map by using a light table and Saltzman projector.

Approximately 75 thin sections of rocks from the Wildrose area were studied in detail. Estimated modal analyses of 40 thin sections and modal analyses of 4 thin sections are presented in this report. An estimated mode is made by estimating the percentages of colored, birefringent, high relief, and opaque minerals and then subtracting to obtain the percentages of quartz and/or feldspar. In an estimated mode the abundance of a mineral is probably within 25 percent of the abundance as determined by point counting. Because the objective of the petrography was to determine the range of composition of the metamorphic rocks the method of estimating modes was more satisfactory than that of point counting. Sample locations are referred to a rectangular coordinate system with the origin at the southwest corner of the map area, near the mouth of Tuber Canyon. For example, the coordinates 37,600;12,800 indicate that the sample locality is 37,600 feet east and 12,800 feet north of the origin.

Previous Investigations

The mineral deposits and other geologic features of the Panamint Range are briefly mentioned in several reports dating back to 1874 (Stetefeld, 1874; Fairbanks, 1894, 1896a, 1896b; Spurr, 1903; Ball, 1907; Gale, 1914; Noble, 1926; Davis, 1927). The first relatively detailed treatments of the geology were presented by F. M. Murphy (1930, 1932); his primary interest was the mining district around Panamint City, but he did make a reconnaissance study of the stratigraphy of an area that corresponds approximately to the present 15-minute Telescope Peak quadrangle. D. E. White (1940) mapped two small parts of the Wildrose area in connection with the World War II strategic minerals investigations program of the U. S. Geological Survey. He studied two antimony deposits on the north and south sides of Wildrose Canyon, but used only descriptive rock names and made no detailed stratigraphic investigation in connection with this work. D. H. Sears later mapped much of the central Panamint Range in reconnaissance, and the results of his work are shown on the Death Valley sheet (1958) of the new Geologic Map of California.

Regional geologic setting

The Panamint Range, which is in the southwestern part of the basin and range physiographic province (Fenneman, 1931, p. 327), is about 100 miles long and 15 to 20 miles wide. It trends approximately N. 25° W. and is a typical asymmetrical basin range, with a very steep western slope toward Panamint Valley and a more moderate eastern slope toward Death Valley. Precambrian metamorphic rocks are dominant within the range, but sedimentary rocks of every Paleozoic system, Triassic sedimentary and volcanic rocks, Cretaceous intrusive rocks, and Tertiary sedimentary and volcanic rocks also are present. In general the Paleozoic rocks trend slightly more northwesterly than the trend of the range itself; thus the Precambrian-Paleozoic boundary trends obliquely across the range in such a way that rocks of Precambrian age are not exposed northwest of Townes Pass, the physiographic boundary between the northern and central parts of the Panamint Range. In the central part of the range, rocks of Paleozoic age crop out mainly on the eastern slope; on the average, these rocks dip gently to moderately eastward.

In a gross way the Panamint Range resembles a relatively simple tilted block, but a study of the eastern slope by C. B. Hunt (oral communications, 1960) indicates that its structural history is very complex. The Panamint Valley fault zone (Noble, 1926, p. 425), which bounds the range on the west side, is marked by numerous physiographic features that indicate very recent movement. The Death Valley fault zone, which lies to the east of the range, is concealed by alluvium and hence its exact location is uncertain. No recent movement is evident along that part of the zone directly east of the Panamint Range.

DESCRIPTION OF THE ROCK UNITS

General statement

The bedrock in the Wildrose area consists predominantly of metamorphosed sedimentary and possible volcanic rocks, mostly micaceous quartzite, amphibolite, schist, marble, and conglomeratic metagraywacke, but also including conglomerate, quartzite, argillite, limestone, and dolomite. These rocks are Precambrian and Precambrian (?) in age. They have been intruded by dikes, sills, and small stocks of Cretaceous granite or quartz monzonite. Younger sandstone and fanglomerate, probably upper Tertiary in age, are restricted in their occurrence to the western part of the area. Surficial deposits consist of various types of alluvial gravel, including some landslide material.

All the pre-Cretaceous rocks have been affected by regional metamorphism. There is no apparent sequence of increasing metamorphic grade within the Wildrose area; instead over the entire area the rocks show the effects of lower middle to middle grade metamorphism. The micaceous rocks contain chlorite, biotite, and very rarely garnet and andalusite as indicator minerals, whereas hornblende is the indicator of metamorphic grade in the amphibolites. Tremolite and diopside are present in the carbonate rocks. Superimposed on the regional metamorphic mineral assemblages are narrow contact metamorphic aureoles around the granitic stocks. In the largest of these aureoles, diopside is developed in amphibolite and garnet in schist.

Earlier designations of stratigraphic units in the central and northern Panamint Range stemmed mainly from the work of

Murphy (1930, 1932) in the Telescope Peak quadrangle and of Johnson (1957) in the Manly Peak quadrangle immediately to the south. To a large extent Murphy utilized local names for rock units, because his work preceded most of the investigations through which the presently accepted stratigraphic nomenclature of the Death Valley region was established. Firm correlations with other areas now make it possible to abandon most of Murphy's local names in favor of the well-defined stratigraphic section in general use east of Death Valley (Hazzard, 1937). These correlations, which are summarized in Table 1, are discussed individually in conjunction with descriptions of the units observed in the Wildrose area.

The stratigraphic section presented herein (Table 1) includes established names wherever possible. The thick section of rocks previously referred to as the Panamint metamorphic complex (Murphy, 1932), was divided by the present writer into lithologic mapping units in order to delineate the large structural features of the area. These units are not given formal names, as their extent beyond the limits of the Wildrose area is not known. The same rock types commonly are present in several of these units, a factor that makes correlation difficult in some parts of the Wildrose area.

No fossils were found in rocks within the Wildrose area. The Noonday dolomite and the Johnnie formation are tentatively assigned to the Precambrian; because these formations lie beneath rocks which contain a lower Cambrian fauna they are assigned a Precambrian (?) age. In Hanaupah Canyon,

Table 1 - Correlation chart of Precambrian formations of the Death Valley region

Hopper (1947) Harrisburg-Aguereberry area, Panamint Range	This Report	Murphy (1932) Panamint City area	Johnson (1957) Manly Peak quadrangle	Hazzard (1937) Nopah Range
Wood Canyon fm.			Wood Canyon fm.	Wood Canyon fm.
Sterling quartzite			Sterling quartzite	Sterling quartzite ? ?
Johnnie fm.	Johnnie fm.: upper mbr.	Death Valley fm.	Johnnie fm. (upper mbr.)	Johnnie fm.
	Johnnie fm.: lower mbr.	Hanaupah fm.	Johnnie fm. (lower mbr.)	
Precambrian rocks: buff dolomitic lst. with interbedded schist	Noonday dolomite	Telescope group: Redlands dolomitic lst.; Radcliff fm.; Sentinel dolomite	Noonday dolomite	Noonday dolomite
Precambrian rocks: quartz-biotite schist, slate, phyllite, dolomitic lst., quartzite, metamorphosed congl.	Kingston Peak fm.: South Park mbr.	Telescope group: Wildrose fm.; Mountain Girl congl.-quartzite; Middle Park fm.	Kingston Peak fm.: South Park mbr.	Algonkian
	Kingston Peak fm.: Sour Dough lst. mbr. ? ? ?	Telescope group: Sour Dough lst.	Kingston Peak fm.: Sour Dough lst.	
	Kingston Peak fm.: Surprise mbr.	Surprise fm.	Kingston Peak fm.: Surprise mbr.	
Precambrian rocks: quartz-biotite schist, crystalline lst., diabase bodies	Panamint metamorphic complex, upper series: unit C: dolomitic marble	Marvel dolomitic lst.	Archean gneiss	Archean granite gneiss
	Panamint metamorphic complex, upper series: unit B: garnet-chlorite-muscovite-quartz schist	Panamint metamorphic complex		
	Panamint metamorphic complex, upper series: unit A: biotite-quartz schist, calcite marble, muscovite-quartz schist, amphibolite			
	Panamint metamorphic complex, lower series: micaceous quartzite with interbedded amphibolite; diorite; marble			

about 5 miles east-southeast of Bennett Peak, which is approximately the southeast corner of the Wildrose area, a lower Cambrian fauna is present in the upper part of the Wood Canyon formation (C. B. Hunt, oral communication, 1960), and at Aguereberry Point, about 6 miles north-northeast of Wildrose Peak, lower Cambrian fossils are present in the upper part of the Wood Canyon formation immediately above the Zabriskie quartzite member (Hopper, 1947, p. 406). Lower Cambrian fossils also have been found in the upper part of the Wood Canyon formation, above the Zabriskie quartzite member, in the Nopah Range (Hazzard, 1937, pp. 310-312). The Noonday dolomite and the Johnnie formation are several thousand feet stratigraphically below the lowest horizon yielding lower Cambrian fossils. However, there is no apparent unconformity in the intervening Sterling quartzite and the major part of the Wood Canyon formation (see Table 1).

The Kingston Peak formation is the upper unit of the Pahrump group, which is usually assigned a later Precambrian age because the Noonday dolomite and overlying rocks are separated from it by an angular unconformity (Hazzard, 1937, p. 301; Noble and Wright, 1954, p. 144; Hewett, 1956, p. 29). Most rocks of the Pahrump group are very mildly metamorphosed, and they are not distinguishable in this respect from the overlying sections of early Paleozoic age. They rest with profound unconformity upon crystalline rocks that are highly metamorphosed. These crystalline rocks, which include the Panamint metamorphic complex, have been assigned an earlier Precambrian age (Noble and Wright, 1954, p. 143).

Panamint metamorphic complex

Name, subdivision, and distribution

The name Panamint metamorphic complex was introduced by Murphy (1932) to describe a group of metamorphic crystalline rocks that underlie a large area, including the Wildrose area, along the western slope of the central and southern Panamint Range. He suggested that the rocks are largely of sedimentary origin. The term Panamint metamorphic complex is used in the same sense by the present writer to include all rocks stratigraphically below the later Precambrian Kingston Peak formation, which is equivalent to all of Murphy's Surprise formation plus a part of his Telescope group. The Marvel dolomitic limestone of Murphy which lies below the Surprise formation is absent in the Wildrose area.

The Panamint complex is here divided into several mappable units on the basis of its occurrence in the Wildrose area. The most general subdivision of the complex is into two units, the lower and upper metamorphic series (Table 1). In addition, the lower series is divided into four lithologic units and the upper series is divided into three units. The lower series contains abundant amphibolite and rare thin lenses of marble whereas the upper series has prominent marble layers and only locally layers or lenses of amphibolite. The schists of the upper series can usually be distinguished from the micaceous quartzites of the lower series, but this distinction is subtle as the mineral composition of these two rock types is quite similar. The metasedimentary rocks of the Panamint complex contain trace amounts of heavy minerals, but it was not possible to distinguish between rocks of the two series on the basis of their heavy mineral suites.

Bedding as such was not recognized in the rocks of the Panamint complex. However, the layers or lenses of marble within the micaceous quartzite of the lower series, the compositional layering within marble units of the upper series, and the contact between marble and quartz-mica schist in the upper series are considered to be related to original bedding. As the foliation of the metamorphic rocks is subparallel or parallel to the contacts or layering, it is concluded that the foliation is probably generally parallel to bedding.

Rocks of the Panamint metamorphic complex form the major part of the bedrock exposure within the Wildrose area. In the vicinity of Tuber Canyon the metamorphic complex is overlapped on the west by Quaternary alluvium; the position of the contact approximately corresponds to the trace of the frontal fault that bounds the Panamint Range. From a point about half a mile south of Wildrose Canyon to the north side of the area, Tertiary sedimentary rocks, principally fanglomerates, lie on the metamorphic block and define its western margin of exposure. Along most of its eastern margin the Panamint metamorphic complex is overlain unconformably by the Kingston Peak formation, but near the north side of the area it is overlain in fault contact by the Noonday dolomite.

Lower metamorphic series

Distribution and subdivision

The lower metamorphic series of the Panamint complex is well exposed in the Wildrose area along the western slope of the Panamint Range, where it forms a north-trending band ranging in width from about half a mile in the area south of Tuber Canyon to more than 2 miles between Williams and Wildrose Canyons. North of Wildrose Canyon most of the lower series is covered by Tertiary fanglomerate. Another north-trending outcrop belt of the lower series occupies the central part of the map area. Along Tuber Ridge this belt is about 6,000 feet wide.

Detailed structure within the lower metamorphic series is probably very complex. The series has been profoundly deformed so that lenses or "blobs" rather than layers are typical. Distinctive marker horizons necessary to unravel the structure are lacking, but it is possible to delineate two contrasting rock types, micaceous quartzite and amphibolite, that together dominate the section. The quartzite weathers to various shades of brown and gray, or rarely red, whereas the amphibolite weathers very dark gray or nearly black. The color contrast between the two rock types imparts a grossly blotchy pattern to the range front that resembles the texture of marble cake (Frontispiece and Plate 1). The actual contact between micaceous quartzite and amphibolite was generally covered by rubble and not seen.

Micaceous quartzite forms the major part of the lower metamorphic series, but amphibolite also is abundant and has been mapped separately. Small bodies of diorite that crop out within the lower metamorphic series also are shown and described separately. Rare thin layers and lenses of marble are scattered through the micaceous quartzite, but most of these are too thin or not extensive enough to map as separate units.

Micaceous quartzite

The micaceous quartzite ranges from light gray to dark brown where fresh, and weathers to various shades of brown or gray. The rock is typically fine-grained with grains generally less than 0.5 millimeters in size but locally as large as 1 millimeter. Most outcrops show a poorly- to well-defined foliation which is the single planar element of the micaceous quartzite. This folitation is defined by planar concentrations of biotite, muscovite, and/or chlorite. Between foliation surfaces the rock is largely quartz. The foliation is parallel to the lenses or layers of marble that occur within the quartzite sections.

The micaceous quartzite has a considerable range in mineral composition. Quartz generally constitutes more than 50 percent of the rock or the sum of quartz plus plagioclase is more than 50 percent. Plagioclase is a major constituent where present, but was identified in only two of the ten thin sections that were studied in detail. The thin sections were not stained, so it is possible that untwinned plagioclase has been included with quartz in some thin sections. Estimated modes of the micaceous quartzite are given in Table 2, samples 1-10.

A fairly large percentage of plagioclase grains are untwinned and most grains are partly or largely altered to fine-grained white mica and clay minerals. On the basis of its occurrence in three thin sections a range of composition of plagioclase from calcic oligoclase to sodic andesine is indicated. The composition was determined utilizing the method of extinction angle of albite twins.

Quartz grains are rounded to sub-rounded and between 0.1

Table 2. Estimated modes of micaceous quartzite, lower series, Panamint metamorphic complex

<table>
<tr><th></th><th>1</th><th>2</th><th>3</th><th>4</th><th>5</th></tr>
<tr><td>quartz</td><td rowspan="2">50</td><td>60</td><td>50</td><td>45</td><td>50</td></tr>
<tr><td>plagioclase</td><td></td><td></td><td></td><td></td></tr>
<tr><td>muscovite</td><td>3</td><td>25</td><td>20*</td><td>30</td><td>1*
2</td></tr>
<tr><td>biotite</td><td>T</td><td></td><td></td><td></td><td><1</td></tr>
<tr><td>chlorite</td><td>40</td><td></td><td>15</td><td>15</td><td>40</td></tr>
<tr><td>graphite</td><td></td><td></td><td></td><td>5</td><td></td></tr>
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<tr><td>hematite</td><td>3</td><td>T</td><td>1</td></tr>
<tr><td>calcite</td><td>3</td><td>5</td><td>T</td><td>5</td><td>5</td></tr>
<tr><td>apatite</td><td>T</td><td>T</td><td>T</td><td>T</td><td></td></tr>
<tr><td>zircon</td><td>T</td><td></td><td>T</td><td></td><td>T</td></tr>
<tr><td>rutile</td><td></td><td></td><td></td><td></td><td>T</td></tr>
<tr><td>clinozoisite-epidote</td><td></td><td></td><td></td><td></td><td></td></tr>
<tr><td>leucoxene</td><td></td><td></td><td></td><td></td><td></td></tr>
<tr><td>cordierite (?)</td><td></td><td></td><td></td><td></td><td></td></tr>
<tr><td>dravite</td><td></td><td></td><td></td><td></td><td></td></tr>
<tr><td>map coordinates</td><td>10,800 E.
16,000 N.</td><td>7,500 E.
21,700 N.</td><td>6,030 E.
20,600 N.</td><td>7,300 E.
7,575 N.</td><td>4,150 E.
12,700 N.</td></tr>
</table>

*fine-grained white mica

Table 2 (Cont'd)

	6	7	8	9	10
quartz	57	35	56	29	37
plagioclase		30		25	T
muscovite	5		15		59
biotite	30	15	25	35	
chlorite		5		T	
graphite	1		2		
magnetite		T			2
hematite	5	15	1	1	2
calcite		T	T	T	
apatite			T	T	
zircon					
rutile					
clinozoisite-epidote	2		<1	10	
leucoxene			T		
cordierite (?)				T	
dravite					T
map coordinates	6,500 E. 10,600 N.	8,900 E. 14,850 N.	4,300 E. 6,560 N.	6,550 E. 4,400 N.	22,900 E. 14,500 N.

and 1.0 millimeters in size. Moderately- to highly-strained grains are the rule, and sutured grain boundaries are common.

Micas or chlorite are present in amounts between 20 and 45 percent. The colorless micas are differentiated on the basis of grain size; that is, the mica is called muscovite if it is coarse-grained or white mica if it is fine-grained. Some type of colorless mica, generally muscovite, is in eight of the ten thin sections. Prismatic, ragged grains of chlorite up to 0.5 millimeters long are in six of the ten thin sections. The chlorite is pleochroic from colorless to pale green. Both anomalous blue and brown interference colors are observed, but grains that show anomalous brown are more abundant. This indicates that in general the chlorites have an Fe/Fe+Mg value of slightly less than 0.52 (Albee, 1960, p. 1813). Biotite, present in six of the ten thin sections, ranges from less than 0.1 to 0.7 millimeters long but the average size is 0.3 millimeters.

In the majority of the thin sections only two varieties of mica and/or chlorite occur together; the assemblage biotite-chlorite is observed more frequently than either muscovite-chlorite or muscovite-biotite. Three of the minerals are found together in two of the ten thin sections. The biotite-chlorite assemblage is in four thin sections, and in two of these biotite and chlorite are intergrown. However, it is not possible to determine whether the minerals formed together or whether one formed at the expense of the other. The assemblage muscovite-chlorite is in three sections and the assemblage muscovite-biotite in two sections. Muscovite-chlorite-biotite is in two sections, but in these the biotite content is less than 1 percent.

Clinozoisite-epidote is a minor constituent in three thin sections of micaceous quartzite. The mineral occurs as discrete grains and does not appear to be an alteration product of plagioclase. Both anomalous first order blue and normal higher order interference colors are observed, which implies some substitution of Fe^{+3} for Al in the clinozoisite (Winchell and Winchell, 1951, p. 448).

Hematite, magnetite, graphite, and perhaps ilmenite are the opaque minerals. Graphite and magnetite are similar in appearance so that in some thin sections graphite may have been identified as magnetite. However, the writer has called any black opaque mineral with a "flaky" appearance in thin section graphite.

Accessory minerals identified in trace amounts include apatite, zircon, and rarely rutile. Calcite is present in most thin sections though only in very small amounts.

Amphibolite

Amphibolite makes up approximately 25 to 30 percent of the lower metamorphic series. It characteristically weathers very dark gray to black and produces a much darker outcrop than does the micaceous quartzite. This color contrast is very helpful in delineating areas of amphibolite on aerial photographs. Amphibolite exposures are slightly more resistant to weathering than those of micaceous quartzite, and commonly amphibolite produces a bouldery outcrop.

Lenses or layers of amphibolite range from a few feet in thickness to several thousand feet in outcrop width; but the largest outcrop width, on the ridge south of Wildrose Canyon, has almost certainly been thickened by folding. Probably individual layers range up to a few hundred feet in thickness.

In general the amphibolite is fine grained and has the same average size range as the micaceous quartzite, but some hornblende grains are a few millimeters in size. The parallel orientation of hornblende, and locally of chlorite and biotite grains, gives the amphibolite a crude foliation. However, this foliation is often difficult to ascertain, especially if the particular amphibolite layer is rich in quartz and/or plagioclase. Foliation is parallel to compositional layering as represented by contacts between amphibolite and micaceous quartzite. In some places a lineation is produced by parallel preferred orientation of hornblende grains.

Hornblende, quartz, and plagioclase are major minerals in all amphibolite specimens that were studied microscopically. Estimated modes of five samples of amphibolite are given in Table 3, samples 11-15. Biotite and chlorite are major minerals in some thin sections. Clinozoisite-epidote is present in most sections in minor amounts. Sphene is an abundant accessory mineral in all amphibolites that were studied. Magnetite-hematite, ilmenite, apatite, zircon, calcite, and white mica are trace minerals in some thin sections.

The hornblende, in general, occurs as subhedral to euhedral grains which range in size from 0.1 to 1 millimeters and, rarely, larger. Most grains have a "splotchy" appearance that results from poorly-defined zoning. That is, they have a core, which is

Table 3. Estimated modes of amphibolite and diorite, lower series, Panamint metamorphic complex

	11	12	13	14	15	16
hornblende	83	75	25	40	60	30
plagioclase	2(An_{29})	<1	5	35(An_{45})	34	45(An_{30})
quartz	5	16	25	7	5	10
potassium feldspar						5
clinozoisite-epidote	3	5	1	3	T	3
biotite	5 (biotite + chlorite)		35	10		
chlorite			5	2		5
sphene	2	1	$\frac{1}{2}$	$<\frac{1}{2}$	T	T
white mica			2			
magnetite-hematite		$\frac{1}{2}$	1	3	1	2
ilmenite		$<\frac{1}{2}$				
apatite				T	T	T
zircon				T		T
calcite		1				T
map coordinates	6,400 E. 2,600 N.	3,900 E. 5,750 N.	10,950 E. 4,350 N.	9,900 E. 9,980 N.	700 E. 9,450 N.	4,460 E. 19,600 N.

11-15 amphibolite
16 diorite

pleochroic from nearly colorless to medium brown, rimmed by material which is pleochroic from very faint green to pale medium green. Extinction angle ($\gamma \wedge C$) ranges from 19^{o} to 22^{o}.

The plagioclase ranges in composition from sodic to calcic andesine as determined in three of the five thin sections. Albite twinning is abundant in some sections and rare in others. Generally plagioclase occurs as anhedral grains 0.1 to 0.4 millimeters in size; patches of white mica alteration are common in most grains.

Quartz occurs as anhedral grains which average about 0.2 millimeters in size. Most grains show slight strain as evidenced by moderately undulant extinction. Sutured grain boundaries and mosaic texture are rare.

Sphene is a characteristic accessory mineral in the amphibolites. Small anhedral, "tear-shaped" grains, generally less than 0.1 millimeters in size, are typical; euhedral grains are rarely observed.

The opaque minerals are magnetite-hematite and ilmenite. In one thin section ilmenite occurs as cores of sphene grains. Intergrown magnetite and hematite are not uncommon and occur as discrete grains.

Diorite

Two small masses of diorite (or metadiorite) form part of the Panamint complex. Diorite is used here as a compositional term and is not meant to imply that the diorite originally crystallized as intrusive masses. Although it might have been an intrusive rock no evidence was observed that would prove its origin. The masses could be classified as amphibolite, but the diorite has certain megascopic differences which allow a separate map unit to be defined. The color index of the diorite is less than 40 whereas that of the amphibolite is generally more than 60. The foliation produced by preferred orientation of hornblende grains is much better in the amphibolite than in the diorite. Planar structures that resemble joints are common in the diorite but are not present in the amphibolite. In addition, differences in mineral composition were observed in thin section. In the diorite plagioclase is more abundant than hornblende and potassium feldspar is a minor constituent, but in the amphibolite hornblende is more abundant than plagioclase and potassium feldspar was not identified.

One mass of diorite crops out in lower Wildrose Canyon in the vicinity of Wildrose Station. It forms cliffs 25 to 75 feet high on both sides of the highway for a distance of approximately 1,000 feet. Another, smaller mass of diorite is exposed along the bottom of lower Tuber Canyon. The diorite in Wildrose Canyon is a dark gray, medium-grained, weakly foliated rock. Three thin sections of this rock were studied and an estimated mode for the least altered specimen is given in Table 3, sample 16.

Hornblende is the chief mafic mineral and is present as subhedral to euhedral grains up to 3 millimeters long. Hornblende is pleochroic from medium brown to dark green in the least altered specimen. Hornblende is altered in patches to chlorite that invariably shows anomalous blue interference colors. The hornblende in the diorite is coarser and generally more anhedral than hornblende in the amphibolite.

Plagioclase, the most abundant mineral in the diorite, occurs as anhedral grains up to 1 millimeter across. Although some grains are unaltered, others are nearly completely altered to fine-grained white mica and/or clay minerals. In the thin section for which the estimated mode is given plagioclase grains that show albite or pericline twinning are generally fresh or only slightly altered whereas untwinned grains are more strongly altered. The plagioclase feldspar is calcic oligoclase (An_{30}) on the basis of extinction angle of albite twins, but the abundance of untwinned grains suggests a composition toward median oligoclase. It also is possible that some of the untwinned feldspar may be albite formed as an alteration product of a mafic mineral. However, albite was not definitely recognized in the thin sections.

Potassium feldspar, in grains up to 1 millimeter in size, ranges in abundance from trace amounts to a few percent, but it is minor compared with plagioclase. Potassium feldspar grains are generally unaltered. Potassium feldspar was identified on the basis of optic sign and poorly developed gridiron twinning.

The diorite contains 10 percent or less quartz that occurs as anhedral grains 0.3 to 0.5 millimeters in size which have unsutured borders and show normal extinction. Although some or all of the quartz may have been produced as a metamorphic reaction product, there is no textural evidence to either prove or disprove this possibility.

Epidote is a minor mineral in the diorite. It occurs as small discrete grains, generally in the vicinity of large hornblende grains. Epidote almost certainly is not a primary mineral; it may have formed either as a deuteric mineral or during subsequent metamorphism.

Magnetite is the only opaque mineral in the diorite. Accessory minerals are apatite, sphene, and zircon, all present in trace amounts. Calcite, also a trace mineral, is probably secondary.

Contact relations between the diorite and micaceous quartzite of the lower metamorphic series are not obvious. The northern mass of diorite seems to occupy the core of a small anticline that plunges gently northward. The attitudes of foliation in the diorite and micaceous quartzite may be slightly discordant, but exposures near the contact are poor. However, the southern mass of diorite has micaceous quartzite on both sides, and the foliation in the rock types seems conformable. Therefore, whether the diorite is older than or perhaps intrusive into and younger than the micaceous quartzite is unknown.

Marble

Several discontinuous layers of calcite marble within the lower metamorphic series of the Panamint complex crop out north of Wildrose Canyon. They range from dark bluish gray to chocolate brown. Most of the marble contains a few percent muscovite as scattered flakes.

A tightly folded marble layer about 15 feet thick (Plate 2, fig. 1) crops out at Wildrose Station, and a tightly folded marble layer crops out in the bottom of Wildrose Canyon about 2,000 feet east of Wildrose Station. Resistant layers and pods of quartzitic and/or dolomitic material which weathers dark brown are common in the marble; and greenish, fine-grained quartzite that contains chlorite is interbedded with it.

All marble layers in the lower metamorphic series are shown with the same map symbol on Plate 1, but the stratigraphic relationship between individual marble layers is not known.

Upper metamorphic series

Unit A

The contact between the upper and lower metamorphic series of the Panamint complex is drawn at the base of the rather distinctive sequence of rock types that forms unit A of the upper series. This sequence comprises three contrasting rock types: biotite-quartz schist, calcite marble, and muscovite-quartz schist. Units of these rocks are somewhat continuous laterally and can be followed along strike across the entire Wildrose area. However, they pinch and swell, are folded and faulted, and are interbedded with one another. Consequently, the succession of rock types differs from place to place in the map area.

In most parts of the area biotite-quartz schist is present at the base of unit A. The schist is medium to dark brownish gray and weathers to a dark gray varnished outcrop. Although it is mineralogically similar to the darker-colored micaceous quartzites of the lower series, the schist of unit A characteristically has a slabby outcrop. These slabs break along the foliation, which is defined by fine- to medium-grained, locally coarse-grained, flakes of biotite. Estimated modes of biotite-quartz schist are given in figure 1 - Tightly folded marble layer in the lower metamorphic series of the Panamint complex, just east of Wildrose Station. The distance between dark colored layers on the limbs of the fold is 2 to 3 feet.

figure 2 - Overturned folds in unit B of the upper metamorphic series of the Panamint complex, looking north from Tuber Ridge. Distance between axial planes of the folds is about 500 feet. Trace of the axial plane of the Williams Canyon syncline lies just west (to the left) of the photograph.

figure 1

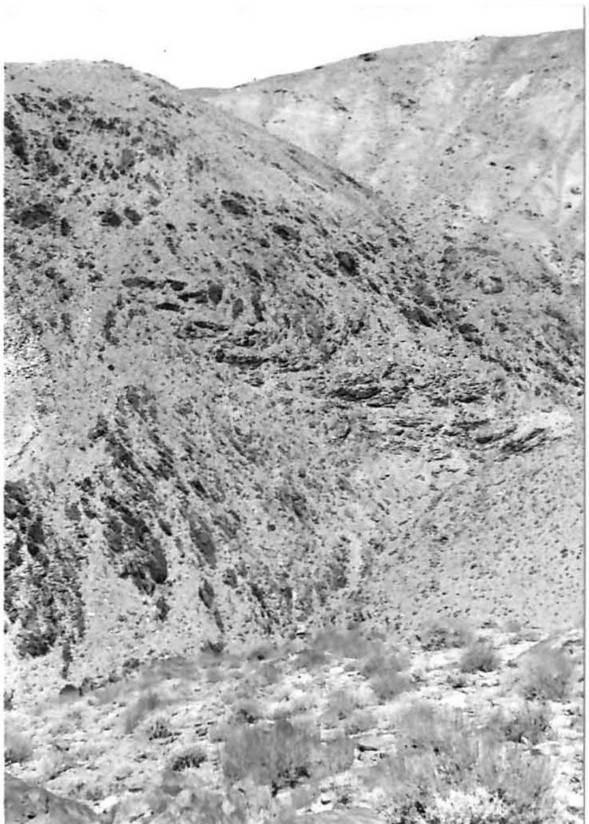


figure 2

Plate 2

Table 4. Plagioclase may locally be a major mineral in this unit (Table 4, sample 18); but on the basis of the average of the modes, the name biotite-quartz schist seems more appropriate.

Most commonly, the biotite-quartz schist is overlain by calcite marble that weathers dark bluish gray. Resistant layers, 1/8 to 1/4 inch thick, of medium- to coarse-grained quartz and minor phlogopite are typical of this marble, but the layers form less than 5 percent of the rock (Table 4, sample 20). The layers weather a rusty brown color and define the only apparent foliation in the marble.

Calcite marble crops out at various positions within unit A, but all marble in the unit is shown on Plate 1 with the same symbol. In certain parts of the area, between Wildrose and Williams Canyons and between Williams and Tuber Canyons, a thin layer of marble crops out between the biotite-quartz schist of unit A and the micaceous quartzites of the lower metamorphic series. On the ridge south of Wildrose Canyon this rock is a very dark gray, fine crystalline calcite marble, which contains tiny, scattered cubes of pyrite. Narrow layers, 1/16 to 1/4 inch thick, of very fine-crystalline calcite are tightly folded. One and one-half miles to the south a calcite marble of different character occupies the same stratigraphic position. This marble has a weathering color that ranges from light brown to dark gray. On a fresh surface the marble is white to light gray and coarsely crystalline with abundant, up to 5 percent, flakes of fine-grained graphite scattered through it.

Table 4. Estimated modes of unit A, upper series, Panamint metamorphic complex

	17	18	19	20	21
quartz	43	52	64	2	2
biotite	25	20	20		
muscovite	10	5	10		
white mica		10			
chlorite	10				
plagioclase		10(An_{31})			
albite	<1				
clinozoisite	10		<1		
magnetite	1	3	5		
hematite				$\frac{1}{2}$	<1
pyrite		T			
calcite			T	95	91
graphite				2	2
phlogopite				$\frac{1}{2}$	
diopside					5
apatite	T	T	T		
zircon	T	T	T		
rutile		T			
map coordinates	9,600 E. 13,000 N.	15,375 E. 16,500 N.	15,550 E. 13,750 N.	20,950 E. 23,400 N.	15,000 E. 18,350 N.

17-19 biotite-quartz schist
20-21 calcite marble

A calc-silicate marble (Table 4, sample 21) crops out directly south of the National Park Summer Headquarters. It lies above the base of unit A but below most of the biotite-quartz schist described previously. This marble is characterized by a layering that is defined primarily by differences in the size of calcite grains, which range in size from 0.1 millimeter in the fine-grained layers to 2 millimeters in the coarse-grained layers. The diopside, which makes up about 5 percent of the marble, is concentrated in the coarser layer.

The third major rock type in unit A of the upper metamorphic series, muscovite-quartz schist tends to be lenticular and rather restricted in lateral extent. At least two layers of muscovite-quartz schist have been mapped in unit A. The lower layer, which crops out on the ridge south of Wildrose Canyon in the axial part of the Williams Canyon syncline, is dark gray and graphitic. A good foliation is produced by flattened micaceous pods up to 2 millimeters thick and several millimeters in length. Most of these pods are fine-grained white mica, although coarser-grained muscovite is scattered through many of the pods. Estimated modes of two samples of the lower layer of muscovite-quartz schist are given in Table 5, samples 22 and 23.

The upper, less graphitic, layer of muscovite-quartz schist, whose normal stratigraphic position is at the top of unit A, becomes more prominent south of Wildrose Canyon. In most areas a marble lies beneath the upper layer, but this marble is lenticular so that in places muscovite-quartz schist is found directly above biotite-quartz schist. On the east limb of the

Williams Canyon syncline, from points less than 1 mile north of Tuber Canyon to the south side of the map area, muscovite-quartz schist is not present and unit B of the upper series is underlain by biotite-quartz schist except in a few places where discontinuous lenses of marble are found between unit B and biotite-quartz schist of unit A.

Unit B

Unit B of the upper series crops out in the Williams Canyon syncline from points 1 1/2 miles north of Tuber Canyon to the southern boundary of the map area. It forms a north-northwest trending belt up to 3,000 feet wide north of Tuber Canyon, but its stratigraphic thickness, which has been increased by folding, is much less. This unit produces bold, rocky outcrops, especially on the north side of Tuber Canyon. These outcrops are typically dark gray and very slabby, and they are commonly varnished and stained reddish brown. Slabs 18 inches on a side and 4 to 6 inches thick are common and slabs up to 3 feet on a side are not uncommon.

Unit B is rather distinctive and homogeneous in hand specimen; estimated modes are given in Table 5. North of Tuber Canyon (Table 5, sample 24) it is muscovite-chlorite-white mica-quartz schist with porphyroblasts of garnet. It weathers a dark brownish gray color and is heavily varnished and stained reddish brown. The characteristic slabby nature of the outcrop reflects the foliation produced by the micaceous minerals (Plate 3, fig. 1). Porphyroblasts of garnet, 1 to 3 millimeters in size and somewhat elliptical in shape, are partly chloritized at their ends. Their centers

Table 5. Estimated modes of units A and B, upper series, Panamint metamorphic complex

	22	23	24	25
quartz	42	58	37	57
muscovite	15	10	10	10
white mica	20	20	30	20
chlorite			15	
biotite			T	3
garnet			5	
plagioclase			1	
hematite				8
magnetite			2	2
graphite	20	10		
clay minerals	3	2		
pyrite		T		
zircon			T	0.1
tourmaline				T
map coordinates	15,000 E. 18,700 N.	14,150 E. 16,700 N.	14,250 E. 8,890 N.	17,650 E. 680 N.

22-23 quartz-muscovite schist, unit A
24-25 unit B

are fine-grained white mica. These presumed alteration minerals may indicate some retrograde effects. South of Tuber Canyon unit B crops out in a similar manner, but its mineralogy is slightly different (Table 5, sample 25). Garnet and chlorite are not present in sample 25, but white mica occurs as flattened elliptical pods which may indicate complete replacement of porphyroblasts of garnet or possibly of aluminum silicates. Little evidence was observed that indicates whether the garnet was formed during the second period of deformation and then altered during the third period of deformation or whether both formation and alteration of garnet occurred during the third deformation. (See section on geologic structure.) However, the foliation produced by the micaceous minerals wraps around the porphyroblasts which seems to imply that the garnet formed during the third period of deformation.

Unit C

Unit C of the upper metamorphic series, the uppermost unit in the Panamint complex, is a white dolomitic marble unlike any other carbonate unit in the Wildrose area. This marble is restricted in its outcrop to the axial portion of a syncline from points just north of Tuber Canyon to the southern border of the area. South of Tuber Canyon the outcrop width of unit C is nearly 5,000 feet but its stratigraphic thickness is probably less than 1,000 feet.

The marble is very coarsely crystalline with average grain size about 3 millimeters. It does not crop out as a resistant unit figure 1 - Unit B of the upper metamorphic series, Panamint metamorphic complex, between Williams and Tuber Canyons, looking south. The slabby outcrop is typical of unit B.

figure 2 - Clast in boulder conglomerate, South Park member of the Kingston Peak formation. On Tuber Ridge, coordinates: 34,400 E. and 5,800 N.

figure 1

figure 2

Plate 3

but instead forms a light gray rubble-covered slope. On a fresh surface the rock is white with rare graphitic (?) partings. The marble is massive and shows no bedding features; it does not contain quartzite or micaceous layers which are common in other marbles stratigraphically below and above this unit. This marble is also distinctive as it is dolomitic, whereas calcite marbles are in general more common in unit A of the upper metamorphic series and in the overlying Kingston Peak formation.

Unit C of the upper metamorphic series might seem to be the apparent stratigraphic equivalent of Murphy's Marvel dolomitic limestone (Table 1). The writer made a brief reconnaissance of the Panamint City area and observed that the Marvel dolomitic limestone apparently pinches out on the north side of Hall Canyon, which is about 3 miles south of the southern boundary of the Wildrose area. In the vicinity of Panamint City, the Marvel is a light-blue gray dolomite, which is distinctly different from the white dolomitic marble of unit C. The exact relationship between the Marvel dolomitic limestone and unit C of the upper metamorphic series is unknown, but they are not considered correlative.

Thickness

It is not possible to measure precisely the stratigraphic thickness of the Panamint complex, owing to the complicated structure and lack of distinctive marker horizons. However, a crude estimate of thickness can be reasoned out assuming the major structure of the complex to be synclinal. Using the ridge immediately north of Williams Canyon as a sample cross-section, the

Panamint metamorphic complex crops out in a belt approximately 23,000 feet wide. The upper series crops out in a zone 6,700 feet wide and the lower series in the remaining 14,300 feet. Dividing these figures by 2 to correct for the synclinal nature of the complex, the resulting "effective outcrop width" becomes 7,150 feet for the lower series and 3,350 feet for the upper series. Assuming that bedding in the complex is everywhere parallel to observed foliation with an average dip of 60°, the "effective thickness" of the lower series reduces to 6,200 feet and of the upper series to 2,900 feet. Finally, these values for "effective thicknesses" must be maximal, because within the Panamint complex, and especially within the lower series, there is widespread evidence for tight folding on a small scale.

Contact between lower series and upper series, Panamint metamorphic complex

The contact between the lower and upper series of the Panamint metamorphic complex is thought to be a profound unconformity; however, the typical physical characteristics of an unconformity have been obscured or destroyed by metamorphism. The marked difference in structural style between the two series is considered the strongest evidence for an unconformity. The structural style of the Panamint complex is described in more detail in the section on geologic structure. Unit A, the sequence of rock types which marks the base of the upper series, is folded and locally faulted, but it is continuous along strike for several miles and can be followed across the entire area. However, even though the lower series consists largely of two distinctive rock types, micaceous quartzite and amphibolite, individual units cannot be traced anywhere in a systematic manner. These observations lead to the conclusion that the lower series was profoundly deformed and consequently metamorphosed prior to deposition of the rocks which presently constitute the upper series.

That the two series are conformable and were deformed at the same time is an alternate hypothesis. This interpretation requires that the presumably more competent rocks of the lower series be severely deformed and sheared or boudined whereas the presumably incompetent marbles of the upper series maintain their stratigraphic integrity. However, at present the knowledge of behavior of materials under geologic conditions is limited, and the behavior of various rock types under deformation cannot be predicted.

The recognition of the contrasting styles of deformation in the two series is critical in defining the stratigraphy and structure of the Panamint complex. Because the lower series has apparently been more intensely deformed it is considered the older unit. The distribution of the lower series (Plate 1) requires that the major structure in the Panamint complex be synclinal, herein called the Williams Canyon syncline. The sequence of units in the upper series can be worked out assuming that their structure is synclinal.

Kingston Peak formation

Distribution and correlation

The sequence of primarily clastic rocks that overlies the Panamint metamorphic complex is assigned to the Kingston Peak formation. Its stratigraphic assignment is based on correlation with the southern Panamint Range where the formation lies above the Crystal Spring formation and below the Noonday dolomite (Johnson, 1957; Wright in Wasserburg, Wetherill and Wright, 1959).

The Kingston Peak formation is the upper unit of the Pahrump group, a sequence of three formations that is commonly present in the southern Death Valley region. Although the Pahrump group lies several thousand feet stratigraphically below the lowest occurrence of lower Cambrian fossils and is therefore Precambrian in age, the group is in other areas metamorphosed to the same mild degree as the overlying early Paleozoic rocks and not highly metamorphosed like the underlying crystalline rocks. The Kingston Peak formation, whose type section is in the Kingston Range, southeast of Death Valley (Hewett, 1940, pp. 239-240), has been mapped in the Wildrose area, but the other two members of the Pahrump group, the Crystal Spring formation and the Beck Spring dolomite, are absent. These latter formations are also absent in the Manly Peak quadrangle (Johnson, 1957, p. 360), but the Crystal Spring formation has been mapped in Warm Spring Canyon which lies east of the Manly Peak quadrangle (Wright in Wasserburg and others, 1959). In the Manly Peak quadrangle Johnson (1957, p. 360) divided the Kingston Peak formation into three members: the Surprise member, the Sour Dough limestone member and the South Park member. These members are recognizable in the Wildrose area, and the nomenclature proposed by Johnson is used in this report. The Kingston Peak formation, as herein defined, includes several units mapped by Murphy (1932): the

Surprise formation and the lower units of his Telescope group, including the Sour Dough limestone, the Middle Park formation, the Mountain Girl conglomerate-quartzite, and the Wildrose formation (see Table 1).

In the Kingston Range the Kingston Peak formation is almost entirely a clastic unit. Hewett (1940, pp. 239-240; 1956, p. 27) reports that the lowest and uppermost thirds of the formation are shaly sandstone with sporadic pebble zones whereas the middle third is coarse conglomerate. The middle third of the unit contains pebbles and cobbles of quartzite, granite, limestone, and dolomite in a minimum of sandy matrix; this part of the formation resembles the fanglomerate material presently being deposited in alluvial fans throughout the region. In the Saddle Peak hills some 30 miles west of the Kingston Range, Wright (1952) has described a complete section of the Pahrump series in which the Kingston Peak formation is broadly divisible into a lower part composed of fine-grained quartzite and shale and an upper part composed mostly of conglomeratic quartzite and conglomerate. In the southern Panamint Range, Johnson (1957, p. 360) describes the Kingston Peak formation as "...dominantly a conglomeratic subgraywacke or pebbly mudstone with subordinate shale, sandstone, conglomerate, and limestone."

The Kingston Peak formation in the Wildrose area contains metagraywacke, schist or phyllite, quartzite, conglomerate, and limestone. The formation crops out as a belt that trends north-northwest across the area. Foliation, which is parallel to bedding, strikes north-northwest to north-northeast and dips moderately to steeply eastward. Along the ridge south of Tuber Canyon this belt is 7,500 feet wide, which indicates a stratigraphic thickness of about 5,000 feet if one assumes an average dip of 45° and assumes no repetition of units. The most complete and structurally uncomplicated section of the Kingston Peak formation crops out along the ridge south of Tuber Canyon. However, the base of the Kingston Peak formation (Surprise member) is very difficult to locate along this ridge, as the lower part of the formation is lithologically similar to the underlying biotite-quartz schist of the Panamint metamorphic complex (upper series, unit A).

Surprise member

The name Surprise formation was introduced by Murphy (1930) to describe predominantly fine-grained flaggy or slaty rocks that occupy an irregular north-trending strip on the west side of the Panamint Range. The Surprise formation is named for exposures in Surprise Canyon in the southern Panamint Range. The Surprise formation nonconformably overlies the Marvel dolomitic limestone or, where the Marvel is missing, the Panamint metamorphic complex. A nonconformity also separates the Surprise formation from the overlying Telescope group (Murphy, 1932, p. 349). Johnson (1957, p. 363) redefined the formation as the Surprise member of the Kingston Peak formation. He assigned to the Surprise member all rocks above the "Archean" metamorphic rocks, which are continuous with the Panamint metamorphic complex (Johnson, 1957, p. 360), and below the Sour Dough limestone.

Within the Wildrose area the lower part of the Surprise member may properly be called metagraywacke. It is a dark gray, fine- to medium-grained biotite-muscovite-quartz-white mica rock with a poor to fair foliation that is spaced at intervals of $\frac{1}{2}$ inch to 2 inches. A specimen of metagraywacke from the Surprise member (Table 6, sample 26) has larger grains of biotite, muscovite, chlorite, and quartz set in a "matrix" of white mica and quartz. Presumably, metamorphism of a graywacke would produce an increase in grain size and the coarser-grained rocks of the Surprise member, herein called metagraywacke, could be the resulting product. That the micas in the metagraywacke are not detrital is indicated by their parallel preferred orientation which produces the foliation of the rock. Comparison of estimated modes of metagraywacke from the lower Surprise member (Table 6, sample 26) and biotite-quartz schist from the upper metamorphic series of the Panamint complex (Table 4, samples 17-19) shows that these two rock types are mineralogically similar.

The upper part of the Surprise member is mainly conglomeratic metagraywacke. Conglomeratic lenses occur throughout the member, but these lenses are rare in the lower part. In this report conglomeratic metagraywacke is defined as a rock with a metagraywacke matrix that contains less than 25 percent pebble- to cobble-size clasts. A rock with more than 25 percent pebbles and/or cobbles is called conglomerate. The matrix of most conglomerate consists predominantly of micaceous minerals; such rocks are called argillaceous conglomerate. In the Surprise member the conglomeratic portions are generally poorly sorted and have 10 to 15 percent clasts in a metagraywacke matrix.

Table 6. Estimated modes of the Kingston Peak formation

	26	27	28	29
quartz	39	25	45	40
biotite	15	25	30	30
muscovite	10	30	7	4
chlorite	1	<1	7	
graphite (?)	3			
magnetite				1
hematite	2	3	1	
white mica	30			
clinozoisite-epidote		15		
sphene		1		
tourmaline	T			
zircon	T			T
calcite			T	
"knots"			10*	25**
map coordinates	29,125 E. 3,950 N.	31,400 E. 11,450 N.	23,500 E. 9,025 N.	34,000 E. 8,250 N.

*ovoid clots of white mica that contain larger grains of quartz, muscovite, biotite, and calcite

**ovoid clots of white mica that contain scattered grains of quartz and magnetite and rare grains of muscovite and biotite

26-28 Surprise member
29 South Park member

The clasts range in size from $\frac{1}{4}$ inch to 2 inches. Light to dark gray quartzite clasts are the most abundant type, but clasts of granite or granitic gneiss are not uncommon. Most of the clasts are relatively well rounded but some are slightly flattened parallel to the poorly developed foliation of the metagraywacke matrix.

Although the metagraywacke of the Surprise member appears rather monotonous, some variation in composition does occur. Sample 27, Table 6, is representative of the upper part of the member. The only real difference between sample 27 and sample 26 from the lower part of the member is the presence of clinozoisite-epidote and the absence of white mica in the former. Sample 27 is apparently somewhat calcareous whereas sample 26 is not; it cannot be concluded, however, on the basis of two thin sections that the upper part of the Surprise member is somewhat calcareous whereas the lower part is not.

A limestone bed, not unlike the overlying Sour Dough limestone locally is exposed in the upper part of the Surprise member. The bed does not exceed 50 feet in thickness and apparently is lenticular; it crops out only in upper Wildrose Canyon and hence it is not a good marker horizon. The limestone is medium crystalline and light to medium gray with rude color banding.

The upper part of the Surprise member includes two other lithologic varieties. One is medium to dark gray metagraywacke characterized by a very good planar cleavage. This cleavage is as thinly spaced as 1/8 inch and probably reflects a higher content of micaceous minerals than in other parts of the member. The second variety is dark gray "knotty" schist that contains $\frac{1}{4}$ inch "knots" in a phyllitic or schistose matrix of quartz and biotite. The knots are aggregates of biotite, muscovite and quartz grains that range from 0.3 millimeters to 2 millimeters in size. Probably the knots originally grew as porphyroblasts of some unknown mineral. The biotite in the matrix has a strong preferred orientation that bends around the knots, which also implies that the knots represent altered porphyroblasts. Sample 28, Table 8, is an estimated mode of a representative specimen of "knotty" schist. These two lithologic varieties are gradational into normal metagraywacke and conglomeratic metagraywacke.

The thickness of the Surprise member can be estimated reasonably well at two places in the area. If a dip of 70° is assumed the thickness of the member at the bottom of Tuber Canyon is approximately 3,800 feet. On the north side of Wildrose Canyon it is approximately 4,150 feet thick if a dip of 50° is assumed. For reasons stated previously the base of the member is difficult to locate precisely. Even so, a stratigraphic thickness of approximately 4,000 feet is indicated for the Surprise member of the Kingston Peak formation.

Sour Dough limestone member

The name Sour Dough limestone was proposed by Murphy (1932, p. 350) for a coarsely crystalline, micaceous, arenaceous gray limestone with alternating white and gray stripes. The Sour Dough limestone is the lowest formation in Murphy's Telescope group. The formation is named after Sour Dough Canyon, a tributary of Surprise Canyon near Panamint City. Johnson (1957, p. 365) redefined the Sour Dough limestone as the middle member of the Kingston Peak formation, and his nomenclature is used in this report.

The Sour Dough limestone is the only distinctive horizon marker within the Kingston Peak formation, and its presence makes it possible to divide the formation into three members. It forms a prominent bluish gray unit near the middle of the dark brown-colored series of rocks that make up the remainder of the Kingston Peak formation. Exposure of the limestone is in marked contrast to the rather subdued slope produced by the underlying Surprise member; the limestone crops out as a low resistant unit or locally forms a small cliff. It ranges in thickness from 40 feet on the ridge south of Tuber Canyon to 120 feet north of Wildrose Canyon.

The Sour Dough limestone member is a coarse-crystalline bluish-gray limestone. On a fresh surface it is irregularly banded very light gray and medium to dark gray; on a weathered surface the color banding is light yellowish gray and medium to dark gray. This color banding probably reflects distribution and concentration of graphite. The color layering is roughly parallel to the upper and lower contacts of the limestone and thus reflects the attitude of bedding. Locally it is tightly folded. In addition to color layering there are quartz-rich layers which are more resistant and weather out as rusty-colored ridges. Just west of Piñon Mesa the Sour Dough limestone contains rare very small cubes of hematite after pyrite. North of Wildrose Canyon cubes up to 5 millimeters on a side of hematite after pyrite make up 2 to 3 percent of the limestone. At the latter locality coarse-grained muscovite is a minor constituent of the limestone, but in general the Sour Dough limestone member is not appreciably micaceous.

It is possible that an unconformity separates the Surprise and Sour Dough limestone members of the Kingston Peak formation. Murphy (1932, p. 349) suggested that a nonconformity separates the Surprise formation and the Sour Dough limestone, the lowest unit of his Telescope group (see Table 1). His primary evidence was the apparent difference in deformation of rocks above and below the nonconformity. In the Wildrose area there are no apparent differences in the degree of deformation of rocks within the Kingston Peak formation. However, though the two members seem conformable in most places, the Sour Dough limestone apparently truncates bedding or foliation in metagraywacke of the Surprise member on the ridge west of Piñon mesa. The abrupt change from metagraywacke to limestone may also suggest a hiatus; that is, a time interval of indefinite duration during which the conditions of sediment deposition changed. But, interbedded limestone and clastic units are relatively common in the geologic column and the present writer is not suggesting an intraformational unconformity on this evidence.

South Park member

The South Park member forms about the upper one-third to one-half of the Kingston Peak formation. The member was defined by Johnson (1957, p. 365) for a well-exposed section near

South Park basin in the northern part of the Manly Peak quadrangle where the member ranges in thickness from 285 feet in Goler Wash in the south part to 1,000 feet in South Park Canyon near the northern boundary. The South Park member is equivalent to three formations of the Telescope group of Murphy (1932, p. 350): the Middle Park formation, the Mountain Girl conglomerate-quartzite, and the Wildrose formation (see Table 1).

The stratigraphic section of the South Park member which the writer studied in greatest detail crops out along Tuber Ridge; much of the following description of the member is derived from this section. The South Park member produces a dark gray, subdued slope except for small cliffs of conglomerate and quartzite. Foliation, which parallels bedding in the member, in general dips moderately to the east. The South Park member overlies the Sour Dough limestone and in turn is overlain, probably with slight unconformity, by the Noonday dolomite.

The lowest part of the South Park member, directly above the Sour Dough limestone, is conglomeratic metagraywacke which is similar to conglomeratic portions of the Surprise member. A layer or lens of "knotty" schist that is not continuous along strike and is gradational into conglomeratic metagraywacke both above and below, commonly crops out about 50 to 100 feet above the Sour Dough limestone. An estimated mode of this schist is given in Table 6, sample 29. It is seen by comparison with "knotty" schist from the Surprise member (Table 6, sample 28) that there is no striking difference in mineral composition. Likewise, in the hand specimen, there is no criterion which will enable one to distinguish between the members on the basis of the occurrence of this type of schist. Johnson (1957, p. 366) reports that in the Manly Peak quadrangle a "...spotted porphyroblastic rock that occur(s) only in the South Park member" along with some other lithologic types serve to distinguish the South Park member from the Surprise member. However, in the Wildrose area the "knotty" schist which is considered the same as Johnson's porphyroblastic rock, is common to both the Surprise and South Park members.

A few hundred feet above the top of the Sour Dough limestone is the base of argillaceous conglomerate, a distinctive lithologic type peculiar to the South Park member of the Kingston Peak formation. Locally, conglomerate lies directly above the Sour Dough limestone and it generally forms a series of small cliffs which are very dark gray in color and heavily varnished. The overall dark color of the conglomerate reflects the composition of the matrix which is biotite-rich; the clasts are predominantly very light gray or white quartzite that weather a light gray color in marked contrast to the dark matrix.

For the most part clasts in the conglomerate are pebble size but pockets of cobble conglomerate are not uncommon. Along Tuber Ridge near the top of the unit is a minor occurrence of boulder conglomerate (Plate 3, fig. 2). Rough pebble counts of conglomerate outcrops show that nearly all of the clasts are white to light gray quartzite. In thin section most clasts are a mosaic aggregate of quartz grains with rare grains of plagioclase, biotite, and muscovite. However, rare clasts of quartz-rich granitic and gneissic rocks are observed.

The matrix of the argillaceous conglomerate seems to be primarily quartz and biotite in the hand specimen; but in thin section muscovite and magnetite also are seen to be important constituents. Composition of the matrix is variable throughout the area so that the conglomerate is somewhat schistose where biotite along with muscovite is predominant or quartzitic where the percentage of micaceous minerals is low. Schistose conglomerate crops out along Tuber Ridge and in highway roadcuts north of the map area between A Canyon and Harrisburg Flat; quartzitic conglomerate is general on the north side of Wildrose Canyon.

A characteristic feature of the conglomerate is the texture produced by pronounced stretching or flattening of the clasts. The clasts are disc-shaped and flattened into the plane of the foliation. If a clast is cut parallel to the foliation the cross section is oval to nearly circular. In general the flattening ratio of clasts ranges from 2 1/2:3:1 or 5:6:1 or, rarely, larger; where b lies in the plane of the foliation perpendicular to its strike, a lies in the plane of the foliation perpendicular to b, and c is perpendicular to the ab plane.

Feldspathic quartzite, conglomeratic in part, and very light gray in color, is interbedded with and intertongues with argillaceous conglomerate. The quartzite contains a few percent of potassium feldspar and minor biotite. Detrital grains are plainly visible, especially on a weathered surface, and the quartzite as a whole is medium to coarse grained. Argillaceous conglomerate and feldspathic quartzite are mapped as separate types because they are distinctive lithologic varieties. In the South Park member conglomerate, quartzite, both, or neither may be present in a random stratigraphic section. The most striking features of the sequence of conglomerate and feldspathic quartzite are the lenticular nature of the rock types and their marked lateral variations in thickness.

Another type of conglomerate, limestone fragmental conglomerate, is prominent in the South Park member north of Wildrose Canyon. This limestone conglomerate has a dark brown-weathering matrix that is light to medium gray on a fresh surface. The matrix is slightly calcareous and is composed largely of medium sized quartzite and quartz grains and scattered flakes of biotite. Clasts of light gray to tan medium-grained limestone that weather light brown make up 10 to 20 percent of the rock. The limestone clasts are flattened, generally with a flattening ratio larger than that observed in the quartzite clasts in the stretched pebble conglomerate described previously. Although limestone clasts are the rule, quartzite clasts also occur but they are not abundant.

A nearly white bed of dolomite crops out in the upper part of the South Park member north of Wildrose Canyon. This dolomite is only a few feet thick and probably lenticular as it is not found south of Wildrose Canyon, except as a thin bed in the fault slice east of Antimony Canyon.

Dark-colored, relatively pure quartzite is abundant near the top of the South Park member. Most of this quartzite is uniform or massive, but locally, it contains a few percent of biotite and is thinly bedded. Some of these beds show very faint grading.

Noonday dolomite

The section that overlies the Kingston Peak formation in the Wildrose area is composed primarily of carbonate rocks that probably are equivalent to the Noonday dolomite, one of the more widespread and easily recognized formations in the southern Death Valley region. The type locality of the Noonday dolomite is in the Nopah Range, California, where the formation consists of about 1,500 feet of cream-colored algal dolomite, sandy in places (Hazzard, 1937, p. 300). It is the only thick series of carbonate rocks in the stratigraphic interval, which is several thousand feet thick, between the Beck Spring dolomite of the Pahrump series and limestone in the upper part of the Wood Canyon formation. Neither of these latter formations is present in the Wildrose area, but the Wood Canyon formation is present just east of the area. The Noonday dolomite is characteristically tan or buff in color and is approximately 1,000 feet thick in the Wildrose area. The formation is broadly divisible into three parts that are roughly equal in thickness, but members as such have not been differentiated. The lower third of the formation is arenaceous limestone with subsidiary calcilutite, the middle third is laminated micaceous limestone, and the upper third is dolomite, locally arenaceous. In the vicinity of Panamint City, Murphy (1932) divided this part of the stratigraphic column, herein called the Noonday dolomite, into three formations, the Sentinel dolomite at the bottom, the Radcliff formation, and the Redlands dolomitic limestone at the top (see Table 1). However, these formations are not distinctive units in the Wildrose area so Murphy's nomenclature is not used in this report. Instead, rough correlations with his section are suggested in the description of the Noonday dolomite in the Wildrose area which follows. In general the middle one-third of the Noonday dolomite retains the characteristic lithology which Murphy assigned to the Radcliff formation, but the upper and lower parts of the Noonday dolomite commonly bear little resemblance to the Sentinel dolomite and the Redlands dolomitic limestone.

In the Wildrose area there is a pronounced lithologic break between the dark-colored metagraywackes of the South Park member of the Kingston Peak formation and the overlying Noonday dolomite, which consists primarily of light-colored carbonate rocks. Continuous exposures of the Noonday dolomite are restricted to the eastern part of the map area; in upper Wildrose Canyon the base of the formation is located about half a mile west of the Charcoal Kilns. Outliers of the Noonday dolomite crop out on ridges north and south of Wildrose Canyon, west of the belt of continuous exposure.

In its type locality and other parts of the Death Valley region, the base of the Noonday dolomite is an angular unconformity (Hazzard, 1937; Noble and Wright, 1954). In the southern Panamint Range most exposures of the contact between the Noonday dolomite and the Kingston Peak formation indicate apparent conformity, but an unconformity has been suggested on the basis of possible truncation of the uppermost unit of the Kingston Peak formation (Johnson, 1954, p. 354). In the Wildrose area the Noonday dolomite and the South Park member of the Kingston Peak formation in places seem to be conformable, but north of Wildrose Canyon the truncation of units in the South Park member by the base of the Noonday dolomite suggests an unconformable relationship.

The lower part of the Noonday dolomite crops out as a prominent tan band, the estimated thicknesses of which are 125 feet on Tuber Ridge and 155 feet north of Wildrose Canyon. The most characteristic lithology in the lower part of the Noonday dolomite is tan- to light brown-weathering, coarse-grained arenaceous limestone that is cream to light tan on fresh surface. Rounded quartz grains, 1/64 to 1/8 inch in size, make up 30 to 40 percent of the rock, and on a weathered surface these more resistant quartz grains give the rock a granular appearance. Muscovite and magnetite are minor constituents of the arenaceous limestone. An estimated mode of a sample of arenaceous limestone is given in Table 7, sample 30. The arenaceous limestone forms the lower part of the Noonday dolomite except as noted below, and the contact with the underlying South Park member of the Kingston Peak formation is particularly well exposed on Tuber Ridge west-southwest of Hummingbird Spring. Light gray-weathering quartzite, light gray on a fresh surface, that is mostly coarse-grained quartz but also includes about 2 percent fine-grained biotite forms the base of a small cliff (Plate 4, fig. 1). This quartzite belongs to the South Park member of the Kingston Peak formation. The lower 6 to 7 feet of the Noonday dolomite, immediately above the quartzite, is light brown-weathering calcareous quartzite conglomerate. This conglomerate contains

Table 7. Estimated modes of the Noonday dolomite

	30	31		32	33		34
		arenaceous limestone	calcilutite		mica-rich layers	mica-poor layers	
calcite	64	56	71	74	54	93	
dolomite							96
quartz	35	15	15				4
muscovite	$\frac{1}{2}$	10		10	25	3	T
biotite		10	10	15	20	3	
chlorite		5					
epidote		3	3				
pyrite				1			
magnetite	$\frac{1}{2}$	<1	<1				
hematite					1	1	T
map coordinates:	23,850 E. 14,350 N.	36,050 E. 7,200 N.		38,950 E. 10,150 N.	37,450 E. 9,400 N.		46,300 E. 15,300 N.

figure 1 - Contact between the Noonday dolomite, above, and the South Park member of the Kingston Peak formation. On Tuber Ridge, coordinates: 35,300 E. and 5,200 N.

figure 2 - Mixed rock (Kmr), lower Tuber Canyon. Amphibolite of the lower metamorphic series, Panamint complex cut by dikes of Cretaceous granite. Height of face is about 75 feet.

figure 1

figure 2

Plate 4

about 60 percent white to light gray, sub-angular to well-rounded quartzite clasts, which range from 1/8 inch to 1/2 inch in size, in a matrix of cream-colored, medium-crystalline limestone. Trace amounts of fine-grained biotite and muscovite are scattered through the matrix. The contact between the conglomerate and the underlying quartzite is very sharp and generally quite flat, but there is local relief on the contact of 1 to 1 1/2 feet. The conglomerate grades upward into brown-weathering arenaceous limestone as described previously.

The base of the Noonday dolomite in other parts of the area is commonly marked by the appearance of arenaceous limestone of an essentially different type. This limestone consists of interbedded arenaceous limestone and calcilutite. The interbeds range in thickness from 1/32 inch to 1/2 inch, and the alternate arenaceous limestone layers weather out in relief so that the rock has a corrugated or laminated appearance. The arenaceous limestone layers weather medium brown or gray and the layers are light gray on a fresh surface. Medium-grained quartz, calcite, and biotite in decreasing order of abundance compose the layers. Some layers contain a few percent coarse-grained muscovite which shows up especially well on a weathered surface. The calcilutite layers weather a greenish gray and are slightly darker gray on a fresh surface than are the limestone layers. The calcilutite layers are predominantly very fine- to fine-grained calcite with minor biotite, An estimated mode of this interbedded arenaceous limestone and calcilutite (Table 7, sample

31) shows that, although the mineralogy of the arenaceous limestone is similar to that of the calcilutite, the relative amounts of individual minerals vary between the two lithologic varieties. Muscovite is restricted to the arenaceous limestone layers; but calcite, quartz, magnetite, biotite, chlorite, and epidote are common to both arenaceous limestone and calcilutite. The two textural types interfinger with the coarser-grained arenaceous limestone layers seemingly more continuous; lenses of arenaceous limestone are found within calcilutite layers. It seems likely that there is lateral variation in the lithologic character of the interbedded arenaceous limestone-calcilutite component of the Noonday dolomite.

The upper contact of the interbedded arenaceous limestone-calcilutite sequence is sharp. That is, even though arenaceous limestone interbeds are mineralogically similar to the tan arenaceous limestone that forms the major portion of the lower part of the Noonday dolomite, there is no obvious gradation between the two types of limestone that might be produced by merely lensing out the calcilutite interbeds.

Quartzite interbeds occur within the tan arenaceous limestone. These quartzite beds are medium to dark brown and weather a dark brown, rusty color. The quartzite is calcareous; individual beds range from a few inches to a few feet in thickness. A fine color lamination and a thinly spaced cleavage characterize the quartzite; this cleavage is parallel to bedding and is probably related to accumulations of mica.

A relatively thin layer of white dolomite often crops out in the lower part of the Noonday dolomite. This dolomite is probably the Sentinel dolomite as defined by Murphy in the vicinity of Panamint City (1932, p. 349). The white dolomite has not been mapped as a separate unit, but the writer noted that the position of the white dolomite within the formation seems to vary throughout the area. Whether this means that there is more than a single white dolomite layer or that the tan arenaceous limestone is variable in thickness laterally is not known.

The characteristic lithologic type in the middle part of the Noonday dolomite is similar to the Radcliff formation in the vicinity of Panamint City (Murphy, 1932, p. 349), but the Radcliff formation is not mapped as a separate unit in this report. In the Wildrose area the middle part of the formation is a continuous rocky belt and small cliffs are general. These cliffs are discontinuous along strike and are evidently not controlled by stratigraphy alone. A typical section of the "Radcliff lithology" crops out north of Wildrose Canyon on the west flank of Wildrose Peak; at this locality the middle part of the Noonday dolomite is about 400 feet thick. The "Radcliff lithology" is laminated micaceous limestone whose lamination is compositionally controlled and is produced by alternating mica-rich and mica-poor layers. On a weathered surface the mica-rich layers are rusty brown and are slightly more resistant than the light gray to tan mica-poor layers. On a fresh surface the mica-poor layers are light gray and the mica-rich layers are medium to dark gray. The layers are very uniform and continuous and range from 1/32 inch to 3/8 inch in thickness, with the mica-poor layers everywhere thicker than the mica-rich layers. Estimated modes for two thin sections of laminated micaceous limestone are given in Table 7, samples 32 and 33. Quartz is also a constituent of the resistant layers, but it was not identified in the two thin sections. The laminated micaceous limestone averages about 25 percent mica, muscovite plus biotite, and 75 percent calcite with minor pyrite and hematite. The pyrite occurs as irregular pods, which are elongate parallel to the layering; the pods are a few millimeters or less in length. On the north side of Wildrose Canyon about half a mile north of the Charcoal Kilns, tiny cubes, about 1/16 inch on a side, of hematite after pyrite are common along the boundary between mica-rich and mica-poor layers.

The "Radcliff lithology" of the middle part of the Noonday dolomite is spectacularly deformed (Plate 5, figs. 1 and 2); the more resistant, rusty-weathering layers are complexly folded and ruptured. An axial-plane foliation is locally developed in the intervening mica-poor layers; this foliation is defined by paper-thin dark-colored layers of graphite.

The most characteristic lithology in the upper part of the Noonday dolomite is similar to the Redlands dolomitic limestone described by Murphy (1932, p. 349). But, as rock types other than the white crystalline dolomitic limestone of Murphy are abundant in the upper part of the Noonday dolomite a separate unit is not mapped in the Wildrose area. The upper part of the formation crops out as a prominent light gray to nearly white layer on the flanks of Wildrose, Bennett, and Rogers Peaks. This layer is easily distinguished from the medium brown or gray of the underlying middle part of the Noonday dolomite and the dark brown of figure 1 -Typical deformation in the "Radcliff lithology", Noonday dolomite, coordinates: 38,600 E. and 9,100 N.

figure 2 - Small, tight folds in the "Radcliff lithology", Noonday dolomite, coordinates: 39,700 E. and 7,700 N. Note hammer in center of photograph for scale.

figure 1

figure 2

Plate 5

the overlying Johnnie formation. The base of this white layer crosses Wildrose Canyon a few hundred feet west of the Charcoal Kilns.

White to light gray fine-crystalline dolomite is the major rock type in the upper part of the Noonday dolomite. The bedding is very indistinct and the formation produces a massive rocky outcrop. In places the dolomite is slightly arenaceous; quartz ranging in size from less than 1/16 inch to 1/4 inch is scattered through it. An estimated mode of a thin section of this gray dolomite is given in Table 7, sample 34. Quartz locally occurs as irregular dark blue pods up to a few inches long. These pods are now aggregates of highly strained quartz grains up to 2 or 3 millimeters across, but the pods may represent original chert nodules. A narrow white zone that surrounds these quartz pods is, in thin section, about 1 millimeter wide and is fine-grained possibly altered material that contains remnants of what seem to be grains of diopside.

A snow-white dolomite that contains rosettes of very coarse tremolite grains is restricted to the upper part of the Noonday dolomite. Individual grains up to 1 1/2 inches long are common in the Wildrose area, but tremolite crystals several inches long can be seen in the Redlands dolomitic limestone in the vicinity of Panamint City. These tremolite rosettes are concentrated along foliation planes, or at least are easily seen on the surfaces of slabs that weather out along the foliation. The foliation is considered parallel to bedding.

A stratigraphic section of the Noonday dolomite that is not complicated by folding or faulting is difficult to find in the Wildrose area. However, a section of the formation that is well exposed along the ridge which runs southwest from Wildrose Peak is described in Table 8. The thicknesses of units in the section are not measured precisely but are estimated from measurements of elevation.

The stratigraphic section of Table 8 is not typical of the entire Noonday dolomite and hence it presents some ambiguities. Units 1 and 2 of the Noonday dolomite represent what the writer has called the lower part of the formation. On Tuber Ridge tan arenaceous limestone lithologically similar to unit 2 forms the lowest part of the formation and overlies quartzite of the Kingston Peak formation. Thus the relationship on the north side of Wildrose Canyon implies marked lateral variation in the composition of the lower part of the Noonday dolomite. Unit 1 of the formation is probably correlative with Murphy's Sentinel dolomite and unit 3, the laminated micaceous limestone that is the most consistent lithologic type in the Noonday dolomite, is correlative with his Radcliff formation. Unit 4 represents the upper part of the formation though the characteristic white tremolitic dolomite that is common south of Wildrose Canyon, is absent.

Johnnie formation

The Johnnie formation was defined by Nolan (1929) in the Spring Mountains region, Clark County, Nevada. Nolan describes the lower 3,500 feet of the formation as largely fine-grained quartzites with a few shale zones, and the upper 1,000 feet as

Table 8. Stratigraphic section of the Noonday dolomite

The section was measured in a northeasterly direction along the ridge southwest of Wildrose Peak between elevations of approximately 7,500 and 8,050 feet. The thickness of units has been estimated from measurements of elevation.

Johnnie formation (lower 20 feet only)

medium gray, fine-grained quartzite that weathers reddish brown. A fine lamination at 1/2 to 1 inch intervals probably represents bedding. Platy fragments break along surfaces parallel to the lamination.

Noonday dolomite (total estimated thickness 640 feet)

unit no.	thickness (feet)	
4	75	limey dolomite, light and bluish gray color-banded.
3	400	laminated micaceous limestone. The resistant mica-rich layers are bluish gray and weather rusty brown. The mica-poor layers are very light gray and weather medium gray. Individual layers range from 1/32 to 1/4 inch thick.
2	115	tan arenaceous limestone that weathers light brown and has poor to massive bedding. On a weathered surface, resistant medium to coarse quartz grains give the rock a granular appearance.
1	40	white, very fine-crystalline dolomite that typically has small irregular granules and pods of quartz which weather out in relief.

South Park member, Kingston Peak formation (upper 15 feet only)

limestone fragmental conglomerate that weathers dark brown. The matrix is dark gray and consists of medium-grained quartz and minor biotite. Clasts make up 10 to 20 percent of the rock; the clasts are mostly tan to brown, fine- to medium-grained limestone that weathers light brown. The clasts range in length from 1/4 inch to 12 inches and have a length:thickness ratio of about 4:6:1. About 10 percent of the clasts are light bluish gray limey dolomite; less than 5 percent of the clasts are quartzite.

quartzite with a larger percentage of shale and a few beds of dolomite. In the Spring Mountains the base of the Johnnie formation is not exposed and its upper contact is a thrust fault. Hazzard (1937), in the Nopah Range some 30 miles west of the Spring Mountains, measured a section of the Johnnie formation 2,550 feet thick. The lower part of the formation is mainly interbedded quartzite and sandy dolomite and is gradational from the carbonate rocks of the underlying Noonday dolomite to the predominantly clastic rocks, sandstone, shale, quartzite, and dolomite, of the upper part of the Johnnie formation. In the southern part of the Panamint Range Johnson (1957, p. 373) assigns a section of rocks 2,000 to 2,500 feet thick to the Johnnie formation; the lower member of the formation consists of quartzite, sandy dolomite, and dolomite, and the upper member consists of shale, siltstone, sandstone, and minor dolomite. There is some uncertainty in the precise correlation of the Johnnie formation in the Wildrose area with the formation in the southern Panamint Range. Johnson reports that the contact between the Noonday dolomite and the Johnnie formation is gradational; he defines the base of the Johnnie formation at the lowest occurrence of thin quartzite beds that interrupt the sequence of carbonate beds that compose the Noonday dolomite (1957, p. 373). In the Wildrose area the contact between clastic rocks of the Johnnie formation and carbonate rocks of the Noonday dolomite is abrupt. The Johnnie formation has been divided into two members but they are not very similar in lithology to those described by Johnson. Correlation of the rocks herein called the Johnnie formation with the standard section of the Death Valley-Mojave Desert region is based on its position between the Noonday dolomite and the fossiliferous Wood Canyon formation.

The lower part of the Johnnie formation, in addition to the entire Noonday dolomite, has been reported as absent in the central part of the Panamint Range (Hopper, 1947, pp. 403-405). In the Harrisburg-Aguereberry area the Johnnie formation has been described as shale, slate, and phyllite, with subordinate limestone and dolomite (Hopper, 1947, p. 403). These rocks are very similar to and have been correlated with the upper member of the Johnnie formation in the Wildrose area. In addition the writer suggests that the clastic rocks in the upper part of the "pre-Cambrian" section of the Harrisburg-Aguereberry area (Hopper, 1947, p. 402) are probably correlative with the lower member of the Johnnie formation.

The sequence of clastic rocks that overlies the carbonate rocks of the Noonday dolomite along the eastern border of the Wildrose area is assigned to the Johnnie formation. Murphy (1932, p. 349) named these rocks the Hanaupah formation in the vicinity of Panamint City, but the standard Death Valley-Mojave Desert nomenclature (Hazzard, 1937) is used in this report.

The Johnnie formation forms a belt along the east side of the Wildrose area but only the lower part of the formation lies in the map area. Its contact with the overlying Sterling quartzite and the Wood Canyon formation of Hazzard (1937), lies east of the range crest. The Johnnie formation crops out in a belt some

2 1/2 miles wide south of Tuber Canyon and at least 6,000 feet wide near the north side of the area. In general the formation dips gently to moderately eastward although local structural modifications do occur.

The Johnnie formation produces a dark-colored slope in marked contrast to the very light-colored slope that underlies the upper part of the Noonday dolomite. The Johnnie formation can be divided into two members that are different in lithology; this difference is expressed, in particular south of Wildrose Canyon, by contrasts in color, type of outcrop, and characteristic vegetation. The lower member weathers a dark brown color and it produces a rather massive, rubbly float whereas the upper member weathers a greenish gray color and it produces a very slabby or slaty float. The color difference is best seen from a distance; the type of float is commonly the best criterion for distinguishing the members at short range. Outcrop of the Johnnie formation is generally only fair, and the slopes which underlie the formation are covered by talus or rubble. A modest cover of piñon pine and juniper is general on the lower member at elevations above about 6,400 feet, but the upper member is commonly barren of vegetation.

The relationship between the Noonday dolomite and the members of the Johnnie formation is best shown due east of the buildings located at the southeast edge of Piñon Mesa. At this place the lower member of the Johnnie formation seems to overlie conformably white tremolitic dolomite of the Noonday dolomite. The lower member is about 500 feet thick and consists predominantly of chlorite-white mica-quartz rock. The rock has no well-developed planar features, but breaks along surfaces that apparently are parallel to bedding. Chlorite occurs as stubby porphyroblasts up to 6 millimeters long in a matrix of fine-grained quartz and muscovite and very fine-grained white mica. Estimated modes of two specimens of the porphyroblastic chlorite rock are given in Table 9, samples 35 and 36.

The lower member of the Johnnie formation thickens to the south of Tuber Canyon. North of Wildrose Canyon outcrop of the Johnnie formation is poor; the contact between the two members cannot be located precisely though argillite of the upper member of the formation crops out on Wildrose Peak. The contact between members as shown on Plate 1 is inferred by keeping the thickness of the lower member approximately equal to its thickness south of Wildrose Canyon.

Along the ridge south of Tuber Canyon the lower member of the Johnnie formation consists of porphyroblastic andalusite-biotite-muscovite-quartz rock. This rock weathers dark brown, but on a fresh surface it is light gray and is characterized by bluish gray porphyroblasts of andalusite in a matrix of medium-grained biotite, muscovite, and quartz. The porphyroblasts of andalusite are stubby rectangular prisms that range in length from 10 millimeters to 50 millimeters; they have square cross sections that are commonly 10 to 15 millimeters on a side. Although micaceous minerals are quite abundant the random distribution of andalusite porphyroblasts tends to destroy any schistosity in the rock. In thin section the constituent minerals fall into three size

Table 9. Estimated modes of the Johnnie formation

	35	36	37	38 "sand"	38 "silt"
quartz	42	10	35	72	50
plagioclase (?)					
muscovite	5	2	2.5	1	½
white mica	21	59	19	10	40
biotite			15	15	3
chlorite	30	25	1.5		3
andalusite			26		
hematite	2	3			
magnetite		1	1	2	3½
zircon				T	
map coordinates	38,040 E. 6,400 N.	41,620 E. 9,400 N.	39,500 E. 0 N.		43,400 E. 2,800 N.

categories. In addition to the porphyroblasts, which are largely altered to white mica, the "groundmass" of the rock contains grains of biotite, chlorite, muscovite, and andalusite from 0.7 to 1.0 millimeters in size in a fine-grained matrix of quartz, white mica, and magnetite grains that are less than 0.1 millimeter in size. The andalusite typically occurs as anhedral, highly poikilitic grains that are slightly pleochroic and commonly twinned. The large porphyroblasts of andalusite poikilitically enclose biotite, muscovite, magnetite, and "groundmass" andalusite. The distribution of minerals enclosed within the large andalusite porphyroblasts is the same as that in the "groundmass" which implies that the porphyroblasts grew in a somewhat homogeneous biotite-muscovite-andalusite-magnetite-quartz rock. The chlorite is a minor mineral that occurs as an alteration product of biotite. An estimated mode of a thin section that contains about 30 percent andalusite porphyroblasts is given in Table 9, sample 37. The outcrop area of the porphyroblastic andalusite rock is restricted to the ridge south of Tuber Canyon and to the upper plate of the low-angle fault on Tuber Ridge.

Along the ridge south of Tuber Canyon bedding can be recognized near the base of the lower unit of the Johnnie formation. The porphyroblasts of andalusite are smaller, 15 millimeters or less in length, and the groundmass shows bedding that is defined by color lamination. Mineralogically this distinct lamination consists of alternate quartz-rich and biotite-rich layers that are less than 0.1 millimeter thick. Slabs of this rock, 2 to 3 inches thick, break along a surface that is parallel to the bedding.

The upper member of the Johnnie formation may in general be described as a sequence of argillites. A typical section of the upper member crops out on the ridge west of Bennett Peak on the south side of Tuber Canyon. The lower part of the member characteristically is greenish gray in color and breaks into slabby fragments 1/2 to 2 inches thick. Surfaces of breakage are parallel to bedding that is defined by light and dark gray laminae 1/16 to 1/4 inch thick. Well-developed rosettes of chlorite, about 1/8 inch across on the average, are abundant, particularly on bedding surfaces. In thin section the argillite is composed largely of fine-grained quartz, white mica, and magnetite but also includes larger porphyroblasts of chlorite, biotite, and magnetite. Bedding of the argillite is defined by quartz-rich and white mica-rich layers. The chlorite is arranged in radiating clusters or rosettes with individual grains 0.3 to 1.5 millimeters in length. It shows an abnormal brown interference color that indicates Fe/Fe+Mg is less than 0.52 (Albee, 1960, p. 1813). Individual chlorite grains are commonly arranged transverse to the quartz-mica layering as are smaller grains of greenish biotite up to 0.5 millimeters in length. Magnetite is present as anhedral grains 0.2 to 0.5 millimeters across and also as very fine-grained opaque material less than 0.02 millimeters across.

The chlorite gradually decreases in abundance and the color lamination becomes more prominent going upward in the upper member. As chlorite rosettes decrease in abundance, markings which might be direction current features can be observed on the bedding surfaces. Most of these markings are linear "rill marks" but round to oval "molds" or "casts" also occur. Some of the latter markings are up to 1 inch in length.

Above a position approximately 550 feet above its base the upper member is variegated argillite. The colors of laminae include dark gray, purple, and reddish brown. These laminae range from 1/32 to 3/8 inch in thickness and many are lenticular or "wispy". Some laminae, the reddish brown ones in particular, have a fine grained sandy appearance. Epidote is abundant and is found along bedding surfaces and along fractures transverse to bedding. The physical features of the argillite suggest that it was deposited as a siltstone and the presence of epidote implies a calcareous component. An estimated mode of a specimen of upper Johnnie argillite is given in Table 9, sample 38. This specimen shows all of the previously described features except that epidote is absent.

On the ridge south of Tuber Canyon a single thin bed of dolomite crops out about 50 feet above the base of the upper member. The dolomite is light gray and it weathers to a brown color. The more resistant layers are probably quartzose. A few percent of mica flakes are scattered through the rock. The dolomite is not present north of Tuber Canyon.

Tertiary rocks

The Tertiary rocks include the Nova formation, which is probably Plio-Pleistocene in age, and older Tertiary strata of unknown age. The Nova formation has been designated as possibly

Miocene in age (Hopper, 1947), but it resembles the Funeral fanglomerate of the Death Valley area and may be correlative with it (Wright and Troxel, 1954).

Near the mouth of Wildrose Canyon unnamed older Tertiary strata (Wright and Troxel, 1954, pp. 23-24) unconformably overlie micaceous quartzite and amphibolite of the lower series of the Panamint metamorphic complex. The older Tertiary strata crop out both north and south of lower Wildrose Canyon, in Nemo Canyon slightly more than 1 mile from its intersection with Wildrose Canyon, and in a small canyon just south of Williams Canyon. A stratigraphic section was not measured, but these older Tertiary strata are probably not more than about 200 feet thick.

These rocks are chiefly sandstone and mudstone and include a thin basal member of fresh-water limestone (Wright and Troxel, 1954, p. 24). The limestone is light gray and weathers light to medium gray. It is fine to medium crystalline and contains less than 1 percent pyrite. The limestone member is 6 to 8 feet thick and its contact with the underlying metamorphic rocks is knife-edge sharp. Erosional remnants of limestone are mapped between Wildrose and Williams Canyons (Plate 1). These remnants are east of the major exposure of older Tertiary strata which implies that the former areal distribution of older Tertiary strata was greater. The basal limestone is not present everywhere; in the canyon south of Williams Canyon tuffaceous sandstone rests directly on micaceous quartzite. The larger part of the older Tertiary formation is yellowish to reddish brown sandstone and mudstone that are typically non-resistant and poorly sorted. They contain rock fragments up to small pebble size in a very fine-grained matrix. The matrix material and some of the fragments are tuffaceous. It seems likely that the older Tertiary strata consist largely of detrital material of volcanic origin.

The contact between older Tertiary tuffaceous sandstone and the overlying fanglomerate of the Nova formation is probably an unconformity, but the contact is generally covered by debris that weathers out of the fanglomerate and sloughs down over the tuffaceous sandstone.

The Nova formation was defined by Hopper (1947, p. 414) to describe a section in Nova Canyon, which is located some 8 miles northwest of Wildrose Canyon. In the Wildrose area the Nova formation is wholly fanglomerate, but to the north basalt flows, rhyolitic tuffs, and lenses of monolithologic breccia are intercalated with the fanglomerate. The fanglomerate consists of angular to subrounded clasts, ranging in size from pebbles to large boulders, in a fine-grained sandy matrix. The fanglomerate has a crude bedding defined by alternate coarser- and finer-grained lenses or layers. The approximate strike of the bedding is north-south and it dips 5° to 10° eastward. Clasts of all of the bedrock formations of the Wildrose area are represented in the fanglomerate, and as the distribution of the fanglomerate is adjacent to Panamint Valley this suggests that the material was derived from the east. Farther north, Hopper (1947, p. 414) concluded that the fanglomerate was derived from an eastern source area because its clasts consist exclusively of rock types from the central Panamint Range. The Nova formation dips gently to the east or southeast. In the Wildrose area it is several hundred feet thick, but farther north the Nova formation is at least

3,000 feet thick (Hopper, 1947, p. 414).

Quaternary deposits

Two types of Quaternary deposits are mapped in the Wildrose area. The category of alluvium includes stream gravels, fan gravels, and terrace gravels, whereas landslide material includes all debris that presumably has been deposited by landslides. The largest area of alluvium occupies 5 or 6 square miles in middle and upper Wildrose Canyon. Smaller areas of alluvium are mapped in A and Tuber Canyons and east of Nemo Canyon. The various types of gravel offer a bulk sample of the bedrock geology of the area. Clasts in the gravels range in size from granules to large boulders and are generally sub-rounded to rounded. The only real difference between the Quaternary gravels and the fanglomerate of the Tertiary Nova formation is the better induration of the latter.

Remnants of older alluvium are left as terraces in upper Wildrose Canyon and along much of Tuber Canyon. The local drainage baselevel is controlled by the elevation of the floor of Panamint Valley, and this baselevel is changed whenever there is movement along the Panamint Valley fault zone. Scarps indicate that the most recent movement along the fault zone has been predominantly vertical, with the east side relatively up. Consequently, dissection of the older alluvium has taken place as the drainage has adjusted to its new local baselevel. The depth of dissection approaches 100 feet at the mouth of Tuber Canyon; terrace remnants 50 to 75 feet high are found in upper Wildrose Canyon.

Quaternary landslide deposits are mapped at two places in upper Wildrose Canyon; one locality is on the north side of Wildrose

Canyon about half a mile west of the Charcoal Kilns and the other locality is southeast of Piñon Mesa. Both of these deposits contain unconsolidated, poorly sorted debris or rubble derived from the Johnnie formation that crops out upslope.

Igneous rocks

Granite

A few small stocks and numerous dikes and sills of granitic rock of Cretaceous age crop out in the Wildrose area. The largest mass of intrusive rock is located in the northeastern part of the area on the north side of Wildrose Canyon. Smaller granitic masses farther east on the north and south sides of Wildrose Canyon are probably extensions of the main stock, the intervening portions being covered by alluvium. Smaller stock-like masses crop out in lower Tuber Canyon and in upper Antimony Canyon. Numerous dikes and sills are scattered through the area and the largest ones are shown on Plate 1. The intrusive rock crops out as conspicuous white areas that are easily distinguished from the dark gray slopes underlain by rocks of the Panamint metamorphic complex and the Kingston Peak formation. The intrusive rocks are slightly more resistant to weathering than the metamorphic country rock and form rocky, somewhat raised exposures. The rocks are typically coarse grained, in places pegmatitic, and are non-foliated.

The intrusive rocks are mostly granite in composition though there is some variation. Quartz, potassium feldspar, plagioclase, muscovite, and commonly garnet are major minerals. Very coarse grain size, pegmatitic in places, characterizes the granite. The composition of the intrusive rocks is in the range of simple granite pegmatites. Modes are given in Table 10 for 3 samples taken from two stocks and 1 sample from a large dike.

Potassium feldspar is the most abundant mineral in all thin sections, except sample 41 from the Tuber Canyon stock. The potassium feldspar is white and very coarse grained; individual grains are as much as several millimeters across. In sample 39 the potassium feldspar is microcline microperthite, but in the other intrusive bodies the feldspar is not perthitic. Gridiron twinning is very common in sample 39 but untwinned grains are more abundant in the other samples.

Plagioclase on the average has the composition of calcic oligoclase. Grains of plagioclase are smaller than those of potassium feldspar; the maximum size of plagioclase grains is about 2 millimeters. Albite twinning is common and carlsbad twinning is not uncommon in the plagioclase, but it is unzoned. The degree of albite twinning varies between thin sections and the distinction between untwinned plagioclase, untwinned potassium feldspar, and quartz becomes a problem. However, plagioclase grains generally have a light alteration of fine-grained white mica and/or clay minerals whereas potassium feldspar grains are generally quite fresh. This alteration together with the fact that all indices of refraction of plagioclase are less than those of quartz serve to distinguish the two minerals.

Quartz occurs as anhedral grains that fill the interstices between feldspar grains. Quartz grains are as much as a few millimeters in size and they show normal to slightly undulant extinction.

Table 10. Modes of Cretaceous granite

	39		40	41
	coarser-grained phase	finer-grained phase		
quartz	18.2	21.5	39.7	44.5
potassium feldspar	51.6	50.4	45.1	10.7
albite*	10.6	1.5		
plagioclase	14.2	20.0	7.7	34.2 (An_{25})
muscovite	4.6	6.2	7.4	9.4
iron oxide	0.7		0.1	0.1
garnet				0.3
apatite				0.1
zircon				0.1
myrmekite	0.1	0.4		0.6
map coordinates	40,000 E. 17,150 N.		22,600 E. 24,750 N.	8,700 E. 2,480 N.

*exsolution blebs in perthite

Muscovite is present as large plates, commonly as much as several millimeters in length. In hand specimen books of muscovite up to 1/4 inch thick can be seen.

Garnet is an abundant accessory mineral in most of the granitic masses. It is common in the Wildrose Canyon stock from which sample 39 is taken, but the thin sections did not happen to cut across any grains of garnet. In hand specimen it occurs as euhedral crystals up to 1 millimeter across, but the average size of the crystals is less than 0.5 millimeters. The garnet is reddish in color and presumably is spessartitic in composition.

The granitic intrusive rocks in the Wildrose area are believed to be Cretaceous in age. These rocks are intrusive into the Panamint metamorphic complex and into the Kingston Peak formation so that, on a strict basis, the field relations in the area only allow an age assignment of post-later Precambrian. However, a K-Ar age of 73 million years has been measured on muscovite from granite in the Wildrose area (Part II of this report).

Mixed rock

Zones of so-called mixed rock are mapped around the stock in lower Tuber Canyon. The term mixed rock refers to those areas that contain country rock mixed with abundant granitic material in the form of an interwoven network of dikes and veinlets (Plate 4, fig. 2). Individual dikes are as much as several feet thick but all are too small to map separately. Therefore, the areas designated as mixed rock on Plate 1 are the zones in which dikes are abundant. Within such areas in lower Tuber Canyon micaceous quartzite and amphibolite of the Panamint metamorphic complex show pronounced contact metamorphic effects; a discussion of these effects is presented farther on.

The masses of intrusive rock in Wildrose Canyon are not surrounded by a zone of mixed rock like that found in the vicinity of the lower Tuber Canyon stock. In addition, no obvious contact metamorphic aureole surrounds the Wildrose Canyon intrusive bodies though some mineralogic changes are observed in thin section (see section on metamorphism).

Other intrusive rocks of the Panamint Range

Granitic rocks of Mesozoic age are described from parts of the Panamint Range both north and south of the Wildrose area. In the Ubehebe Peak quadrangle, about 25 miles northwest of the Wildrose area, J. F. McAllister (1956) has named and described in detail the Hunter Mountain quartz monzonite. A typical specimen of Hunter Mountain quartz monzonite contains 40 percent orthclase (including albite in perthite), 40 percent plagioclase (oligoclase (An_{22}) to albite), 14 percent quartz, 4 percent hornblende, nearly 1 percent each of magnetite and sphene, and traces of biotite, apatite, epidote, and sericite. Widely varied facies, including olivine-bearing and alkaline rocks, are present along the margins of the main quartz monzonite mass and in separate smaller masses. The Hunter Mountain quartz monzonite intrudes the Bird Spring (?) formation (McAllister, 1956) of Pennsylvanian and Permian age and lies nonconformably beneath continental deposits of late Pliocene or Pleistocene age.

In the Manly Peak quadrangle, about 20 miles south of the Wildrose area, Johnson (1957, pp. 388-393) describes Mesozoic intrusive rocks of two different ages. Granodiorite, regarded as Triassic in age (Johnson, 1957, p. 391) contains on the average plagioclase (oligoclase) 45 percent, potassium feldspar 15 percent, quartz 15 percent, biotite 15 percent, augite 3 percent, magnetite 1 percent, and trace amounts of sphene and apatite. Plutons of quartz monzonite also crop out in the Manly Peak quadrangle. It is porphyritic and is characterized by large phenocrysts of pink microcline. A thin section of quartz monzonite (Johnson, 1957, p. 392) contains quartz 40 percent, microcline as phenocrysts and as traces in the groundmass 15 percent, orthoclase 15 percent, andesine 20 percent, biotite 5 percent, and traces of hornblende, sphene, apatite, magnetite, and zircon. Johnson tentatively regards the quartz monzonite as late Jurassic in age. It intrudes the possible lower Triassic Warm Spring formation (Johnson, 1957, p. 393) and is overlain by volcanic rocks that are probably middle to late Tertiary in age (Johnson, 1957, p. 397).

Small stocks of possible Cretaceous granitic rocks crop out in the Panamint Range between the Manly Peak and Ubehebe Peak quadrangles. Prominent among these, in addition to the previously described intrusive rocks of the Wildrose area, are stocks in lower and upper Surprise Canyon, in Jail Canyon, and in the Harrisburg-Skidoo area. The lower Surprise Canyon and Jail Canyon masses are two feldspar-muscovite-quartz rocks with rare garnet, which are very similar to the granite of the Wildrose area. A specimen along the Harrisburg-Aguereberry Point road contains biotite instead of muscovite, but otherwise it is very similar to the granite in the Wildrose area. The upper Surprise Canyon stock, called the Little Chief porphyry by Murphy (1930, p. 311), is different. It is a holocrystalline porphyritic rock that contains three types of phenocrysts, which average about 3 millimeters in diameter, in a fine-grained groundmass (Dale Simpson, written communication, 1960). Two types of phenocrysts contain potassium feldspar and plagioclase in various zonal and rimmed relationships whereas the third type contains augite rimmed by hornblende and biotite.

It is apparent that the "Wildrose-type granite" does not resemble the other intrusive rocks of the Panamint Range. It is possible that two different ages of intrusive rocks are present in the range. The results of age studies are presented in Part II of this report.

GEOLOGIC STRUCTURE

General statement

The structural features of the Wildrose area include folds and faults which are ascribed to at least four periods of deformation. Structural features of the earlier deformations have been further deformed and to a certain extent obscured by the later deformations. The first deformation is evidenced by the irregular distribution of amphibolite within the lower metamorphic series of the Panamint complex. The second deformation is expressed by larger folds involving the rock sequence of the upper metamorphic series of the Panamint complex. The third deformation is evidenced by the folded thrust sheets of the Kingston Peak formation and later rocks which transect the structural features of the second deformation. The fourth period of deformation is expressed by high angle faults which cut the Tertiary and Quaternary sediments with final tilting going on today. The first two periods of deformation produced pre-Kingston Peak structural features; that is, structures developed before deposition of the Kingston Peak formation. Post-Kingston Peak structural features were produced by the third and fourth periods of deformation which occurred after deposition of the Kingston Peak formation.

Pre-Kingston Peak structures

The pre-Kingston Peak structures include all the recognizable structural features, except faults, that are developed in rocks of the Panamint metamorphic complex. These structures include major and minor folds, foliation or schistosity, and lineation. No faulting of pre-Kingston Peak age can be definitely demonstrated; consequently, all faults are assigned a post-Kingston Peak age.

Few small folds were observed in the Panamint complex, primarily because the micaceous quartzites are lithologically monotonous and generally do not have any layering, but also because the outcrop is poor in many places. In the upper series abundant tight folds are common within the marble layers, but these folds generally have no apparent relationship to either the stratigraphy or the major folds.

The terms foliation and schistosity have been used interchangeably by the writer to describe any planar structure that induces a rock to part along parallel, or nearly parallel, planes. These structures include layers of nonequant minerals such as mica, laminae which consist of different minerals, and lenses, pods, or layers of contrasting rock types. The perfection of a foliation or a schistosity depends primarily on the content of micaceous minerals, and in general, the foliation of rocks in the Wildrose area is only fair.

The linear features that have been mapped include mineral lineations and the intersections of planar features. Hornblende is the only mineral that produces a distinctive lineation; most of the lineations on Plate 1 were measured on hornblende. In addition, the intersection of the plane of foliation or apparent bedding with a vertical fracture or cleavage in places produces a lineation in the micaceous quartzites of the lower series.

The first period of deformation has been inferred because of the characteristic structural style of the lower series of the Panamint metamorphic complex in which no systematic pattern can be discerned in the overall irregular distribution of amphibolite. In some individual outcrops and small areas of amphibolite the foliation wraps around as in the nose of a small fold, but the amphibolite layers generally cannot be followed any distance in a systematic manner. Boudins or lenses and discontinuous layers of amphibolite characterize the deformation of the lower series and suggest profound deformation and perhaps intense shearing and plastic flow of the rocks. The amphibolites are believed to have formed from mafic volcanic material, but no evidence was observed suggesting whether dikes, sills, flows, or detritus was a more likely protolith.

Following the first period of deformation the upper series of the Panamint metamorphic complex was deposited on the intensely deformed lower series. Although rocks of the upper series are also deformed, a sequence of units can be traced across the entire area. This difference in structural style is the primary reason for inferring an unconformity between the two series. The entire Panamint metamorphic complex was folded and deformed and the major pre-Kingston Peak structural features, the Williams Canyon syncline and associated folds, were produced during the second period of deformation.

The Williams Canyon syncline (Plates 1 and 6) is a large syncline whose axis trends approximately north-south across the area and which is overturned to the west. The name synclinorium might be more appropriate as it is actually a composite structure (Plate 2, fig. 2). An anticline and another syncline lie east of the Williams Canyon syncline, but these structures are not given names as they are largely covered by younger bedrock or alluvium. In general the foliation or apparent bedding of rocks of the Panamint metamorphic complex dips eastward, so that it is commonly not possible to utilize the reversal of dip in order to locate the fold axes. Consequently the location of the trace of the axial plane of the Williams Canyon syncline has been dictated in places by the stratigraphy of the Panamint complex. Although there are some exceptions, the linear features in general indicate that the Williams Canyon syncline plunges moderately to the north. However, the pattern and stratigraphy of the upper and lower series of the Panamint metamorphic complex indicate that the overall plunge of the Williams Canyon syncline is to the south.

Unit A of the upper series is folded into a number of small folds whose axial planes are subparallel to the axial plane of the Williams Canyon syncline. South of the Summer Headquarters in Wildrose Canyon rocks of unit A crop out in two small folds whose axial planes trend approximately east-west. The axial plane of the Williams Canyon syncline is apparently refolded by the cross-folds which indicates that the east-west folds were developed after the north-south folds. The overall pattern of folds is consistent with a system of east-west compression, but whether the east-west cross folds represent buckling during compression, a minor component of north-south compression, or some other mechanism, is not known.

Post-Kingston Peak structures

Whereas folding of the early Precambrian rocks is prominent, faults are the predominant structures in the younger rocks. The Kingston Peak formation and the younger units generally dip gently to moderately eastward and their structure is essentially monoclinal. Exceptions to this generalization include folds west and southwest of Wildrose Peak and those farther west on the upper plates of the low angle faults. These latter folds probably were produced by fault movement and they are discussed in detail farther on.

Two types of faults have been either mapped or recognized in the Wildrose area. These types include the older low angle faults in the vicinity of Wildrose Canyon that are probably Cretaceous in age and high angle faults which probably formed in late Tertiary time, such as the north-trending Panamint Valley fault zone and the east-trending Tuber Canyon fault.

Low angle faults have been mapped on both sides of Wildrose Canyon. The name low angle fault is used because it is not known whether the younger rocks have been thrust into a position over the older rocks or whether the younger rocks have slid onto the older rocks by some mechanism involving gravity tectonics.

In general the traces of the low angle faults are poorly exposed because talus from the overlying, younger rocks has sloughed down the slope and has obscured the faults. The best exposed low angle fault is the one north of Wildrose Canyon. At this locality tan arenaceous limestone of the Noonday dolomite and conglomerate, quartzite, and metagraywacke of the South Park member of the Kingston Peak formation overlie metagraywacke of the Surprise member of the Kingston Peak formation. The fault itself is a subhorizontal zone of moderately crushed rock, generally less than 12 inches thick, that traces a sinuous course across many small gullies. However, the abrupt lithologic change at the fault was a more satisfactory mapping criterion, particularly in those areas where talus covers the fault trace.

The average dip of the faults on both sides of Wildrose Canyon is about 10° to the west; this common dip and similar elevation of the traces of the faults indicate that the outliers on either side of the canyon are remnants of the same fault-bounded slice. Possibly the isolated area of Noonday dolomite about 2 miles west of Wildrose Peak is also underlain by a low angle fault, but this exposure of arenaceous limestone could just as well be overlying the Kingston Peak formation unconformably.

Direct evidence of the direction of fault movement was not observed. The younger rocks in the upper sheet are folded or rumpled into folds whose axial traces have a general north-south trend. If these folds developed as drag folds at right angles to the direction of tectonic transport, they indicate an east-west direction of transport. If it is also assumed that the low angle faulting took place before the Panamint Range was tilted or rotated to the east, then the initial dip of the faults was westward at 20° to 30°. This analysis requires that a mechanism involving gravity tectonics have its source area to the east because the blocks cannot slide uphill. The age of low angle faulting is discussed farther on.

Possible origins of the low angle faults include thrusting from the west, gravity sliding from the east, and normal faulting along faults that dip gently westward. In a gravity sliding mechanism, large masses of rock would move downslope under gravitational forces. However, the masses in Wildrose Canyon have maintained stratigraphic integrity and are not strongly brecciated. The implications are that they probably did not move as talus piles or as landslides. Thrusting from the west can probably be disregarded because the sense of displacement would be opposite to that recognized farther east by Hunt. Therefore, normal faulting along gently westward dipping faults seems the most likely mechanism to explain the low angle fault-bounded slices.

On the eastern slope of the Panamint Range Hunt has mapped an imbricate series of low angle faults that dip gently to the west (oral communication, 1960). The displacement on these faults ranges from a few thousand feet to more than a mile. The faults trend approximately north-south, and individual faults have been mapped for several miles. Everywhere the rocks above the fault have been moved west relative to those below the fault so that younger rocks overlie older rocks. This is the same sense of displacement as in the Wildrose area. The amount of displacement on the fault in Wildrose Canyon is unknown but it probably is of the order of a few miles.

Hunt also has some evidence concerning the age of low angle faulting. In Hanaupah Canyon a sill of porphyritic rock has been emplaced along a major fault zone that dips gently westward. This intrusive rock is a fine-grained facies of the Little Chief porphyry, which occupies the crest of the range in the vicinity of Panamint City. Although it is not possible to precisely bracket its age the Little Chief porphyry is considered to be Cretaceous in age.

In the Wildrose area it is believed that the low angle faults were developed during the third period of deformation. Whether the faulting is older than Cretaceous cannot be proved, but no evidence has been found in the Death Valley region that indicates low angle faults that are post-Precambrian and pre-Cretaceous in age. The faulting and deformation of the Precambrian (?) rocks, for example the intricate folding in the "Radcliff lithology" of the Noonday dolomite (Plate 5), are considered contemporaneous. Therefore, the age of metamorphic minerals in the Precambrian (?) rocks (Part II of this report) places a lower limit of late Cretaceous on the time of thrusting.

Thrust faults have been mapped at several places in the Death Valley region, and directions of thrusting to the west, south, southwest, southeast, and northeast have been inferred (Noble, 1941; Kupfer, 1960; Cornwall and Kleinhampl, 1960). The thrusting has been dated as Mesozoic or post-Miocene. It is apparent that the structural framework of the Death Valley region is neither simple nor completely understood.

The north-trending Panamint Valley fault zone, named by Noble (1926, p. 425), is a series of breaks along which the movement has been largely vertical. This fault, or more properly fault zone, separates bedrock from alluvium and it marks the western boundary of the range. No individual faults in this zone are indicated on Plate 1, but a fault has been inferred at the contact between rocks of the Panamint metamorphic complex and the Nova formation (Plate 6). Although the Nova formation has probably been displaced it was not possible to map any faults because the formation has no marker horizons. Striking evidence of recent movement along the Panamint Valley fault zone is indicated by the Wildrose graben, a wide, down-faulted trench located just west of the area covered by Plate 1. The graben has steep walls up to 200 feet high, formed of Quaternary alluvium. Just north of Wildrose Canyon triangular-faceted spurs in rocks of the Nova formation also indicate recent movement along the zone.

In his investigation farther north, Hopper (1947, pp. 426-427) mapped five separate faults in the Panamint Valley fault zone. These trend north-northwest and dip steeply to the west; the faults constitute a series of steps with their east sides raised relative to their west sides. The total vertical displacement along this zone is probably more than 25,000 feet (Hopper, 1947, pp. 426-427). Four of the faults offset rocks of the Nova formation and this movement can be dated as late Tertiary in age. However, the fifth break, whose dip slip displacement is estimated as 20,000 feet on the basis of the juxtaposition of possible Devonian rocks on the west and Precambrian rocks on the east, does not offset fanglomerate of the Nova formation (Hopper, 1947, pp. 426-427). Consequently, the age of movement along this fault can only be bracketed as post-possible Devonian and pre-late Tertiary.

The major high angle fault in the Wildrose area has been named the Tuber Canyon fault (Plate 1). Its trend is approximately east-west and Tuber Canyon has been developed along the fault, at least in part. The fault has been mapped for a distance of about 6 miles from west of Bennett Peak to lower Tuber Canyon and it probably continues to the range front, but as the fault enters the lower series of the Panamint metamorphic complex its offset cannot be mapped because distinctive marker horizons are absent. The Tuber Canyon fault is vertical or nearly vertical, and the stratigraphic units and structures offset by the fault require dip slip movement along it with the north side raised relative to the south side. Offset of the Sour Dough limestone member of the Kingston Peak formation on the south side of Tuber Canyon, coordinates: 32,250E. to 34,000E. at 2,100N., indicates the best estimate of the amount of displacement. At this location the horizontal offset is 1,690 feet which, assuming a dip of 55° E. for the limestone, is equivalent to 2,400 feet of vertical movement. Although the relative displacement of the Sour Dough limestone could be produced by horizontal movement, the displacement of the Williams Canyon syncline can only be explained by vertical movement.

Several other high angle faults have been mapped in the area (Plate 1), but with the exception of the faults west of Wildrose Peak most of these can be classified as small. A vertical displacement of about 750 feet has been inferred for the fault west of Wildrose Peak that bifurcates near the northern boundary of the area.

The mechanism that produced the high angle faults in the Wildrose area is not known, but they might have been developed in order to relieve stresses produced in the Panamint block by movement on the Panamint Valley fault zone. That is, the vertical movement along one segment of the Panamint Valley fault zone would introduce stresses into the range block that could be released by faults developing at a high angle to the valley fault zone. In the Wildrose area many of the faults, but not all, have a trend that is at a high angle to the range front. However, there is no way to verify this hypothetical mechanism or the age of the faulting, if such a mechanism did operate.

Movement along the high angle faults probably began in late Tertiary time and has continued to the present time. The overall displacement on the Panamint Valley fault zone has resulted in rotation or tilting of the Panamint Range block to the east. Tilting of 10° to 20° to the east is indicated by the attitude of the fanglomerates of the Nova formation. These fanglomerates at present dip eastward about 5° to 10°; but, assuming that they formed as fan gravels, their initial dip must have been 5° or 10° to the west. That movement has continued to the present is indicated by recent displacement of Quaternary alluvium. Tiltmeter observations on the floor of Death Valley indicate that measurable tilting is occurring at the present time (Greene and Hunt, 1960).

METAMORPHISM

General statement

All bedrock formations of the Wildrose area except Cretaceous granite and Tertiary sedimentary rocks have been affected by regional metamorphism. The metamorphic rocks have mineral assemblages that are typical of middle grade metamorphism. The structural evidence on intensity of deformation suggests that metamorphism must have accompanied the earlier periods of deformation. However, the present mineral assemblages seem to reflect the third period of deformation as essentially no textural or mineralogic evidence remains of the earlier deformations. Hence, the mineral assemblages that follow are ascribed to the third period of deformation. In addition, narrow contact metamorphic aureoles surround the larger masses of granite, which were emplaced at the end of the third period of deformation. The two types of metamorphism are described separately and a summary statement is included at the end of the section.

Regional metamorphism

Estimated modes of typical rock types have been included with the descriptions of individual stratigraphic units. The mineral assemblages are tabulated in Table 11 according to the general lithologic type; that is, arenaceous or argillaceous, mafic, and calcareous. These assemblages indicate a lower middle to middle grade of regional metamorphism.

The typical mineral assemblage in the arenaceous or argillaceous rocks is quartz-muscovite-biotite; chlorite and plagioclase are commonly present. Garnet and andalusite are rare and are commonly partially or completely altered to aggregates of white mica. These relicts might be interpreted as evidence of the earlier metamorphism; however, such features are also present in theJohnnie formation which was not deposited until after the earlier deformations. The rarity of garnet and andalusite might be ascribed to more complete retrogradation in part of the area or to differences in chemical composition of the rocks such that these minerals did not form.

Andalusite is commonly considered a contact metamorphic mineral, but it is also abundant in regional metamorphic terranes of the Buchan type (Read, 1952; Miyashiro, 1958). In regional metamorphism of the Buchan type the critical index minerals are found in rocks that belong to every grade of metamorphism. However, it has been recognized that conspicuous porphyroblasts of andalusite

Table 11. Mineral assemblages of metamorphic rocks of the Wildrose area

<u>arenaceous or argillaceous</u>

quartz-muscovite-plagioclase
quartz-muscovite-biotite
quartz-muscovite-biotite-chlorite
quartz-biotite-chlorite
quartz-muscovite-graphite
quartz-garnet-muscovite-biotite-chlorite
quartz-muscovite-biotite-epidote-sphene
quartz-andalusite-muscovite-biotite-chlorite

<u>mafic</u>

hornblende-quartz-andesine-epidote-sphene
hornblende-quartz-andesine-epidote-biotite-chlorite-sphene
hornblende-quartz-andesine-potassium feldspar-epidote-chlorite

<u>calcareous</u>

calcite-quartz
calcite-quartz-diopside
dolomite-quartz
dolomite-tremolite-quartz
calcite-quartz-muscovite
calcite-quartz-muscovite-biotite
calcite-quartz-muscovite-biotite-epidote

indicate a higher grade than the incipient crystals of "knotted schists" and that the assemblages, in the following order, andalusite-biotite, andalusite-garnet, and andalusite-staurolite indicate an increasing grade of metamorphism (Harker, 1932, p. 232). In the Wildrose area andalusite occurs only in the lower member of the Johnnie formation and this restriction to a single stratigraphic unit presumably means that the chemical composition of many of the other units in the area are not favorable for the development of andalusite. The assemblage andalusite-biotite in the Johnnie formation implies a middle grade of metamorphism.

The most common mineral assemblage in the mafic rocks is hornblende-andesine-quartz-epidote. Hornblende is the characteristic amphibole; actinolite, which is typical of low grade metamorphism does not occur in the amphibolites. Sphene, the typical accessory mineral in the mafic rocks, is common in middle grade rocks. The occurrence of epidote and the absence of garnet implies a lower middle or middle grade rather than an upper middle grade of metamorphism. The mineral assemblages of the mafic rocks seem to indicate a higher grade of metamorphism than those of the argillaceous rocks, but this apparent discrepancy is considered related to the chemical composition of the argillaceous rocks.

The calcareous rocks also indicate a middle grade of metamorphism though the mineral assemblages are not as distinctive as those of the argillaceous (or arenaceous) and mafic rocks. Calcite-diopside-quartz occurs in unit A of the upper series of the Panamint complex and dolomite-tremolite-quartz occurs in the Noonday dolomite.

Both of these assemblages are typical of middle or lower middle grade metamorphism.

The relationship of metamorphic minerals to structural feature is not clear. However, the lack of fracturing in minerals such as the tremolite rosettes in the Noonday dolomite and the preferred orientation of micas with respect to minor structures related to the third deformation implies that metamorphism continued until after the development of the structural features that mark the third deformation, the low angle faults and open folds in the Kingston Peak formation and younger units.

The regional metamorphic rocks of the Wildrose area contain mineral assemblages that Fyfe, Turner, and Verhoogen assign to their hornblende hornfels facies (1958, pp. 205-211). However, they use the term hornfels to imply contact metamorphism; the present writer considers that Buchan type metamorphism is a more appropriate term for these rocks because they have been developed by regional metamorphism.

Contact metamorphism

A distinct contact metamorphic aureole surrounds the granitic stock in lower Tuber Canyon. The aureole varies in thickness but it is probably not more than 100 feet thick. The stock intrudes rocks of the lower series of the Panamint metamorphic complex and it has produced distinctive mineralogic changes in both micaceous quartzite and amphibolite. The micaceous quartzite has been metamorphosed to a coarse-grained garnet-biotite-quartz schist. The grain size of the amphibolite in the contact zone is coarser than in the normal amphibolite but no mineralogic changes are obvious in the hand specimen. However, a monoclinic pyroxene, probably diopside, is an abundant mineral in a thin section (Table 12, sample 42). This specimen of amphibolite was collected in the area of mixed rock (Kmr), Plate 1, within a few inches of a large granite dike.

No well-defined aureole surrounds the granite stock in upper Wildrose Canyon, and a slight coarsening in grain size near the contact is the only indication of any contact metamorphism. A thin section of metagraywacke that was collected about 2 inches from the granite contact (Table 12, sample 43) does not indicate any major changes in mineral composition. The absence of mineralogic changes might be related to the chemical composition of the Kingston Peak formation, but it seems likely that the contact effects of the upper Wildrose Canyon stock are minor.

The observations of the extent and intensity of contact metamorphism associated with intrusion of the Cretaceous granitic rocks lead to the conclusion that the country rock more than about 200 feet from a granite body has not been affected, that is, heated, sufficiently to cause reaction between or recrystallization of the constituent minerals.

Summary of metamorphism

The two most important conclusions that are drawn from the observations of the metamorphic rocks of the Wildrose area are (1) the rocks of every unit, from the Panamint metamorphic complex through the Johnnie formation, have been affected by regional metamorphism, and (2) the metamorphic effects with a direct spatial

Table 12. Estimated modes of contact metamorphic rocks

	42	43
hornblende	38	
quartz	25	33
plagioclase	15 (An_{32})	
muscovite		10
biotite		25
chlorite		15
monoclinic pyroxene	20	
clinozoisite-epidote		15
hematite		1
sphene	2	1
calcite		T
apatite		T
graphite		T
map coordinates	5,950 E. 2,650 N.	36,400 E. 21,850 N.

relation to Cretaceous intrusive rocks have been slight.

The writer concludes that the first two periods of deformation were intense enough to have metamorphosed the rocks that were involved. The evidence for profound deformation and consequent metamorphism of the lower series of the Panamint complex has been presented in an earlier section. The deformation that produced the Williams Canyon syncline and associated folds after deposition of the upper series is also considered to have metamorphosed the entire Panamint complex. Finally, the Panamint complex was involved in the third period of deformation that metamorphosed the Kingston Peak formation, the Noonday dolomite, and the Johnnie formation. This polymetamorphic history is not reflected in either the mineral assemblage or texture of rocks of the Panamint complex, but this is really not too surprising considering the grade of metamorphism presently shown by the terrane. That is, relict high-grade minerals might remain as disequilibrium components in a subsequent middle or low grade metamorphism, but if the earlier metamorphism were of low or middle grade probably no trace of it would be retained in the mineral assemblage. Thus it is concluded that the rocks of the Panamint complex were completely recrystallized during the metamorphic episode that affected the Kingston Peak and younger formation.

The next problem concerns the age of this period of metamorphism. In his study of the Barstow quadrangle O. E. Bowen (1954) mapped open folds and an unconformity in possible Permian rocks that he considers evidence of Paleozoic deformation. But the present writer does not interpret this deformation to be sufficiently intense to metamorphose the rocks involved. The upper Paleozoic section near Barstow, which may also be lower Mesozoic in part (McCulloh, 1954, p. 17), has been affected by both regional and contact metamorphism (Bowen, 1954, p. 31). Although the contact effects are related to the late Mesozoic, probably Cretaceous, intrusive rocks, no date other than pre-intrusion can be assigned to the regional metamorphism.

Other areas of metamorphosed possible Paleozoic rocks crop out in the western Mojave desert (McCulloh, 1954) but it has not been possible to date the time of metamorphism on a stratigraphic basis. In the eastern Mojave desert the Paleozoic and Mesozoic sedimentary rocks are unmetamorphosed except in the vicinity of Mesozoic plutonic masses.

On the basis of this brief analysis of geologic relationships in the Mojave desert the writer concludes that the regional metamorphism in the Wildrose area and the western Mojave desert can be assigned to the Nevadan orogeny because it is the only known geologic "event" of sufficient magnitude to cause regional metamorphism over this extensive area. Radioactive ages of metamorphic rocks from the Wildrose area and the Mojave desert, ranging from 75 to 85 million years, indicate that the metamorphism occurred in the Cretaceous period.

The Cretaceous intrusive rocks in the Wildrose area were probably emplaced after the regional metamorphic episode was completed as evidenced by the non-foliated texture of the granite and the contact aureole around the stock in lower Tuber Canyon. In the Wildrose area the intrusive rocks were not observed cutting any structural features of the metamorphic rocks. However, in the similar geologic setting of the Sierra Nevada Range, the folds and faults in regional metamorphic rocks have been cross cut by Cretaceous intrusive rocks (MacDonald, 1941; Taliaferro, 1942; Clark, 1960).

That the contact metamorphic effects of the Cretaceous intrusive rocks in the Wildrose area, as shown by mineralogic changes in the vicinity of the granite stocks, are slight has been described previously. This observation is important because it means that the age of metamorphic rocks in the Wildrose area is the age of regional metamorphism and is not due to contact metamorphic effects.

PART II: GEOCHRONOLOGIC STUDIES IN THE DEATH VALLEY-MOJAVE DESERT REGION, CALIFORNIA

INTRODUCTION

In order to delineate the age and extent of the Precambrian basement in part of southern California, an investigation of the distribution of ages of the Precambrian rocks of the Death Valley-Mojave Desert region was undertaken. The geologic mapping of other workers was utilized wherever possible. It was hoped that the combination of geologic mapping and absolute age determinations might yield a basis for correlating the early Precambrian rocks. Several areas of crystalline rocks were selected for the age studies. In some areas the early Precambrian crystalline rocks are overlain by fossiliferous Cambrian rocks. The fact that the overlying sedimentary rocks are unmetamorphosed indicates that the Precambrian age of the crystalline rocks has probably not been affected by orogenic events since Cambrian time. In other areas the age assignment has been a matter of considerable speculation because of the lack of stratigraphic control. These areas are of great interest to Precambrian geochronology because they make up the major part of the basement of the Mojave Desert.

In order to prevent any confusion, the term Precambrian as used in this report is defined as follows. The top of the Precambrian system is placed at the major unconformity below the strata that contain lower Cambrian fossils. The base of the Cambrian system is ambiguous as it is defined on the basis of the appearance of certain fossils and not on a change in faunal assemblage. In this report the base of the Cambrian is placed at the lowest occurrence of lower Cambrian fossils. The rocks above the unconformity but below the lowest occurrence of lower Cambrian fossils are assigned a Precambrian (?) age; this stratigraphic interval has been designated by other workers as Eocambrian and Infracambrian.

Detailed age studies were carried out in three areas, the Mountain Pass district, the Marble Mountains, and the Wildrose area (Plate 7). In addition, age measurements were made on two other areas, Joshua Tree National Monument and the Kilbeck Hills. A Precambrian age has been assigned to the latter areas, primarily on the basis of degree of metamorphism of the rocks, because no stratigraphic evidence has been found that indicates a Precambrian age.

MOUNTAIN PASS DISTRICT

The Mountain Pass rare-earth district is an area about 7 miles long and 3 miles wide centered near Mountain Pass, California, 60 miles southwest of Las Vegas, Nevada, on U. S. Highway 91 (Plate 7). The Mountain Pass district lies within a block of metamorphic rocks of early Precambrian age that is about 55 square miles in area. This block is bounded on the east and west sides by faults that converge at the north end of the block and which juxtapose lower Paleozoic sedimentary rocks and early Precambrian metamorphic rocks. Quaternary alluvium overlies the block of metamorphic rocks on the south and east sides.

No fossiliferous rocks are found within the Mountain Pass block, but a reasonable correlation can be made between the gneissic rocks at Mountain Pass and strongly sheared granitic gneiss in the Winters Pass Hills, less than 20 miles northwest of Mountain Pass. In the Winters Pass Hills, the gneiss is overlain unconformably by the Noonday dolomite (Hewett, 1956, p. 29). The Noonday dolomite contains no organic remains and is considered Precambrian (?) in age, but it lies conformably beneath rocks which carry a Lower Cambrian fauna. Less than 10 miles west of the Winters Pass Hills the Noonday dolomite overlies rocks of the later Precambrian Pahrump group. The data presented in this work confirm the Precambrian age assignment of the rocks at Mountain Pass.

Rare-earth minerals have been found in the metamorphic block in a N30°W-trending zone about 6 miles long and 3 miles wide in which potassium-rich igneous rocks and carbonate rocks of probable Precambrian age occur. The mineral deposits and associated rocks have been studied in detail by Olsen, Shawe, Pray and Sharp (1954). The potassium-rich igneous rocks range in composition from shonkinite through syenite to granite. The rare-earth mineralization is bounded on the north by the North fault that trends N70°W across the metamorphic block. Northwest-trending shonkinite dikes and carbonate veins which contain rare-earth minerals are abundant south of this fault but none have been found north of it. The larger masses of potassium-rich igneous rock are non-foliated stocks that have sharply discordant intrusive contacts. Around the stocks marginal brecciation of the metamorphic country rock and local contact metamorphism are common (Olsen, Shawe, Pray, and Sharp, 1954). The Sulfide Queen carbonate body contains inclusions of gneiss and igneous rocks, and carbonate veins cut across the igneous and metamorphic rocks.

The metamorphic rocks have been divided by Olsen and others (1954, pp. 5-7) into four rather poorly-defined mapping units: unit A - biotite granite gneiss complex, unit B - mixed granitic gneiss complex, unit C - granitic augen gneiss complex, and unit D - sillimanitic biotite-garnet gneiss complex. The age relationships between these units are not clear, but Olson and others consider that the age sequence is A (oldest) - B - C - D (youngest). Simple pegmatites of granitic composition occur in all the metamorphic units but the largest pegmatites are found in unit B. The pegmatites consist chiefly of pink to white potassium feldspar and quartz, with minor gray to white plagioclase, muscovite, and rare garnet (Olsen and others, 1954, p. 7). In some places the pegmatites have sharp contacts and cut discordantly across the foliation of the metamorphic rocks, whereas in other places the pegmatites are concordant and/or seem to grade into granitic material. Olsen and others (1954, p. 8) report that the pegmatites are commonly gneissose; they believe that the pegmatites were intruded after most of the complex had formed but they have been affected by some of the metamorphism.

Minerals from five different rocks were used in the age study at Mountain Pass. The sample locations are indicated on the generalized geologic map (Figure 2), and estimated modes of four of the rocks and of grain mounts of the mineral concentrates are presented in Table 13.

Biotite was separated from two samples of metamorphic country rock. One of these, MP-7, was collected about 3000 feet

Figure 2- Generalized geologic map of the Mountain Pass district, California

Table 13. Estimated modes of early Precambrian rocks, Mountain Pass district, California, and of mineral concentrates separated from the rocks

rocks

	MP-7	MP-9	MP-21	MP-22
quartz	58	59		
biotite	15	15	40	20
chlorite		T		
muscovite	2		28*	
plagioclase	25 (An_{25})	20		
garnet		5		
potassium feldspar				43
monoclinic pyroxene			25	30
magnetite	T	1	3	5
hematite			1	
apatite			3	2
zircon	T		T	

mineral concentrates

	MP-7	MP-9	MP-21	MP-22
biotite	98.2	98.3	90.4	97.4
quartz	1.2	0.4		
plagioclase	0.6	0.4		
garnet		0.9		
monoclinic pyroxene			9.6	2.6

*fine-grained white mica

northeast of the Birthday shaft. It is a coarse-grained biotite-plagioclase-quartz gneiss from unit B (Olsen and others, 1954). The gneiss produces a rather non-resistant outcrop which is somewhat decomposed, and the biotite on the surface is stained a brown color. However, beneath a thin crust of weathered material, the biotite is very fresh and shows no evidence of weathering. In thin section the biotite has a very good parallel preferred orientation; it is very fresh and has virtually no chloritization.

The other sample of metamorphic rock, MP-9, was collected in a road-cut on U. S. Highway 91. It is a massive, dense, very fresh, coarse-grained garnet-biotite-plagioclase-quartz gneiss from unit D (Olsen and others, 1954). In thin section the parallel preferred orientation of biotite plates is good though not as pronounced as in MP-7. In MP-9 some biotite grains are partly chloritized, but less than half a percent of the total biotite has been altered to chlorite. Migmatitic segregations are common in the vicinity of the sample location. About 200 feet west of MP-9 a small biotite-garnet-plagioclase-quartz pegmatite, about 1 to 2 feet thick, crops out which is conformable with the foliation of the gneiss. The borders of the pegmatite are gradational into the gneiss. The biotite in the pegmatite occurs as plates about 1 inch square; it is discolored, friable, inelastic, and soft. The biotite was water-logged, and apparently the pegmatite acts as a water course. However, such alteration as shown by the pegmatite is not pervasive, and most of the gneiss is very fresh.

Muscovite (MP-1) and pink potassium feldspar (MP-2) from a granitic pegmatite, located about 200 feet east of MP-7, were analyzed. The pegmatite dike is about 6 feet thick and is slightly discordant with the foliation of the gneiss. Pink and white potassium feldspar, muscovite, and quartz form the major portion of the pegmatite; schorl and beryl (?) are minor constituents. Books of coarse muscovite up to 6 inches x 6 inches x 1 inch are abundant in the massive quartz core which is about 4 feet square. The quartz core apparently acted as a buttress and protected the muscovite from later shearing, as this was the only pegmatite found in the area in two days that contained either coarse muscovite or a quartz core. All the other pegmatites, both north of the North fault (Figure 2) and south of U. S. Highway 91, are strongly sheared. They contain quartz and potassium feldspar but do not contain coarse muscovite; the only mica is fine-grained muscovite "smeared" along the shear planes.

A strongly sheared quartzo-feldspathic rock crops out a few hundred feet east of the muscovite-bearing pegmatite described previously. In hand specimen the rock is medium grained and shows marked granulation. In thin section the granulation of minerals produces a distinctive texture; "milled" residual grains of potassium feldspar and quartz are imbedded in a fine-grained matrix of quartz, potassium feldspar, and muscovite. The quartz is highly strained and shows pronounced undulant extinction. The rock was probably originally a granitic rock, but its relation to the granitic pegmatites is not known.

The writer's observations on the shearing of pegmatites confirm Olsen and others (1954, p. 8). The writer concludes that the pegmatites were intruded sometime during the period of metamorphism and were then sheared during a later part of the metamorphism. However, the muscovite (MP-1) is not bent and the potassium feldspar (MP-2) is not badly broken so the later cataclasis has probably not affected the ages of these minerals.

Two samples of shonkinite (MP-21 and MP-22) were collected from the dump at the Birthday shaft. The shonkinite was removed during the sinking of the shaft in 1950 (Russell Wood, oral communication, 1960), and was much fresher than any surface outcrops. In hand specimen MP-21 is medium-grained biotite-pyroxene shonkinite, and MP-22 is very coarse-grained biotite-pyroxene shonkinite which contains scattered gray potassium feldspar. Neither specimen contains abundant pink potassium feldspar which is typical of much of the shonkinite. The biotite in MP-22 is rimmed by a narrow bleached zone that is visible in hand specimen. In thin section, the biotite in both specimens is very fresh. However, that deuteric or hydrothermal alteration took place is indicated by the complete alteration of potassium feldspar to white mica in MP-21 and the corrosion of pyroxene and patchy alteration of potassium feldspar to white mica in MP-22. A sample of syenite from the Birthday body also indicates significant deuteric or hydrothermal alteration as evidenced by the nearly complete alteration of potassium feldspar to clay minerals.

The experimental data are presented in Table 14. The results on muscovite and the Rb-Sr age of potassium feldspar from the pegmatite (MP-1 and MP-2) are concordant and are interpreted as the absolute age of intrusion of the pegmatite. The age of approximately 1650 million years is in general agreement with

Table 14. Analytical data and radioactive ages for minerals from the Mountain Pass district, California

sample	A^{40*}(cc STP/gm)	A^{40*}/A^{40}	percent K	K-A(m.y.)	Sr^{87*}(ppm)	Sr^{87*}/Sr^{87}	Rb(ppm)	Rb-Sr(m.y.)
MP-1	0.913×10^{-3}	0.96	8.62	1660	5.50_8	0.95	841	1640
MP-2					3.61_6	0.32	566	1600
MP-7	0.723×10^{-3}	0.96	7.45	1560	6.51_1	0.93	1122	1460
			7.47					
			avg: 7.46					
MP-9	0.732×10^{-3}	0.92	7.39	1580	3.26_6	0.78	629	1310
MP-21	0.408×10^{-3}	0.99	6.11	1200	4.27_1	0.22	762	1410
MP-22	0.418×10^{-3}	0.65	7.55	1040	3.59_1	0.09	649	1390
			7.62					
			avg: 7.58		3.87_1 (a)			1500

MP-1 : muscovite-pegmatite
MP-2 : potassium feldspar - pegmatite
MP-7 : biotite - gneiss
MP-9 : biotite - gneiss
MP-21 : biotite - shonkinite
MP-22 : biotite - shonkinite

(a) isotope ratio analysis

the concordant age of approximately 1700 million years measured on muscovite and potassium feldspar from a pegmatite in the southern Panamint Range (Wasserburg and other, 1959). The gneissic rocks that the pegmatite cross-cuts have K-Ar ages that are approximately 5 percent younger than the age of the pegmatite but they are not in serious disagreement. The data give results that disagree with the field relationships; this relationship between the radioactive ages of pegmatites and country rock has been found in other areas, but the reason for the discrepancy is not presently known.

The biotite from both samples of gneiss and the biotite from both samples of shonkinite have discordant K-Ar and Rb-Sr ages. The discrepancies, almost all of which are outside of experimental error, range from 7 to 28 percent. No regularity is observed in the discordance; in the gneiss samples the K-Ar ages are greater than the Rb-Sr ages, and in the shonkinite the reverse is true. The cause of the discordant ages is not known, but the fact that all the discrepancies were obtained on biotite may be significant. In other areas the common discordance in ages is such that the K-Ar age is lower than the Rb-Sr age. The biotite from the shonkinite shows this type of discordance. The biotite from the gneisses illustrate the less common case where the Rb-Sr age is lower than the K-Ar age. However, this second type of discordance has also been obtained on biotite from the Llano region of Texas (R. E. Zartman, 1961, oral communication). Zartman reports that in a strongly foliated granite, which is part of a composite pluton, the Rb-Sr age of biotite is about 25 percent less than the Rb-Sr age of coexisting potassium feldspar, and the K-Ar age of the biotite agrees with the Rb-Sr age of the feldspar. Also, in an associated pegmatite, the K-Ar age of biotite is 15 percent less and the Rb-Sr age is 20 percent less than the Rb-Sr age of potassium feldspar. Although the reason for the discordance is not known it is apparent that the interpretation of ages measured on biotite must be done with caution.

The K-Ar age of 1200 million years for MP-21 is in good agreement with the concordant ages of biotite from the early Pre-cambrian granite of the Marble Mountains (Table 16). However, the K-Ar age of MP-22, 1030 million years, is considerably younger than any other age at Mountain Pass. A K-Ar age of 980 million years has been measured on biotite from the southern Nopah Range (Wasserburg and others, 1959). The Rb-Sr ages of the shonkinite samples are significantly older than the K-Ar ages; an average value for the three Rb-Sr ages is 1430 $\pm$ 40 million years. The difference of about 7 percent between the two Rb-Sr ages of MP-22 is to be expected because the Sr^{87} is only 9 percent radiogenic. For MP-22 the age of 1500 million years, determined by isotope ratio analysis, is the better number. The potassium feldspar in MP-21 and MP-22 is largely altered and not suitable for dating purposes. A specimen of shonkinite which contained fresher potassium feldspar was obtained from Professor L. T. Silver, but it was not possible to measure a RB-Sr age because the common strontium content was too high. By X-ray fluorescence the Rb/Sr ratio was approximately 1/3.

The present writer did not do any analytical work on the youngest Precambrian unit at Mountain Pass, the carbonate rock that is presumably of intrusive origin (Olsen and others, 1954). However, other workers have analyzed monazite from the Sulfide Queen carbonate body (Jaffe, 1955; Tilton, written communication, 1961). For monazite SQ-81 Jaffe obtained the following data:

Pb - 0.113 weight percent (spectrographic)

U - 0.002 weight percent (fluorimetric)

Th - 2.65 weight percent (colorimetric)

The lead has the following isotopic composition, in atom percent (analysis by the Carbide and Carbon Chemicals Co., Oak Ridge, Tennessee):

204	206	207	208
0.0168	0.782	0.258	98.943

From the data Jaffe calculated a Pb^{208}/Th^{232} age of 925 million years, which was in good agreement with Pb-α ages measured on other monazites from the carbonate body and on zircon separated from the shonkinite. However, his lead concentration has subsequently been proved low and all of his ages at Mountain Pass are at least 40 percent low. Tilton (written communication, 1961), obtained a value of 0.175 weight percent by isotope dilution for the lead concentration of monazite SQ-81. Using Tilton's lead concentration and Jaffe's thorium concentration and lead composition, the present writer calculates a Pb^{208}/Th^{232} age of 1425 ± 70 million years. Within experimental error this age agrees with the

Rb-Sr ages of biotite from the shonkinite and is, therefore, consistent with the geologic conclusion that the potassium-rich igneous rocks and the carbonate rocks are genetically related and are products of a single geologic episode (Olsen and others, 1954, p. 59).

Other workers have found that commonly the most reliable ages are those measured on pegmatite minerals. Because it is believed that the pegmatites were emplaced during the metamorphic episode, probably near the end, the concordant pegmatite ages are interpreted as both the age of injection of the pegmatite and of metamorphism of the terrane. All the biotite ages are discordant, and, as the most common cause of discordance is loss of daughter isotope, the biotite ages should be considered minimum values. However, the K-Ar ages of biotite from the gneiss are in good agreement with the pegmatite ages and tend to confirm them.
An important question that is not resolved by the data is the actual age of the potassium-rich igneous rocks and the associated rare-earth deposits. The field relationships indicate that these rocks are younger than the metamorphic rocks, but their difference in age cannot be predicted. Pb^{208}/Th^{232} ages have commonly been found to be lower than Pb/U ages, and the monazite age of 1425 million years should be considered a minimum age. The average Rb-Sr age of 1430 million years for MP-21 and MP-22 also should be considered a minimum age. The age studies in the Marble Mountains (Plate 7) indicate emplacement of granitic rocks approximately 1350 million years ago. Therefore, the writer believes that these data suggest a younger intrusive event in the area,

distinct from the 1650 million year metamorphism and pegmatite intrusion and older than 1430 million years.

MARBLE MOUNTAINS

The Marble Mountains are a northwest-trending range about 20 miles long and up to 4 miles wide that is located approximately 90 miles east of Barstow, California, on U. S. Highway 66 (Plate 7). The range is critical to Precambrian geochronology as it is the most southwesterly locality in the continental United States in which sedimentary rocks that contain lower Cambrian fossils unconformably overlie Precambrian crystalline rocks. Hazzard and Crickmay (1933) studied the Cambrian stratigraphy and paleontology in detail, but they did not describe the Precambrian rocks. Therefore, the writer mapped a portion of the Marble Mountains at a scale of 1:24000 (Plate 8); he concentrated in particular on subdividing the Precambrian rocks.

The lower Cambrian units include the Prospect Mountain quartzite at the base, the Latham shale, and the Chambless limestone (Hazzard, 1954). No fossils are found in the Prospect Mountain quartzite, but it is apparently conformable with the overlying Latham shale which contains a lower Cambrian fauna. A short description of these units, as taken from Hazzard and Crickmay (1933), is presented below.

The Prospect Mountain quartzite is a white to reddish brown, cross-bedded quartzite that is locally conglomeratic. Lenses of conglomerate are common in the lower part of the formation but a typical basal conglomerate is not present. Pebble counts in the conglomerate lenses indicate 35 to 40 percent each of white quartz and banded red jasper and a few percent of chert, quartzite, rhyolite, and basic volcanic rock. The pebbles of jasper, including agate and chalcedony, resemble amygdule fillings and they imply that the source area for the quartzite contained volcanic rocks. A few angular fragments of vein quartz, which were observed only a few inches from quartz veins in the underlying granite, can be related to a local source; otherwise, the quartzite is apparently free of rock fragments of local origin (Hazzard and Crickmay, 1933, pp. 61-63).

The Prospect Mountain quartzite ranges in thickness from 390 to 450 feet and it is overlain by the Latham shale which is 40 to 45 feet thick (Hazzard and Crickmay, 1933, pp. 61-63). The shale is greenish gray and commonly weathers a reddish color. It contains abundant fragments of trilobites that have been referred to the Paedeumian age of Olenellidae (Hazzard and Crickmay, 1933, p. 72). A more primitive genus of Olenellidae, Nevadia, has been identified in other parts of the Death Valley area (Nelson, 1961).

The Chambless limestone, a gray limestone that ranges in thickness from 100 to 120 feet, is characterized by the presence of ovoid nodules that are presumably algal in origin (Hazzard and Crickmay, 1933, p. 63). Lower Cambrian trilobites have been found in shaly horizons in the limestone.

The present writer divided the early Precambrian crystalline rocks into four mappable units: metamorphic rocks, granodiorite, granite, and roof pendants which include various lithologic types. A marked unconformity, a smooth surface that has very little relief, separates the early Precambrian rocks from the overlying Prospect Mountain quartzite. In the southern part of the Marble Mountains the unconformity truncates scattered quartz veins and pegmatitic stringers in the granite. No evidence of later fault movement along the unconformity was observed. The lower Cambrian formations showed no visible evidence of metamorphism.

The metamorphic rocks, primarily poorly-foliated quartz-feldspar-biotite-(hornblende) gneisses that are interbedded with thin layers of marble, crop out only in the northern part of the area (Plate 8), but similar gneissic rocks plus garnetiferous varieties occur in roof pendants south of U. S. Highway 66. The metamorphic rocks are presumably the oldest units in the Marble Mountains basement.

Granodiorite is considered the next younger unit though no evidence of its relationship to the metamorphic rocks was observed. The granodiorite is a holocrystalline, medium- to coarse-grained, non-foliated rock. It is dark colored and might seem to be quite mafic; however, the high content of plagioclase (labradorite) is responsible in part for its dark appearance. Because of its high quartz content the rock is called granodiorite, but it is close to diorite in composition (Table 15, MM-11).

The granite is the youngest rock type that is early Precambrian in age. It intrudes the granodiorite and contains xenoliths and roof pendants of various types of metamorphic and more mafic igneous rocks. The most abundant rock type in the roof pendants is a hybrid rock (Table 15, MM-1) that is considered to be

Table 15. Modal analyses of early Precambrian rocks of the Marble Mountains, California

	MM-3[1]	MM-1[2]	MM-11[2]
quartz	28.1	45	15
potassium feldspar	30.2	15	10
plagioclase	21.6 (An_{32})	15 (An_{35})	40 (An_{54})
muscovite	0.1	--	--
biotite	16.6	10	25
chlorite	0.1	2	--
hornblende	0.3	15	7
magnetite	1.2	2	3
hematite	0.2	--	--
sphene	--	--	T
zircon	0.8	T	T
apatite	0.2	1	T
epidote	0.6	T	--

1 - average of 3,500 points counted on 3 thin sections

2 - estimated mode

granodiorite which has been contaminated by the granite magma. The hybrid rock may possibly be a third variety of igneous rock; in this case its relationship to the granodiorite is not known. The hybrid rock is a holocrystalline porphyritic rock which has a color index of approximately 30. It consists of phenocrysts 5 to 10 millimeters in size in a groundmass whose average grain size is less than 1 millimeter. Hornblende, biotite, chlorite, quartz, plagioclase, potassium feldspar, and magnetite are the principal components of the groundmass, and the phenocrysts are quartz and potassium feldspar. The potassium feldspar phenocrysts are commonly euhedral, but the quartz phenocrysts are typically anhedral, ovoid masses. The potassium feldspar phenocrysts resemble those in the granite, and as the abundance of the phenocrysts increases near the contacts between roof pendant and granite it seems likely that the phenocrysts were developed during "soaking" of the roof pendant in the granite magma. The quartz phenocrysts presumably developed in the same manner, but their relationship is not clear.

The granite is a porphyritic, coarse-grained, non-foliated rock that ranges in color from gray to red depending on the degree of weathering. In the vicinity of U. S. Highway 66 the granite is typically decomposed or disaggregated and the flakes of biotite are covered with a hematite stain. The granite is cut by numerous small faults, and pervasive shearing is a characteristic feature. However, farther south the shearing is not apparent and the granite is much fresher. It consists of grains of biotite, plagioclase, and quartz from 1/16 to 1/4 inch in size and phenocrysts of potassium feldspar up to 1 inch or more across. The phenocrysts contain scattered inclusions of biotite, plagioclase, and quartz. Grain size of the groundmass decreases to the north. Although the granite is very fresh and apparently unweathered, the effects of possible deuteric alteration can be seen in thin section. The plagioclase (sodic andesine) is largely altered to white mica, and the potassium feldspar is altered to white mica in places. The epidote is also probably a deuteric mineral. Zircon is an abundant accessory mineral that occurs in two different habits, tiny euhedral prisms and larger, more abundant, irregularly-shaped grains up to 1 millimeter in size. A modal analysis of a sample of granite from the southern end of the Marble Mountains (Plate 8) is presented in Table 15 (MM-3).

Highly reacted inclusions of more mafic rock are abundant in the granite, particularly at the southern end of the range. Thin pegmatitic stringers and narrow quartz veins are scattered through the granite. The pegmatites are generally less than 6 inches thick, but a few pegmatites up to 3 feet thick were seen. They contain quartz and pink potassium feldspar, but no muscovite-bearing pegmatite was seen. North of U. S. Highway 66 the pegmatites intrude the granodiorite described previously. The pegmatites are considered to be genetically related to the granite and to have formed during the late stages of crystallization of the granite mass.

At the southern end of the Marble Mountains the granite is cut by a few thin mafic dikes. The dikes trend east-west and, from a distance, they seem to be truncated by the unconformity at the base of the Prospect Mountain quartzite. However, the dikes only change directions at the contact and have been emplaced in the unconformity itself for distances of 25 to 50 feet before again changing directions to intrude the Prospect Mountain quartzite. The dikes are probably related to the Tertiary volcanic activity in the region that, north of U. S. Highway 66, is evidenced by rhyolitic tuffs and basalt flows.

The absolute ages of three minerals from the Marble Mountains were measured and the results are presented in Table 16. Biotite and potassium feldspar were separated from the granite at the south end of the range (MM-3); the sample locality is approximately 1.15 miles N.35°E. of Cadiz (Plate 8). The biotite concentrate (MM-3b) contains biotite 97.5 percent, hornblende 1.9 percent, and feldspar 0.6 percent. The potassium feldspar concentrate (MM-3f) contains no observable biotite impurities, and quartz plus plagioclase is less than 1 percent. Biotite (MM-1) was separated from the hybrid rock of a roof pendant; the sample locality is approximately 0.7 miles S.5°W. of Cadiz Summit (Plate 8). The biotite concentrate contains biotite 82.9 percent, chlorite 10.6 percent, hornblende 1.5 percent, quartz 4.5 percent, and magnetite 0.5 percent.

The biotite from the granite (MM-3b) has concordant K-Ar and Rb-Sr ages of approximately 1200 million years. The coexisting potassium feldspar (MM-3f) has a Rb-Sr age of 1370 million years by direct analysis and 1340 million years utilizing an isotope ratio analysis of the strontium. The K-Ar age of MM-3f is 19 percent less than the K-Ar age of MM-3b. It has commonly been observed that the K-Ar age of mica is 25 to 35 percent greater than the K-Ar

Table 16. Analytical data and radioactive ages for minerals from the Marble Mountains, California

sample	A^{40*}(cc STP/gm)	A^{40*}/A^{40}	percent K	K-A(m.y.)	Sr^{87*}(ppm)	Sr^{87*}/Sr^{87}	Rb(ppm)	Rb-Sr(m.y.)
MM-1	0.306×10^{-3}	0.99	4.31	1250	2.81_4	0.52	540	1310
MM-3b	0.431×10^{-3}	0.99	6.44		4.71_7	0.63	968	1230
	0.425×10^{-3}	0.95			4.82_6	0.64	991	1220
	avg: 0.428×10^{-3}			1190				
MM-3f	0.577×10^{-3}	0.96	11.36	970	2.91_8	0.12	536	1370
					2.84_9(a)			1340

MM-1 : biotite-roof pendant
MM-3b : biotite - granite
MM-3f : potassium feldspar - granite

(a) - isotope ratio analysis

age of coexisting potassium feldspar (Wetherill and others, 1955; Wasserburg and others, 1956); this difference has been considered the result of loss of radiogenic argon from the feldspar. Because of presumed argon loss, MM-3f does not date a geologic event, but its apparent K-Ar age of 970 million years is important as it implies that the granite has not been significantly heated in the past billion years.

Professor L. T. Silver has measured Pb-U ages on zircon from a sample of granite collected approximately 1,500 feet east of MM-3 (Plate 8). He has analyzed three portions of the zircon concentrate retained on a 200 mesh screen: a portion of the total concentrate (A), a less radioactive portion of the concentrate (B), and a more radioactive portion of the concentrate (C). He very generously permitted the use of his data in this thesis for purposes of comparison with other ages. His data are as follows:

	apparent ages (m.y.)		
	Pb^{206}/U^{238}	Pb^{207}/U^{235}	Pb^{207}/Pb^{206}
A	760	920	1320
B	1135	1235	1400
C	525	680	1245

The ages are discordant. When the Pb/U atom ratios are plotted on a Pb^{207}/U^{235}-Pb^{206}/U^{238} diagram, the three points fall on a straight line that intersects ''concordia'' at ages of 1450 and slightly less than 200 million years. From the model of Wetherill (1956) these data can be interpreted to indicate crystallization at 1450 million years ago and episodic loss of lead slightly less than 200 million years ago.

The Rb-Sr age of the potassium feldspar (MM-3f) is in reasonably good agreement with the episodic loss model zircon age obtained by Silver, the difference in ages being approximately 7 percent. The discrepancy in apparent age between the Rb-Sr age of the potassium feldspar and the concordant ages of the coexisting and presumably cogenetic biotite is approximately 10 percent, which is considered outside of experimental error. The discordance is not understood at present, but it is clear that the discrepancy must reflect the history of the rock since its time of crystallization. Other workers have found that biotite is less resistant to recrystallization and consequent loss of radiogenic daughter whereas potassium feldspar in some cases is more resistant to loss of strontium than is biotite to loss of either argon or strontium.

MM-1 from the roof pendant has slightly discordant ages that agree within experimental error with the ages of MM-3b, though the ages of MM-1 are about 5 percent higher. Although the roof pendant is certainly older than the granite, the difference in age is not known. The ages of MM-1 must be considered minimal because they are less than the Rb-Sr age of potassium feldspar from the granite. It seems likely that any memory of an older age would have been erased during reaction between the roof pendant and the granite.

It is concluded that the granite of the Marble Mountains was emplaced as an intrusive rock approximately 1350 million years ago. The discordance between the Rb-Sr age of MM-3f and the concordant ages of MM-3b is interpreted as indicating the partial recrystallization of the biotite at some later time.

The data place an upper limit of 1350 million years on the age of the lower Cambrian. It is apparent that these methods are not particularly valuable in defining the age of the lower Cambrian in the Marble Mountains.

THE WILDROSE AREA

It has been shown in Part I of this report on the basis of stratigraphic evidence that rocks of early Precambrian, later Precambrian, and Precambrian (?) age are present in the central Panamint Range (Plate 7). In the Wildrose area of the Panamint Range K-Ar ages were measured on four metamorphic rocks and one igneous rock. On stratigraphic grounds three of the metamorphic rocks are early Precambrian in age and one is Precambrian (?) in age. The intrusive rock is post-later Precambrian in age on stratigraphic evidence, but it is surmised on the basis of regional geology that the igneous rock is possibly Cretaceous in age. Sample descriptions and estimated modes are given in Part I.

W-10 and W-15 are from the lower metamorphic series of the Panamint complex and W-18 is from the upper metamorphic series. A high-purity muscovite concentrate was separated from W-10, a micaceous quartzite. Very fresh muscovite flakes, without any visible signs of alteration, were obtained in the -100+150 mesh fraction. Very fine-grained magnetite is scattered through the muscovite flakes. The concentrate contains less than 1 percent quartz. Coarser muscovite occurs in other outcrops of micaceous quartzite in the vicinity of W-10, but the books of muscovite are commonly discolored by hematitic stains and frayed along their edges.

Hornblende was concentrated from W-15, an amphibolite. The outcrop is on the western front of the Panamint Range, and although the locality is adjacent to the Panamint Valley fault zone, no cataclastic effects are visible in either hand specimen or thin section. A hornblende concentrate was obtained in the -80+100 mesh fraction. The hornblende shows no visible alteration, but the plagioclase in the amphibolite is altered in patches to fine-grained white mica.

Biotite was concentrated from W-18, a biotite-quartz schist. The concentrate contains biotite 78.5 percent, quartz plus plagioclase 17 percent, magnetite 4 percent, and muscovite 0.5 percent. Very fresh biotite was obtained in the -80+100 mesh fraction. The biotite is pleochroic from pale brown to deep brown; some layers are altered to chlorite, but the chlorite is less than half a percent of the concentrate.

A high-purity biotite concentrate was separated from W-32, laminated micaceous limestone from the middle part of the Precambrian (?) Noonday dolomite. The biotite is pleochroic from colorless to pale brown. It occurs, together with a minor amount of muscovite, in mica-rich quartzose layers that weather out in relief. The less resistant interbeds contain very little mica. The biotite was concentrated from the entire rock; the concentrate contains less than 5 percent calcite.

Coarse muscovite, larger than +9 mesh, was hand picked from W-40, a dike of coarse-grained granite that intrudes the lower part of the Surprise member of the Kingston Peak formation. Although some grains were hematite stained a concentrate of very fresh muscovite plates about 1/8 to 1/4 inch on a side was obtained.

The early Precambrian rocks used in this study were selected to give a sampling over as wide a stratigraphic range and areal coverage as possible. Three different minerals were utilized in order to compare the ages obtained on different mineral species. It was discovered during the field study of the Wildrose area that the Precambrian (?) rocks, the Noonday dolomite and the Johnnie formation, are metamorphosed, which is in marked contrast to their occurrence as unmetamorphosed sedimentary rocks in the southern Panamint Range and east of Death Valley. In the Wildrose area the Precambrian (?) formations and the older rocks have similar mineral assemblages that are typical of middle grade metamorphism. Therefore, biotite from the Noonday dolomite was dated in order to compare its age with those of the older rocks. Finally, a date was obtained on the granitic rock in order to determine its relation to the ages of the metamorphic rocks.

Mineral concentrates from the metamorphic rocks have unfavorable Rb/Sr ratios and the Rb-Sr method cannot be utilized. K-Ar ages of the metamorphic rocks and igneous rock are presented in Table 17. The data indicate that the Wildrose area was involved in regional metamorphism during the later part of the Cretaceous period that resulted in complete recrystallization of the minerals. With the exception of W-18 the metamorphic rocks have K-Ar ages of 75 to 85 million years which are interpreted as the age of regional metamorphism in the area. The field relations indicate that the granitic rocks were intruded after the episode of regional metamorphism (see section on metamorphism, Part I of this report), and the K-Ar age of 73 million years is consistent with the field relationships.

Table 17. Analytical data and radioactive ages for minerals from the Wildrose area, California

sample	A^{40*} (cc STP/gm)	A^{40*}/A^{40}	percent K	K-A(m.y.)
W-10	2.69×10^{-5}	0.72	8.19[a]	80
			8.27	
			avg: 8.23	
W-18	2.45×10^{-5}	0.77	5.19[a]	115
			5.20	
	2.16×10^{-5}	0.89	avg: 5.20	101
W-15	0.117×10^{-5}	0.23	0.43	67 ± 10
W-32	1.91×10^{-5}	0.76	5.95	79
W-40	2.62×10^{-5}	0.81	8.87	73

a - K analysis by T. Wen

W-10 : muscovite - micaceous quartzite
W-18 : biotite - biotite-quartz schist
W-15 : hornblende - amphibolite
W-32 : biotite - Noonday dolomite
W-40 : muscovite - Cretaceous granite

The agreement in the dates of 101 and 115 million years for W-18 is not good, but it is possible that the low result represents incomplete fusion of the sample. In any case the results on W-18 indicate an apparent age that is more than 25 percent older than the 75 to 85 million year group. Biotite in metamorphic rocks from other areas has shown this type of discrepancy. In the metamorphic terrane of the southern Appalachians, biotite ages have been measured that are about 20 percent older than the apparent age of metamorphism (Long, Kulp, and Eckelmann, 1959). In the contact zone of a Tertiary intrusive in Colorado, Hart (1961) has found discrepancies in apparent K-Ar age of 35 percent between different samples of biotite. Two mechanisms that can explain the discrepant ages are partial recrystallization of the biotite and inheritance of radiogenic argon by the biotite. Because biotite seems less resistant to metamorphism than hornblende (Hart, 1961) and also than muscovite, the latter mechanism may be more likely in considering the results on W-18. However, the writer can only conclude that the data on the metamorphic rocks indicate a clear-cut Cretaceous age, and that there is no memory whatsoever of a Precambrian age.

The age of W-40 agrees with the K-Ar ages, ranging from 77 to 95 million years, obtained on various plutons in the Sierra Nevada and Coast Ranges (Curtis and others, 1958). Isotopic ages have been measured on portions of the southern California batholith that also indicate extensive emplacement of igneous rocks during the Cretaceous period. A Pb^{206}/U^{238} age of 103 ± 2 million years has been obtained on zircon from the Rubidoux Mountain granite (Banks and Silver, 1961). A K-Ar age of 92 ± 2.5 million years

(Reynolds, 1957) and a Rb-Sr age of 110 million years (Aldrich and others, 1956) have been measured on pegmatites near Pala.

There is evidence of a younger event in the Death Valley region. Wasserburg and others (1959, pp. 706-707) determined a K-Ar age of 30 million years for coarse-grained muscovite from a pegmatite in Monarch Canyon in the Funeral Mountains (Plate 7). The extent of this younger event is certainly unknown at present.

Several Pb-α ages, ranging from 92 to 131 million years, have been measured on zircons from plutons of the Sierra Nevada and southern California batholiths (Larson and others, 1958). The ages are consistent with the stratigraphic limits of intrusion, but because of the inherent uncertainties of analytical accuracy and the common lead correction these Pb-α ages are difficult to evaluate.

The Wildrose data have an interesting implication to the general study of geochronology as here is a case in which Cretaceous mineral ages are obtained on rocks that are early Precambrian in age on a stratigraphic basis. The study of the stratigraphy and structure of the area indicates a polymetamorphic history for the early Precambrian rocks. However, in any fresh, unaltered middle grade metamorphic rock obtained from the Panamint complex there is no evidence in either the mineral assemblages or texture which indicates that the rock has been involved in more than one period of metamorphism. It could be predicted that the Precambrian age of the rocks had been altered after it was observed that the Noonday dolomite and younger rocks were metamorphosed. Assuming complete recrystallization of the minerals and consequently no memory of the Precambrian age, a Cretaceous age for the minerals could be predicted on the basis of the known geologic history of the region. However, it is important to realize that this deductive reasoning is possible only because the stratigraphic age of the rocks is known to be Precambrian. Many areas of crystalline rocks occur as isolated blocks without any stratigraphic evidence of their age. If the early Precambrian rocks of the Wildrose area had been such an isolated block, it would not be possible, on the basis of the radioactive ages that were obtained, to prove the Precambrian age of the rocks.

Although all pre-Cretaceous rocks in the area indicate the same grade of metamorphism, the degree of deformation is quite different in the Precambrian (?) and later Precambrian rocks than it is in the early Precambrian Panamint complex. The later Precambrian and Precambrian (?) formations dip monoclinally to the east or in places are thrown into rather open folds. Rocks of the Panamint complex have been highly deformed, and the structural evidence on intensity of deformation suggests that metamorphism must have accompanied the earlier periods of deformation. Thus, the structural evidence implies that the early Precambrian rocks were completely recrystallized in the Cretaceous period without again being severely deformed. Therefore, the Wildrose area is an example of recrystallization where any memory of an older absolute age has been erased but the original fabric of the rock has not been altered by mobilization or reconstitution of the minerals.

It is possible that zircons from the metamorphic rocks might show a memory of a Precambrian age, and this possibility should be investigated. Zircon is present in W-10 and W-18, but in both rocks its abundance is less than 0.1 percent.

JOSHUA TREE NATIONAL MONUMENT

Crystalline rocks of unknown age form the basement of the Joshua Tree National Monument in the vicinity of Twenty-Nine Palms, California (Plate 7). All the geologic units in the area were defined by Miller (1938) who considered them Precambrian in age except for the White Tank quartz monzonite, which he assigned a late Jurassic age. Miller assigned a Precambrian age to these units on the bases of degree of metamorphism and lithologic similarities to other areas; however, no stratigraphic evidence of the age of the rocks has been found. The White Tank quartz monzonite intrudes all other units in the area. It is similar to and has been correlated with the plutonic rocks of the Sierra Nevada batholith. The writer utilized existing geologic maps (Miller, 1938; Rogers, 1954) for sampling purposes.

The oldest unit in the area is the Pinto gneiss which has been intruded by the Gold Park gabbro-diorite, the Palms granite, and the White Tank quartz monzonite. A sample of the Pinto gneiss, JT-4, was collected from the SW$\frac{1}{4}$, SW$\frac{1}{4}$, sec. 35, T. 1 S., R. 9 E. The gneiss is medium grained and has a good foliation. In outcrop it is massive and breaks into large blocks. The gneiss is very fresh and shows no evidence of weathering in either hand specimen or thin section. JT-4 has the following estimated modal composition: quartz 45 percent, potassium feldspar 30 percent, plagioclase 10 percent, biotite 15 percent, and trace amounts of muscovite, zircon, and apatite. A biotite concentrate was obtained from the -35+80 mesh fraction; the concentrate contains biotite 97.4 percent, muscovite 0.4 percent, quartz plus feldspar 1.8 percent, and zircon 0.4 percent. The biotite in JT-4 is concentrated in poorly defined layers, and the parallel preferred orientation of biotite plates is only fair. The biotite is pleochroic from light to very dark brown, and it is not chloritized or otherwise altered. Grains of zircon and pleochroic halos are scattered through the biotite flakes, but the abundance of zircon in the rock is less than 0.1 percent.

A K-Ar age of 83 million years was measured on the biotite JT-4 (Table 18). The fabric and mineral assemblage of JT-4 indicate that it is a regional metamorphic rock; but as the sample locality is only about three quarters of a mile from the nearest outcrop of the White Tank quartz monzonite, the possibility of contact effects cannot be ignored. However, in other areas of southern California in which contact metamorphism can be ruled out, K-Ar ages of regional metamorphic rocks have been obtained that range from 75 to 85 million years (Table 17).

The age of JT-4 establishes a maximum age for the White Tank quartz monzonite. If it is assumed that JT-4 was recrystallized in the contact zone of the White Tank stock and that the biotite has not inherited radiogenic argon, then the White Tank quartz monzonite must be 83 million years old or slightly younger. This age confirms the correlation of the White Tank quartz monzonite with plutonic rocks of the Sierra Nevada batholith.

Table 18. Analytical data and radioactive ages for minerals from the Joshua Tree National Monument, California, and the Kilbeck Hills, California

sample	A^{40*}(cc STP/gm)	A^{40*}/A^{40}	percent K	K-A(m.y.)	Sr^{87*}(ppm)	Sr^{87*}/Sr^{87}	Rb(ppm)	Rb-Sr(m.y.)
JT-4	2.26×10^{-5}	0.83	6.71	83				
MM-34	1.13×10^{-5}	0.56	7.16	39	0.255	0.12	852	76
					0.282 (a)			84

JT-4 : biotite - gneiss, Joshua Tree National Monument
MM-34 : biotite - gneiss, Kilbeck Hills

(a) isotope ratio analysis

THE KILBECK HILLS

The Kilbeck Hills, located about 16 miles southeast of Cadiz, California (Plate 7), are underlain by a metamorphic complex that has been assigned an "Archean" or early Precambrian age (Hazzard and Dosch, 1936). About 30 miles northeast of the Kilbeck Hills in the Piute Mountains, the Essex series, which is the oldest major unit of the metamorphic complex, is unconformably overlain by a quartzite of possible lower Cambrian age (Hazzard and Dosch, 1936, p. 309). However, no stratigraphic evidence of the age of the Essex series has been found in the Kilbeck Hills.

The writer collected a sample of quartz-feldspar-biotite gneiss from the upper part of the Essex series at 34°22' north latitude and 115°17'30" west longitude. The gneiss, MM-34, is presumably a metasedimentary rock as it is interbedded with marble. The gneiss is fine grained and does not have a good foliation. It breaks into large, rather irregular blocks whose outer surfaces are heavily varnished. Beneath a 1/4 inch weathered rim the gneiss is very fresh and apparently unweathered. In thin section the minerals are very fresh; an estimated modal composition of the gneiss indicates quartz 50 percent, potassium feldspar 28 percent, biotite 10 percent, plagioclase 9 percent, magnetite 2 percent, muscovite 0.5 percent, sphene 0.5 percent, and trace amounts of chlorite, zircon, and apatite. A biotite concentrate was obtained in the -80+120 mesh fraction; the concentrate contains biotite 99.6 percent, quartz 0.2 percent, chlorite 0.2 percent, and a trace amount of zircon. The biotite is pleochroic from pale to medium brown. Scattered layers of the biotite are chloritized, but chlorite makes up less than 1 percent of the biotite concentrate.

The experimental results for MM-34 biotite are presented in Table 18. The uncertainty in A^{40*} is probably 25 to 50 percent because the argon was extracted from a small sample and the corrections for air contamination and A^{40} in the tracer amounted to more than 60 percent of the total A^{40}. Because of the experimental uncertainty the K-Ar date of 39 million years cannot be related to a specific orogenic event and should be considered a minimum age. The most precise Rb-Sr age of MM-34 biotite is 84 million years, which is in good agreement with the 83 million K-Ar age of the Pinto gneiss. The data indicate that the biotite has been a closed system to rubidium and strontium only since the Cretaceous period.

A total rock rubidium-strontium analysis was made on MM-34 in order to determine if the gneiss had retained any memory of an older age. A 2 inch x 2 inch x 1 inch piece of gneiss was crushed to pass a -80 mesh screen. The measured Sr^{87}/Sr^{88} ratio of MM-34 total rock is 0.0858 ± 0.0003. The quality of the data is good and the mean deviation is considered a good estimate of the error. In order to correct for discrimination, the Sr^{86}/Sr^{88} ratio of 0.1210 ± 0.0003 was normalized to a value of 0.1186. The corrected Sr^{87}/Sr^{88} ratio is 0.0849, which indicates an enrichment in Sr^{87} of approximately 1 percent. The total rock sample contains 145 ppm rubidium and 165 ppm strontium as measured by X-ray fluorescence. Thus, if the rock is 1 billion years old the strontium will be slightly more than 4.5 percent radiogenic. However, it has been our experience that the rubidium content of biotite determined by

X-ray fluorescence is invariably low. For MM-34 biotite the rubidium content is 852 ppm by isotope dilution compared with 320 ppm by X-ray fluorescence. Because biotite contains most of the rubidium in the rock, the X-ray fluorescence value of 145 ppm rubidium for MM-34 total rock is low, possibly by a factor of about 2. Therefore, the strontium in MM-34 total rock should be 8 to 10 percent radiogenic if the rock is a billion years old. Thus, on the basis of its measured strontium composition, the apparent age of MM-34 total rock is approximately 100 million years. This result is consistent with either one of two models, which are (1) the Essex series is Cretaceous in age, or (2) the Essex series is Precambrian in age but during Cretaceous metamorphism the system, as defined by the total rock sample, was completely open to strontium.

REGIONAL RELATIONS OF THE EARLY PRECAMBRIAN ROCKS

In line with the principal objective of this investigation, correlations of the early Precambrian rocks between certain areas are suggested. These correlations are made utilizing both the absolute ages and lithologic characteristics of the rocks.

It is necessary to define the term "orogenic event" before correlations can be made on the basis of absolute ages. The definition of an event requires a coherence of data obtained from both geological relationships and mineral ages. It is well known that at times any single dating method gives results that are discordant with those obtained by other methods, and that the ages of cogenetic minerals are commonly discordant. Therefore, a reliable estimate of the age of an event requires dates obtained by as many different methods as possible on different minerals. If the dates are concordant and agree with the geological evidence, an event can be defined without any trouble. If the ages are discordant the pattern of the data may make it possible to set limits on the event. For example, it has been shown that for biotite the K-Ar age, the Rb-Sr age, or both ages are commonly low and that the Rb-Sr age of potassium feldspar is the best estimate of the true age of the rock. Finally, an orogenic event should be regional in extent. Thus, similar mineral ages obtained over a large area are strong evidence of a discrete event.

The oldest rocks in the region are probably more than 1800 million years old. The lead-uranium ages of zircons from meta-igneous rocks in the Southern Panamint Range yield Pb^{207}/Pb^{206} ages of 1720 ± 20 and 1780 ± 20 million years (Silver and others, 1961). The isotopic ages are discordant and the Pb-Pb ages are considered to be minimum ages; the prior history of the meta-igneous rocks as evidenced by the discordant ages of the zircons remains to be investigated. The meta-igneous rocks and associated metasedimentary rocks have been intruded by pegmatites that have concordant K-Ar and Rb-Sr ages of approximately 1700 million years (Wasserburg and others, 1959). The data have been interpreted to indicate an event of profound metamorphism and igneous intrusion about 1700 million years ago.

The metamorphic rocks and associated granitic pegmatites in the Mountain Pass district also indicate this same event. Although the ages measured at Mountain Pass are slightly younger, they are within 3 percent of those in the southern Panamint Range.

The metasedimentary rocks in the southern Panamint Range have concordant K-Ar and Rb-Sr ages of approximately 1500 million years, which is 14 percent younger than the age of the pegmatite that is intrusive into the schists (Wasserburg and others, 1959). The agreement of the ages of the schist in the southern Panamint Range with the Rb-Sr age of biotite from a granite gneiss in the southern Nopah Range has been suggested as possibly indicating a single event at about 1500 million years (Wasserburg and others, 1959, p. 707).

It is believed that the granite of the Marble Mountains was emplaced approximately 1350 million years ago as indicated by the Rb-Sr age of potassium feldspar. From the Pb-U ages obtained by Silver, the episodic lead loss model suggests intrusion of the Marble Mountains granite approximately 1450 million years ago. In the Mountain Pass district, an average Rb-Sr age of 1430 $\pm$ 40 million years was obtained on biotite from the shonkinite but the K-Ar ages were much lower; recalculation of the Pb^{208}/Th^{232} age of monazite from the carbonate mass (Jaffe, 1955; Tilton, written communication, 1961) yields an age of 1425 $\pm$ 70 million years. These data suggest a period of igneous intrusion in the 1350 to 1450 million year interval, though the dates obtained at Mountain Pass should be considered minimum ages. On the basis of the 1500 million year ages obtained on metamorphic rocks described above, the region was apparently affected by metamorphism as well as igneous intrusion between 1350 and 1500 million years ago, though the data on the metamorphic rocks is more suggestive than positive.

In the western United States a period of igneous intrusion approximately 1350 million years ago has been suggested (Aldrich, Wetherill, and Davis, 1957). Micas from eleven different pegmatites and granitic rocks in Arizona, New Mexico, Colorado, and Wyoming yield an average K-Ar age of 1340 ± 50 million years with a spread of 1160 to 1420 million years and an average Rb-Sr age of 1380 ± 50 million years with a spread of 1300 to 1500 million years. The Rb-Sr age of potassium feldspar from the granite of the Marble Mountains is in good agreement with these ages and suggests a westward extension into California of the 1350 million-year old granitic rocks. The Lawler Peak granite at Bagdad, Arizona, is the only one of the eleven rocks in which a coexisting zircon was analyzed. The ages for muscovite and zircon are as follows (Aldrich and others, 1958, p. 1,129):

ages (m.y.)

	K-A	Rb-Sr	$\frac{Pb^{206}}{U^{238}}$	$\frac{Pb^{207}}{U^{235}}$	$\frac{Pb^{207}}{Pb^{206}}$	$\frac{Pb^{208}}{Th^{232}}$
muscovite	1360	1390	—	—	—	—
zircon	—	—	630	770	1210	270

The zircon shows the same pattern of discordance as the zircon from the Marble Mountains, but the meaning of the discordance at Bagdad will require additional work.

Giletti and Damon (1961) report additional Rb-Sr ages from Arizona that, in part, confirm the results of Aldrich and others (1957). Giletti and Damon measured six Rb-Sr ages of micas that range from 1210 to 1550 million years and they concluded that these data "extend the 1350 million year orogeny to northwestern and southern Arizona" (1961, p. 639). This conclusion is not consistent either with their estimated error of ± 3 percent for each age (Giletti and Damon, 1961, p. 640) or with the definition of an orogenic event used by the present writer. The six rocks utilized by Giletti and Damon belong to two different geologic settings. A pegmatite and migmatitic zone in the Vishnu schist, Grand Canyon, represent a much different environment than the four "granites" from northwestern and central Arizona. The 1550 million year Rb-Sr age of muscovite from a pegmatite and the 1390 million year Rb-Sr age of biotite from a migmatitic zone in the Vishnu schist illustrate the same type of discordance observed between the pegmatite and metamorphic rocks at Mountain Pass. The pegmatites in the Vishnu schist and at Mountain Pass may be correlative though the difference in ages is about 6 percent. Biotite from the four "granites" have Rb-Sr ages of 1210, 1300, 1350, and 1450 million years (Giletti and Damon, 1961, p. 640). Assuming loss of strontium from the biotites that have low apparent ages, these results agree with the 1350 million year old ages measured by Aldrich and others (1957). Therefore, it is believed that the data of Giletti and Damon suggest the same two early Precambrian events that are recognized in southern California, not just the "1350 million year orogenic event."

No compelling evidence of a Precambrian event younger than 1350 million years has been found in the Death Valley-Mojave Desert region. The younger K-Ar ages of certain biotites, MP-21, MP-22, MM-3b, and biotite from the southern Nopah Range (Wasserburg and others, 1959), can all be interpreted as due to argon loss at a later, but unknown time. Evidence for a younger event has been found in the San Gabriel Mountains, which are west of the Mojave Desert. Silver and others (1960) obtained concordant Pb-U ages of 1200 million years on zircon from a pegmatite that may be genetically related to a gabbro-anorthosite complex. Slightly discordant ages of 1300 to 1400 million years have been obtained on zircon from a gneiss that the gabbro-anorthosite complex may intrude.

The dates ranging from approximately 70 to 120 million years, obtained by several workers on various minerals, have been ascribed to metamorphism and intrusion during the Cretaceous Nevadan orogeny. Age measurements in the Wildrose area indicate that early Precambrian rocks were completely recrystallized 75 to 85 million years ago. Similar ages of metamorphic rocks are obtained at Joshua Tree National Monument and the Kilbeck Hills. But because stratigraphic control is lacking in the latter areas, the mineral ages can be interpreted either as metamorphic rocks of Cretaceous age or as older rocks that were metamorphosed in the Cretaceous period. The dates obtained on plutonic rocks suggest an extended period of orogeny. However, the zircon ages are commonly several percent higher than K-Ar ages of micas, and in one case the Rb-Sr and K-Ar ages of a pegmatite are badly discordant. Therefore, additional work utilizing different dating methods will be required to delineate the limits of the Nevadan orogeny.

Certain generalizations may be made about the early Precambrian rocks of the Death Valley-Mojave Desert region. The region may be divided into two provinces that are contrasted by general differences in lithology though the differences are not striking. In the "eastern province," east of Death Valley and south of the Black Mountains, a basement of granitic and dioritic gneisses and later, seemingly unmetamorphosed, granitic intrusive rocks is typical. In the "western province," the Avawatz, Owlshead, Quail, and Funeral Mountains and the Panamint Range, the basement consists of schist and phyllite, marble, quartzite, amphibolite, and gneiss; early Precambrian intrusive rocks are present in some areas (Noble and Wright, 1954). If the overall composition of the early Precambrian terrane is uniform, the apparent lower grade of metamorphism in the western province could be interpreted as the result of being involved in the Cretaceous orogeny. Consequently, the pattern of Cretaceous ages observed in the Wildrose area of the Panamint Range also might be expected in other parts of the western province.

The porphyritic granite in the Providence Mountains is lithologically similar to the granite of the Marble Mountains, and the same sequence of lower Cambrian formations, the Prospect Mountain quartzite, the Latham shale, and the Chambless limestone, unconformably overlies the granite in the Providence Mountains (Hazzard, 1954). A correlation of the Precambrian rocks of these areas is very reasonable, and a similar pattern of mineral ages would be expected. A correlation between the granite of the Marble Mountains and the potassium-rich igneous rocks of the

Mountain Pass district is suggested on the basis of chemical similarity, in particular the high potassium content. The mineral age measurements are consistent with this correlation. The occurrence of pink granite intruding diorite in the Avawatz Mountains (L. A. Wright, oral communication, 1961) is similar to the relationship of the igneous rocks of the Marble Mountains. If these areas are correlative, the 1350 million year age might be expected, but the effects of the Cretaceous orogeny are an unknown factor. The granitic rocks in the Owlshead Mountains are probably the same age as those in the Avawatz Mountains.

The assignment of a Precambrian age to a granite in the eastern Mojave Desert just because it is porphyritic must be done with caution. The granite of the Marble Mountains is actually close to a quartz monzonite in composition, and porphyritic quartz monzonites are major components of the Cretaceous plutons scattered through the region.

That the outline of the zone affected by the Cretaceous orogeny is not simple becomes apparent when the mineral ages at the Kilbeck Hills are considered. Only a few miles to the north early Precambrian ages are obtained in the Marble Mountains. The distribution of Precambrian and Cretaceous mineral ages south and east of the Marble Mountains cannot be predicted because at the present time the stratigraphic age of the rocks is not certain. It is apparent that both detailed field study and additional mineral ages are required before the Essex series and associated rocks, the Pinto gneiss, and large areas of basement rocks farther south and east which have been assigned a Precambrian age can be related to the Precambrian basement of the Death Valley-Mojave Desert region.

As a result of this investigation the eastern margin of the Cretaceous Nevadan orogenic belt has been extended into the Panamint Range, west of Death Valley. East of Death Valley, two separate orogenic events are recognized in the geologic relationships and mineral ages of the early Precambrian rocks. The older event is evidenced by rocks approximately 1650 million years old; other workers in southern California and in Arizona have measured ages in this range. The younger event represents the westward extension of 1350 million year old granitic intrusive rocks in the western United States.

APPENDIX

SAMPLE PREPARATION

A 25 to 30 pound sample of rock that was as fresh and free from weathering and alteration as possible was taken from the outcrop. The large sample was broken into smaller fragments that were crushed in a Bico-Braun chipmunk crusher and, if further comminution was necessary, pulverized in a Bico-Braun pulverizer. The material was sized in a sequence of brass wire cloth screens. The coarsest fraction in which individual grains occur, rather than binary aggregates, was selected for further concentration. Mica samples were concentrated in the magnetic fraction of a Frantz isodynamic separator. When biotite or muscovite was the only magnetic mineral it was generally possible to get a 95 percent concentrate or better with the Frantz separator. Potassium feldspar was concentrated by passing the non-magnetic fraction from the Frantz separator through a tetrabromoethane-acetone mixture of specific gravity 2.565; quartz, plagioclase, and heavier minerals sank and potassium feldspar floated.

In order to insure as uniform a sample as possible the mineral concentrate was split by quartering. Between 2 grams and 5 grams, depending on the age of the mineral and its potassium content, were used for an argon extraction. About 0.5 grams of sample were used for a potassium determination and 0.3 to 0.6 grams for a rubidium-strontium analysis.

One total rock rubidium-strontium analysis was done. The total rock sample consisted of a piece of rock about 2 inches x 2 inches x 1 inch that was crushed in a steel mortar to pass a 80 mesh screen. The entire sample was quartered to get about 0.5 grams for the analysis.

THEORY

In order to utilize long-lived natural radioactive isotopes for the absolute age of minerals the following assumptions are made:

(1). The mineral has been a closed system with respect to the parent and daughter isotopes since its crystallization. For igneous rocks the age is interpreted as the time since the crystallization of a mineral from a magma and in metamorphic rocks it is interpreted as the time since the mineral was last recrystallized during a metamorphic episode.

(2). The concentrations of parent and daughter isotopes in the mineral can be determined experimentally, and the amount of daughter isotope initially present can be measured and corrected for.

The basic relationship utilizing in geologic dating, the so-called age equation, is as follows:

$$t = \frac{1}{\lambda} \ln \left(1 + \frac{N_2 - N_2^{o}}{N_1}\right) \qquad (1)$$

where t = the age of the mineral

λ = the decay constant of the parent isotope

$N_2 - N_2^{o}$ = the increase in the number of atoms of radiogenic daughter between the time the system became closed and the present.

N_1 = the present number of atoms of radioactive parent

The age equation is used in this form for the rubidium-strontium method in which $N_2 - N_2^o = Sr^{87*}$ and $N_1 = Rb^{87}$. In the potassium-argon method the fact that K^{40} decays to produce both A^{40} by electron capture and Ca^{40} by beta decay must be taken into account. If the decay constant to A^{40} is λ_e and that to Ca^{40} is λ_β, the age equation is as follows:

$$t = \frac{1}{\lambda_e + \lambda_\beta} \ln \left[1 + \left(\frac{\lambda_e + \lambda_\beta}{\lambda_e} \quad \frac{A^{40*}}{K^{40}} \right) \right] \qquad (2)$$

The decay constants for the systems of geologic interest have been reviewed by Aldrich and Wetherill (1958). The following decay constants are used throughout this report.

K^{40}:	$\lambda_e = 5.85 \times 10^{-11}$ year^{-1}	(Wetherill, 1957)
	$\lambda_\beta = 4.72 \times 10^{-10}$ year^{-1}	(Aldrich and Wetherill, 1958)
Rb^{87}:	$\lambda = 1.39 \times 10^{-11}$ year^{-1}	(Aldrich and others, 1956)

The decay constant of Rb^{87} is the geologically determined one that was found by comparing Rb-Sr ages of minerals with concordant U-Pb ages of coexisting uraninites. It yields a half-life of $(50.0 \pm 2.0) \times 10^9$ years compared with the most recent laboratory value of $(47.0 \pm 0.5) \times 10^9$ years as determined by liquid scintillation counting(F lynn and Glendenin, 1959). The difference in these measurements of the Rb^{87} half-life is 6 percent and will result in a 6 percent difference in a Rb-Sr age. It is believed that the true value of the Rb^{87} half-life lies between the two values.

The abundance of parent isotopes used in the calculations are 1.19×10^{-4} atom percent for K^{40} (Nier, 1950b) and 27.85 atom percent for Rb^{87} (Nier, 1950a).

CALCULATIONS

Potassium-argon method

The amount of radiogenic argon (A^{40*}) is given by

$$\frac{(40*)}{(38)_T} = \frac{\begin{array}{l}(40/38)_M - (40/38)_M(38/36)_N(36/38)_T \\ -(40/38)_T + (36/38)_M(38/36)_N(40/38)_T \\ -(36/38)_M(40/36)_N + (40/36)_N(36/38)_T\end{array}}{1 - (38/36)_N(36/38)_M} \qquad (3)$$

where $(38)_T$ = number of atoms of $argon^{38}$ in the tracer

M = absolute ratio of atoms measured in mixture of sample and tracer

T = absolute ratio of atoms measured in tracer

N = absolute ratio of atoms in normal atmospheric argon

Rubidium-strontium method

A similar equation to (3) is used to calculate the amount of radiogenic strontium. Sr^{87*} is radiogenic strontium, Sr^{86} is used as the tracer isotope, and Sr^{88} is used for the normal strontium correction as follows:

$$\frac{(87)^*}{(86)_T} = \frac{\begin{array}{l}(87/86)_M-(87/86)_M(86/88)_N(88/86)_T-(87/86)_T \\ +\ (88/86)_M(86/88)_N(87/86)_T-(88/86)_M(87/88)_N \\ +\ (87/88)_N(88/86)_T\end{array}}{1-(86/88)_N(88/86)_M} \qquad (4)$$

where $(86)_T$ = number of atoms of strontium ^{86} in the tracer
N = absolute ratio of atoms in normal strontium
and M and T have the same meaning as in (3)

The amount of Rb^{87} in the sample is calculated using the following equation:

$$R = \frac{(87)_N}{(87)_T} = \frac{(85/87)_M - (85/87)_T}{(85/87)_N - (85/87)_M} \qquad (5)$$

where $(87)_N$ = number of atoms of normal Rb^{87} in the sample
$(87)_T$ = number of atoms of "spike" Rb^{87} in the tracer
N, M, and T are the ratios of measured ion currents, uncorrected for discrimination, in normal rubidium, in a mixture of sample and tracer, and in the tracer

ANALYTICAL PROCEDURE

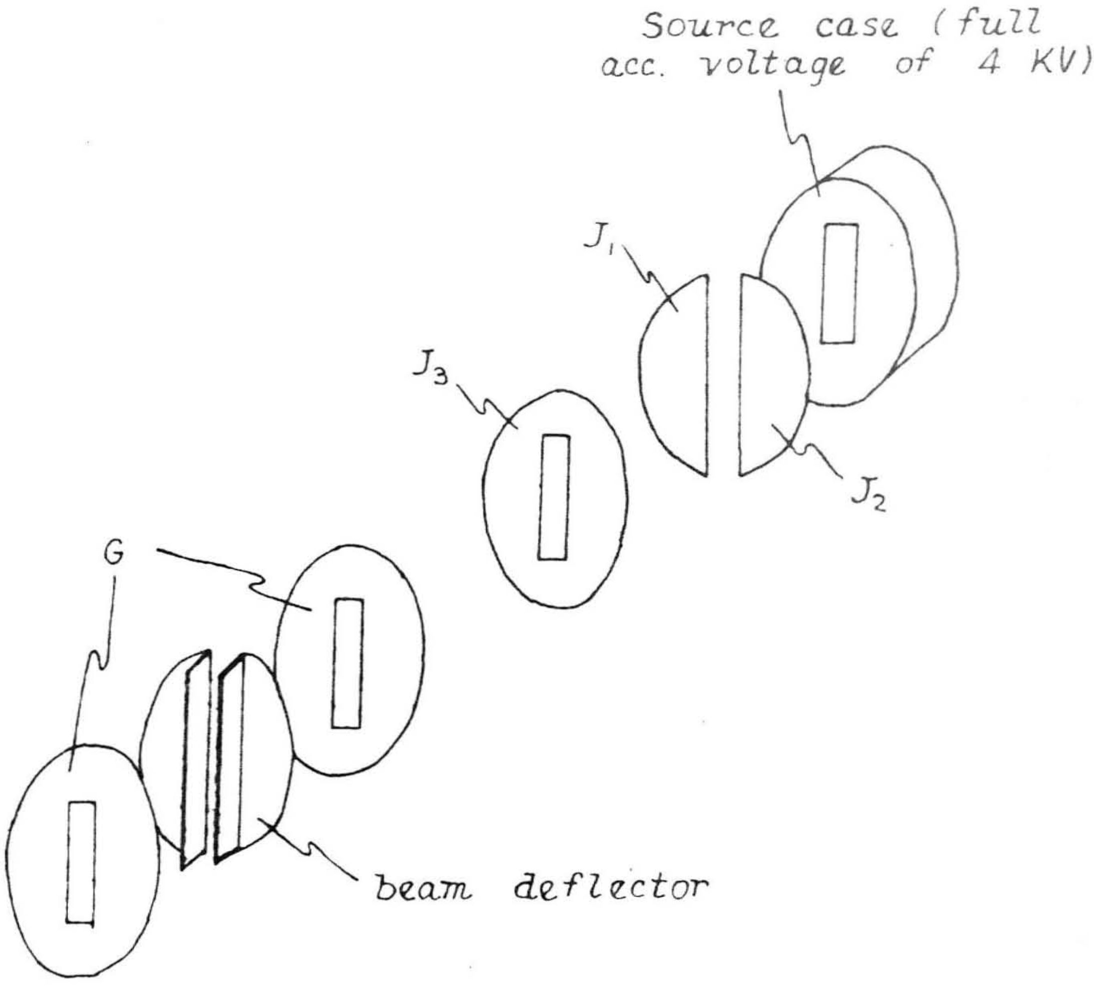
Potassium-argon method

The sample is placed in a molybdenum crucible mounted on a molybdenum rod fitted into a well in the bottom of a quartz reaction vessel. The reaction vessel is attached to the gas purification train through a steel flange which seats on a soft copper gasket. The gas purification train is built inside a marinite oven which contains a metal strip heater. After the sample crucible is loaded and a break-seal tube is attached, the oven is heated to about 200° C. and allowed to outgas overnight.

The system is tested for vacuum tightness the following day by isolating it in three parts and monitoring the pressure with a McLeod gage. If the pressure build-up over a period of 2 to 3 hours is less than 10^{-4} millimeters of mercury, an A^{38} tracer ("spike") is pipetted into the extraction line and frozen out on activated charcoal cooled with liquid nitrogen. The A^{38} tracer is held in a reservoir whose volume is 2,170 cc and is released through a gas pipette whose colume is 3.404 cc. Approximately 10^{-4} cc STP of A^{38} were contained in the initial release, and the decay constant of the reservoir is 0.157 percent per release. The A^{38} was obtained from the Oak Ridge National Laboratories and it has the following isotopic ratios: $A^{40}/A^{38} = 0.1240 \pm 0.0014$ and $A^{36}/A^{38} = 0.00124 \pm 0.00001$. The A^{38} concentration of the tracer was determined by isotope dilution with a known volume of normal argon of spectroscopic purity.

After the tracer is transferred, the argon is extracted by fusing the sample between $1{,}200^{\circ}$C. and $1{,}500^{\circ}$ C. in a 10-kilowatt induction furnace. The gas released is continuously collected on activated charcoal cooled with liquid nitrogen. No discharge was observed in the reaction vessel during an extraction. The fusion is considered complete when material begins to plate out on the walls of the reaction vessel. The gas is then thoroughly mixed with the tracer by alternately freezing out on charcoal and thawing with hot water. The mixing procedure is repeated several times, and the gas mixture is then purified by passing it over hot copper oxide and hot titanium sponge. The purified gas is collected on activated charcoal and sealed in a break-seal sample tube.

All isotopic argon analyses were made on a 6-inch, 60-degree sector, single-focusing Nier-type instrument (Nier, 1947) with a design modified by C. R. McKinney and G. J. Wasserburg. In order to obtain some Z-focusing these workers calculated trajectories of the actual magnetic field and obtained focal points for quarter-plane focusing, and they utilized a 52-degree sector magnet with 60-degree deflection. The gas is analyzed by a dynamic technique in which gas flows continuously into the mass spectrometer ion source and then is pumped away by a mercury diffusion pump in series with a mechanical forepump. A schematic diagram of the ion source is given in Figure 3. The ion beam is accelerated by a constant voltage of 4,000 volts. Mass spectra are obtained by varying the magnetic field with a resistance-capacitance circuit. Two collectors whose entry slits are 0.040 inches wide are utilized. One collector is a simple Faraday cage that is used to measure the A^{40}/A^{38} ratio of large samples. The second collector is a ten-stage Dumont copper-beryllium electron multiplier operated at 250 volts per stage from which gains of 2000 to 4000 are obtained. The electron multiplier is used to measure the A^{36}/A^{38} ratio of all samples and the A^{40}/A^{38} ratio of small samples. The resolved ion current is amplified on a vibrating reed electrometer whose output voltage is recorded on a strip chart. Each collector is attached to an Applied Physics model



J - focusing plates
G - collimating plates

Figure 3 - Schematic diagram of the ion source, 6-inch gas mass spectrometer (HENEAK)

31F vibrating reed electrometer. 10^{10} and 10^{11} ohm resistors are used on the simple collector and 10^{9} and 10^{10} ohm resistors on the electron multiplier. Masses 40 and 38 are normally measured on the simple collector and ion beams of the order of 10^{-11} to 10^{-12} amperes are general; mass 36 is normally measured on the electron multiplier and the ion beam from mass 36 is generally less than 5×10^{-15} amperes. The background ion beam at mass 36 is approximately 5×10^{-16} amperes and it generally represents about 10 to 30 percent of the total mass-36 ion beam. The samples are isolated and generally compressed slightly, and the depletion in masses 40 and 38 is about 12 percent per hour. Higher compressions are used during the mass-36 measurements and depletions of 40 to 50 percent per hour are common. The argon extractions of minerals older than 1000 million years were over 90 percent radiogenic, and the extractions of minerals younger than 100 million years were over 70 percent radiogenic except for W-15 and MM-34, which were 23 and 55 percent radiogenic respectively. The total amount of radiogenic argon in individual extractions ranged from 0.6×10^{-5} cc STP to 1.8×10^{-3} cc STP. A blank determination indicates less than 5×10^{-6} cc STP A^{40*} for the extraction-purification train.

The discrimination of the instrument was determined periodically by analyzing a sample of atmospheric or spectroscopic argon. All measured ratios were normalized to a value of $A^{40}/A^{36} = 295.6$ that was calculated from the abundances: $A^{40} = 99.600$, $A^{38} = 0.063$, and $A^{36} = 0.337$ (Nier, 1950a). The discrimination of the simple collector, as determined in several experiments, ranged from 1.1 to 1.4 percent for 2 mass units in favor of the higher mass; values between 302.1 and 304.0 were obtained for the A^{40}/A^{36} ratio. The discrimination of the electron multiplier relative to that of the simple collector ranged from 2.4 to 2.8 percent for 2 mass units in favor of the lower mass. This is approximately a square root of the mass effect.

The precision error in measuring a series of 10 to 12 isotope ratios is less than 0.3 percent. The A^{38} concentration of the spike is considered correct to within 1 percent. The reproducibility of A^{40*} for duplicate determinations is approximately 1 percent (see Table 16). The error in determining A^{40*} depends on the correction terms in (3) and is estimated to be 0.2 to 0.3 percent for minerals older than 1000 million years and 1.5 to 3.0 percent for minerals younger than 100 million years. Therefore, the total estimated error in determining A^{40*} is 1.5 to 1.6 percent for minerals older than 1000 million years and 2.8 to 4.3 percent for minerals younger than 100 million years.

The potassium was measured on a Perkin-Elmer flame photometer using an analytical method modified from Shapiro and Brannock (1956, pp. 43-44). 400 ppm lithium internal standard is used to increase sensitivity. The error in the potassium measurement is estimated to be 1 to 2 percent of the reported value except in the case of hornblende where the true value is estimated to be within 5 percent of the reported value.

The estimated error in A^{40*}/K^{40} is 2.5 to 3.6 percent for minerals older than 1000 million years and 3.8 to 6.3 percent for minerals younger than 100 million years. The estimated errors are considered maximal and it is believed that the actual errors are less. The K-Ar ages presented in Part II are considered accurate to $\pm$ 5 percent, but the error is probably closer to $\pm$ 3 percent for minerals older than 1000 million years.

The uncertainty in the half-life of K^{40} introduces an error into the accuracy of a K-Ar age that is exclusive of the errors in the measurement of the A^{40*}/K^{40} ratio.

Rubidium-strontium method

Reagents

Hydrochloric acid was made by bubbling hydrogen chloride through triple-distilled water. Reagent grade hydrofluoric and perchloric acids were used without further purification. Triple-distilled water was used in all the experiments. All glassware was cleaned by boiling it in nitric acid, and new glassware was used for each experiment. The contamination level of the reagents was determined by isotope dilution and the results obtained are as follows:

	Rb	Sr
2.5 N HCl	0.24×10^{-9} gram/gram	0.05×10^{-9} gram/gram
6 N HCl	0.047×10^{-9} "	----
$HClO_4$	0.75×10^{-9} "	8.1×10^{-9} "

Blank experiments using about twice the normal amounts of reagents indicate the following total contamination:

Rb: 0.0083×10^{-6} grams

Sr: 0.45×10^{-6} grams

Procedure

The analytical procedure is essentially that used at the Carnegie

Institution of Washington, and it was adopted by us after visiting their laboratory. The mineral sample was twice dissolved in a mixture of hydrofluoric and perchloric acids and each time evaporated to dryness. In most experiments a weighed quantity of strontium-86 tracer solution was added after the first evaporation. In the experiments in which the strontium composition was measured, the sample was made up to volume after the second evaporation and an aliquot was removed for the composition analysis before the tracer was added. The tracer solution has the composition $Sr^{88}/Sr^{86} = 0.173$, $Sr^{87}/Sr^{86} = 0.0252$, and $Sr^{84}/Sr^{86} = 0.000188$. Its concentration is 0.01554×10^{-6} moles Sr^{86} per gram of solution. After the second acid treatment and evaporation, the residue was dissolved in triple-distilled water; in some experiments it was necessary to add a few milliliters of 6N-hydrochloric acid in order to completely dissolve the residue. The sample was transferred to a weighed 150 milliliter pyrex beaker and triple-distilled water was added to make a volume of about 100 milliliters. The beaker plus sample was weighed and an aliquot, generally about 3 percent, was removed for rubidium analysis. The rubidium aliquot was weighed and to it was added a weighed quantity of rubidium-87 tracer. The tracer solution has the composition $Rb^{85}/Rb^{87} = 0.048$ and a concentration of 3.30×10^{-6} grams rubidium/gram solution. The two "spiked" solutions were evaporated to dryness.

The cation exchange columns used to separate strontium from rubidium contain 15 milliliters of Dowex 50-X8, 200-400 mesh, medium porosity, styrene type resin. Before an experiment the resin was backwashed and settled in 2.5N-hydrochloric acid and after each experiment the column was stripped with 6N-hydrochloric acid. The separation point of rubidium and strontium was determined by passing a strontium solution through the column and checking for the first appearance of strontium with a sodium rhodizonate spot test (Feigl, 1937, p. 133). The spot test was good and proved very convenient to use. However, it was necessary to increase the concentration of the strontium solution above the limit of detection stated by Feigl in order to obtain a positive test. The column cutoff was verified by mass spectrometric analysis of the strontium portion of a mineral sample; the experiment indicated complete separation of strontium from rubidium.

The entire solution, after removing the aliquot for rubidium analysis, was evaporated to dryness. The residue was dissolved in 2.5N-hydrochloric acid and centrifuged to remove insoluble particles. The supernatant liquid was decanted into the ion exchange column and it was then eluted with 65 to 75 milliliters of 2.5N-hydrochloric acid (the amount of acid varies from column to column depending on the amount of resin). The eluate which contains iron, aluminum, potassium, and rubidium was discarded. The column was eluted with 12.5 milliliters of 2.5N-hydrochloric acid and this portion, which contains the strontium, was collected in a 15 milliliter pyrex beaker and evaporated to dryness.

In order to facilitate proper spiking the approximate concentrations of rubidium and strontium in the mineral sample were determined by X-ray fluorescence. The smallest precision error in the calculations are obtained by adjusting the amounts of tracer solution to give a Rb^{85}/Rb^{87} ratio of approximately 0.5 for the rubidium mixture and a Sr^{87}/Sr^{86} ratio of approximately unity for the strontium mixture. In those experiments in which the enrichment of Sr^{87} was low the strontium composition was determined on an aliquot that was removed before the tracer was added. The total concentration of strontium rather than that of Sr^{87*} was determined on the remaining solution; consequently, the amount of tracer was adjusted to make the Sr^{88}/Sr^{86} ratio approximately unity.

The rubidium and strontium isotope ratio measurements were made on a 12-inch radius, 60-degree sector, single focusing mass spectrometer described by McKinney (Chow and McKinney, 1958). Single filaments were made of tantalum ribbon, approximately 0.030 x 0.001 inches, spot welded onto kovar pins which fit into steel filament blocks in such a way that the filaments are centered behind the first slit of the ion source as described by Inghram and Chupka (1953), which was operated to produce ions with a 5-k.e.v. energy. Mass spectra were obtained by uniformly varying the magnetic field with a servo mechanism. The resolved ion beam was passed through an adjustable slit, which was generally operated at an opening of 0.025 to 0.030 inches, to impinge on the conversion dynode of a ten-stage silver-magnesium electron multiplier that was operated to have a gain of about 900 (Chow and McKinney, 1958, p. 1500). The electron current was passed through a 2×10^{9}-ohm resistor to a vibrating reed electrometer whose output voltage was recorded on a strip chart. A signal of 100 millivolts, which is equivalent to an ion current of approximately 5×10^{-14} amperes, was obtained for rubidium at about 1 ampere filament current and for strontium at about 2.5 amperes filament current.

In order to correct for discrimination the Sr^{86}/Sr^{88} ratio of normal strontium was normalized to a value of 0.1186 (Gast, 1961, oral communication). A value of 0.0840 was used for the Sr^{87}/Sr^{88} ratio of normal strontium (Gast and Wetherill, 1961, oral communication). Several analyses of normal strontium and internal calculations from tracer calibrations indicate that the discrimination ranged from 2.02 to 2.53 percent for 2 mass units in favor of the lower mass; values of the Sr^{86}/Sr^{88} ratio ranging from 0.1210 to 0.1216 were obtained. Analyses of normal rubidium over a period of more than a year indicate a range of more than 2 percent for the Rb^{85}/Rb^{87} ratio. In this report a value of 2.610 is used for the Rb^{85}/Rb^{87} ratio; Nier's value for this ratio is 2.591 (1950a). A change in the isotope ratios was noted in some of the longer spectrometer runs. In all cases the change in the ratio was less than 0.5 percent and was in favor of the lower mass, which indicates that the effect is not due to fractionation.

The mass spectrometer ion source was modified so that a unit consisting of the defocusing and discriminating plates plus the filament block was completely removable. The unit was cleaned in hot nitric acid after each experiment, and the plates were rubbed with "wetordry paper." For a rubidium analysis a new filament was made and the filament-ion source unit was outgassed for 2 to 3 hours at about 3 amperes filament current to remove trace amounts of rubidium. The ion source was removed from the filament block and a sample was loaded by adding a few drops of 6N-hydrochloric acid to the residue and then pipetting the liquid onto the filament. The ion source was replaced on the filament block and the entire unit was loaded into the spectrometer. For a strontium determination the sample was loaded on a new filament, the cleaned ion source was attached to the filament block, and the entire unit was outgassed overnight in the mass spectrometer at 1.8 to 2.0 amperes filament current to bake out any rubidium. The filament current was increased slowly until stable strontium peaks appeared in the range 2.3 to 2.7 amperes filament current. Generally twenty to thirty scans of the 86, 87, and 88 mass positions were made for a strontium analysis and twenty to thirty scans of the 85 and 87 mass positions for a rubidium analysis. During a strontium analysis the mass 85 position was checked constantly for rubidium contamination. In some experiments the rubidium apparently came from either the filament or sample and the rubidium was normal in composition. However, in the other examples of contamination the rubidium was not normal in composition and it was apparently emitted from the ion source rather than the filament. In either case a correction was made by measuring the intensity of the mass 85 peak and the isotopic composition of the rubidium and then subtracting the contribution of rubidium from the mass 87 peak.

The precision error in measuring a series of 10 to 14 isotope ratios is less than 0.3 percent. The concentration of each tracer solution is considered accurate to within 1 percent. The estimated error in the weighing of tracers is less than 0.1 percent and is negligible. In the case of both rubidium and strontium an additional uncertainty lies in the apparent variance of the spectrometer discrimination. The problem of the measurement of the isotopic composition of normal rubidium introduces a possible error of $\pm$ 1 percent into the calculations. The uncertainty from variable strontium discrimination is less than 0.3 percent. Because both rubidium and strontium are measured on the same split of a mineral sample the problem of sample inhomogeneity does not occur as it does in the potassium-argon method. The precision of analytical reproducibility is considered to be 0.5 to 1.0 percent; duplicate analyses of a mineral sample yielded Rb-Sr ages that differed by less than 0.2 percent (Table 16). The precision error in a Rb-Sr age is inversely related to the ratio Sr^{87*}/Sr^{87}. That is, if the Rb/Sr ratio of the mineral is small, if it contains a large amount of common strontium, or if the mineral is young, the Sr^{87*}/Sr^{87} ratio could be smaller than 0.2 and the uncertainty in measuring Sr^{87*} could be as large as 10 percent. The Rb/Sr ratio is considered large if an enrichment in Sr^{87} greater than 20 percent is obtained. For a mineral 1000 million years old a Rb/Sr ratio of approximately 4 yields a 20 percent enrichment in Sr^{87} whereas a Rb/Sr ratio of approximately 40 is required to provide the same enrichment for a mineral 100 million years old. In the event of low enrichment, an isotope ratio analysis of the strontium can be made in order to reduce the uncertainty in Sr^{87*} to 1 to 3 percent. Therefore, the maximum estimated error in measuring Sr^{87*}/Rb^{87} is 4.3 to 6.3 percent, except that in the case of low enrichment in Sr^{87}, the error may be larger. The Rb-Sr ages presented in Part II are considered accurate to $\pm$ 5 percent.

The uncertainty in the half-life of Rb^{87} introduces an error into the accuracy of a Rb-Sr age that is exclusive of the errors in the measurement of the Sr^{87*}/Rb^{87} ratio.

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