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Rock Units of the Precambrian Basement in Colorado

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By OGDEN TWETO

GEOLOGY OF THE PRECAMBRIAN BASEMENT IN COLORADO

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Description and distribution of Proterozoic and Archean metamorphic and igneous rocks in outcrop and the buried basement in Colorado

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ROCK UNITS OF THE PRECAMBRIAN BASEMENT IN COLORADO

By OGDEN TWETO

ABSTRACT

The term Precambrian basement as used in this report applies to all rocks of Precambrian age. Basement rocks are exposed in many places in mountain uplifts in central and western Colorado, but are buried beneath younger rocks in more than 85 percent of the State. The exposed basement consists of variously metamorphosed layered rocks and of intrusive igneous rocks, in about equal proportions. The buried basement is of the same general character as the exposed basement except for the presence in southeastern Colorado of a unit of moderately metamorphosed sedimentary and volcanic rocks (Las Animas Formation) not known elsewhere.

The Precambrian rocks are all of Proterozoic age except in a small area near the northwest corner of the State, where Archean rocks at the south margin of the Wyoming cratonic age province are present, mainly in the subsurface. As exposed at the Colorado-Utah border and westward, the Archean rocks consist of gneisses about 2,700 m.y. (million years) in age and of the Red Creek Quartzite. The Red Creek is not conclusively dated but is interpreted as probably older than the 2,500-m.y. age boundary between the Archean and Proterozoic Eons. The Archean gneisses are mainly of granitic composition and igneous origin. They are exposed only in small fault blocks separated from the Red Creek by zones of quartz mylonite. The Red Creek Quartzite consists of at least 4 km (13,000 ft) of metaquartzite, kyanite-bearing quartz-muscovite schist, and minor marble, all extensively intruded by sills now metamorphosed to amphibolite. Faulting that occurred before and at various times after deposition of the Red Creek indicates recurrent tectonism at the border of the Archean Wyoming province.

The oldest and most widespread of the Proterozoic rocks in Colorado are gneisses in a complex that constitutes part of the floor on which younger Precambrian sedimentary units rest and the matrix into which igneous rocks were intruded. The gneisses were derived from a heterogeneous sedimentary and volcanic protolith. They are metamorphosed in the upper or the lower amphibolite facies in most places in Colorado, but are of lower metamorphic rank in a few areas. They are divided broadly into biotitic varieties, largely metasedimentary, and felsic and hornblendic varieties, largely metavolcanic. These rock units occur both in large discrete bodies and intimately interlayered with one another. The principal species of biotitic gneiss is biotite-quartz-plagioclase gneiss, some of which contains sillimanite, garnet, or cordierite. Metaquartzite and marble or calc-silicate rocks occur locally in the biotitic gneisses. In the subsurface in eastern Colorado, a metaquartzite-mica schist facies is predominant. The felsic gneisses are quartzofeldspathic and are thought to have been derived principally from tuffs and subvolcanic sills. The hornblendic gneisses include both amphibolites of basaltic composition and more silicic varieties probably derived from epiclastic rocks and tuffs of intermediate composition. Metamorphism of the gneisses peaked about 1,740 m.y. ago. Uranium-lead, samarium-neodymium, and lutetium-hafnium ages in the range 1,740–1,800 m.y. obtained from igneous components of the gneisses suggest that the age of the protolith was little greater than the age of metamorphism and probably does not exceed 1,800 m.y.

According to recent studies, the Vallecito Conglomerate of the Needle Mountains in southwestern Colorado is younger than and lies unconformably upon the Irving Formation of the Proterozoic gneiss complex. This is a reversal of some former interpretations but is in agreement with the original classification. The Vallecito is a massive unit of conglomerate and pebbly quartzite more than 1,500 m (5,000 ft) thick. The Uncompahgre Formation nearby may be a lateral equivalent of the Vallecito. The Uncompahgre consists of more than 2,650 m (8,700 ft) of quartzite, slate, and phyllite. The Uncompahgre and Vallecito are preserved in three areas in a northwest-trending belt in the western San Juan Mountains region. This belt may approximate the original depositional trough. Ages of the two formations are known only as being younger than a 1,695-m.y. granite and older than a 1,440-m.y. granite.

The Middle Proterozoic Las Animas Formation is recognized only in the subsurface in southeastern Colorado. The formation consists predominantly of dark-gray to black slate, phyllite, graywacke, and chert, but an upper part is dull red and contains minor volcanic and carbonate rocks. A thickness of more than 1,700 m (5,575 ft) has been penetrated in boreholes, without the base being reached. The formation occupies a west-trending belt about 130 km (80 mi) long and as much as 50 km (31 mi) wide. It is interpreted as fill in a rift incised in a terrane of mesozonal granite about 1,400 m.y. in age. The Las Animas is tentatively correlated with rocks in the Wichita Mountains province of southern Oklahoma that are thought to be younger than 1,400 m.y. and older than 1,200 m.y.

The Uinta Mountain Group constitutes the core of the eastern Uinta Mountains in northwestern Colorado and is present in the subsurface eastward at least to Craig. The group, which is more than 7 km (3.7 mi) thick, consists of quartzites accompanied by thick wedges of conglomerate and subordinate shale and siltstone. It lies in part on Archean rocks at the southern border of the Wyoming cratonic age province and evidently was deposited in a fault-bounded trough in that structural zone. The group ranges in age from about 925 to 1,100 m.y. and is equivalent in part to the compositionally similar Belt Supergroup of Montana and Idaho and Grand Canyon Supergroup of Arizona.

Igneous intrusion began during a waning stage of the main period of metamorphism and folding in the gneiss complex, and it continued episodically through the end of the Proterozoic. Numerous age determinations made on the igneous rocks fall statistically into three populations centered near 1,700, 1,400, and 1,000 m.y. On the basis of their distinctive characteristics, the rocks of the 1,700-m.y. age category are herein named the Routt Plutonic Suite, and rocks of the

1,400-m.y. age group are named the Berthoud Plutonic Suite. Rocks of the 1,000-m.y. age category are referred to as rocks of the Pikes Peak batholith.

Rocks of the Routt Plutonic Suite occur in many plutons throughout the State. The largest body is the Rawah batholith, which extends southwest at least 200 km from the west flank of the Front Range near the Wyoming border through the Park Range and the White River Plateau. Rocks of the Routt Suite are mostly granodiorite and quartz monzonite, but they range from gabbro to granite. They are synorogenic, foliated to various degrees, and concordant. They commonly contain inclusions of early mafic facies as well as of wall rocks, and, if porphyritic, are characterized by large prismatic potassium feldspar phenocrysts. Formally named rock units assigned to the suite are the Boulder Creek Granodiorite, Twin Spruce Quartz Monzonite, Cross Creek Quartz Monzonite, Denny Creek Granodiorite, Kroenke Granodiorite, Powderhorn Granite, Pitts Meadow Granodiorite, and Bakers Bridge and Tenmile Granites. All these units as well as the similar but unnamed rocks of various other intrusive bodies have been dated radiometrically. Ages of almost all the rocks fall in the range 1,665–1,700 m.y. as determined by rubidium-strontium, uranium-lead, and lutetium-hafnium methods.

Rocks of the Berthoud Plutonic Suite form many plutons in the exposed Precambrian terranes and occupy large areas in the subsurface in the south half of the State. Rocks of the suite are mostly granite and quartz monzonite but include minor syenite and gabbro. The rocks are anorogenic, predominantly discordant, mainly nonfoliated except for fluxion structure near contacts, and, if porphyritic, commonly contain tabular potassium feldspar phenocrysts, although a few varieties contain prismatic feldspars. Many of the granitic rocks contain both muscovite and biotite. Formally named rock units assigned to the suite are the Silver Plume Granite, Sherman Granite, Cripple Creek Granite, San Isabel Granite, St. Kevin Granite, Vernal Mesa Quartz Monzonite, Curecanti Quartz Monzonite, Eolus Granite, Trimble Granite, and Electra Lake Gabbro. With minor exceptions, these rocks range in age from 1,400 to 1,450 m.y. as determined by rubidium-strontium, uranium-lead, and lutetium-hafnium methods.

Mafic and intermediate dikes among the Precambrian rocks in some areas are mostly of about the age of the Berthoud Suite. Dikes in the northern Front Range follow or parallel faults of north-northwest trend. They cut the Sherman Granite of the Berthoud Suite but are cut by the Silver Plume Granite. Elsewhere in the northern Front Range, a long diabase dike—the Iron Dike—extends north-northwest from the Boulder area to the Wyoming border. This dike is younger than the Silver Plume Granite. In the northern Park Range, andesitic dikes both predate and postdate granite of the Berthoud Suite. In the northern Sawatch Range, metalamprophyre dikes are mainly younger than the St. Kevin Granite, but a few may be older.

The Pikes Peak batholith was emplaced in the last major Precambrian intrusive episode in Colorado. The batholith constitutes a large part of the southern Front Range and, as indicated by aeromagnetic data, extends far eastward in the subsurface beneath the Great Plains. The batholith is anorogenic, epizonal, composite, and sharply discordant. Large subcircular intrusive centers within it may mark the sites of calderas at higher levels. The batholith consists principally of coarse-grained orange-pink Pikes Peak Granite, but it also contains bodies of quartz syenite, alkali gabbro, fayalite-bearing granite, the riebeckite-bearing Mount Rosa Granite, the greisen-associated Redskin Granite, and the potassic Windy Point Granite. All these rocks are cogenetic, and all were emplaced in a period of less than 20 m.y. about 1,010 m.y. ago.

A final igneous event began just before the close of the Precambrian and continued into the Cambrian. At that time, mafic and alkalic intrusion occurred on a small scale in an old Precambrian fault zone in the Gunnison region. The intrusives there are of economic interest for their associated thorium, rare earths, and iron ore. Intrusives of similar character to the east in the Wet Mountains are about 50 m.y. younger, of Cambrian age.

Tectonic events that accompanied and influenced the rock-forming events are only outlined in this report. The protolith of the Proterozoic gneisses accumulated in an east-trending sedimentary trough flanked to the north and south by predominantly volcanic materials. Magnetic lineations and long faults of east trend suggest that the upper crust throughout Colorado is organized in units parallel to the trough. Fault systems at various angles to this trend began to form during folding and metamorphism of the gneiss complex, before intrusion of the Routt Suite. Faults or shear zones of northeast trend dictated the locations of borders and also the trends of some Routt Suite plutons. Later, these same faults as well as others influenced the locations and orientations of plutons of the Berthoud Suite.

INTRODUCTION

As exposed in mountain uplifts in Colorado, the ancient rocks that compose the Precambrian basement can be seen to contain features that, like architectural blueprints, anticipate the lines of future structures. The exposures afforded by the uplifts are scattered through the western three-fifths of the State, but altogether they constitute less than 15 percent of the area of the State. In the remaining 85 percent of the State, a Precambrian basement of the same general character as that exposed in the uplifts lies at various depths beneath a cover of younger rocks. Information on this buried basement is severely limited, but by different means some of its main features can be recognized. Knowledge of these features greatly extends the basis for a construct of the geologic anatomy and evolution of Colorado.

Although the exposed Precambrian rocks provide much evidence of events in the geologic development of Colorado, the evidence is in bits and pieces, in part because the areas of exposure are scattered. Integration of all the available records obviously is essential to an understanding of the basic geologic framework of the State. The nature and history of this framework have economic as well as scientific significance. Some of the rocks contain, or at least possibly contain, genetically related mineral deposits, and the framework fracture system controlled or influenced the sites of igneous activity and mineralization that followed in the Phanerozoic. In view of the geologic and economic incentives, a comprehensive investigation of the Precambrian basement logically followed compilation of the geologic map of Colorado (Tweto, 1979) as part of a U.S. Geological Survey program of investigation of the geologic framework and mineral resources of the nation.

This report, a product of that investigation, deals only with the rocks of the Precambrian basement. Studies of the structure and the influence of the basement on younger geologic features are in progress. In addressing the rocks, a first need is to summarize and bring into a regional perspective the many studies of the Precambrian rocks that have been made in the areas of exposed basement. This topic occupies most of the text that follows. The buried basement provides little opportunity for examination and, with one or two exceptions, the information concerning it is adequate only to classify it in terms of the rocks known in the exposed areas. This classification is expressed in the accompanying geologic map (pl. 1).

DEFINITION OF PRECAMBRIAN BASEMENT

The term Precambrian basement as used in this report applies to all rocks of Precambrian age. In most parts of the State, this basement is a readily identifiable unit of crystalline igneous and metamorphic rocks, such as granite, gneiss, and schist, lying stratigraphically below the Phanerozoic stratified sedimentary rocks. In some parts of the State, however, sequences of Precambrian sedimentary rocks lie upon the crystalline rocks and beneath the Phanerozoic sedimentary rocks. These Precambrian sedimentary rocks differ markedly from the crystalline rocks in physical properties, which resemble those of the Phanerozoic rocks. For this reason, in some types of seismic and tectonic analyses the Precambrian sedimentary rocks are effectively ignored, and the term basement is restricted to the crystalline rocks. Misidentification of Precambrian sedimentary rocks as Phanerozoic in boreholes also has led to an inadvertent restriction of "basement" to crystalline rocks. Thus, usages of the term Precambrian basement other than the one adopted here will be encountered.

REGIONAL FEATURES

Colorado occupies part of the North American craton, an extensive interior part of the continent characterized by a sialic or continental-type crust that was formed in Precambrian time. The uppermost part of the crust that formed at that time constitutes the Precambrian basement in the sense of this report. In Colorado, the total crust is about 48–50 km (30 mi) thick, considerably thicker than in the region to the west (Prodehl and Pakiser, 1980; Jackson and others, 1963). Beneath the crust are the much denser rocks of the Earth's mantle. The crust developed at different times in different parts of the continent (Engel, 1963; Goldich, Muehlberger, and others, 1966; Silver, Anderson, and others, 1977; Condie, 1982). An extensive southwestern region, including almost all of Colorado and large parts of contiguous States to the east, south, and west, was added to the continent between 1,700 and 1,800 m.y. (million years) ago. The oldest rocks in most of Colorado, here called the Proterozoic gneiss complex, were products (and also are the evidence) of that episode of continental growth. To the north, in most of Wyoming and a small area in the northwest corner of Colorado, is an older unit of the continent, generally known as the Wyoming province, of Archean age. The border of that province approaches the Colorado-Wyoming boundary from the northeast in southeastern Wyoming and then turns generally west-southwestward through northwestern Colorado and adjoining Utah (fig. 1). As noted in a following section on Archean rocks, the boundary between the Wyoming province and the province of Proterozoic rocks to the south is tectonic wherever it is observable.

Within Colorado, the Precambrian basement extends to the summits of many of the highest peaks and descends to great depths in some of the adjacent basins. Relief between the highest peaks and the deepest basins is more than 10.4 km (34,000 ft). This great relief resulted from tectonic disturbances in the late Paleozoic, the Laramide, and the late Tertiary summarized elsewhere (Tweto, 1980). In the Denver basin in eastern Colorado (fig. 2), the Precambrian basement is about 4 km (13,000 ft) below the surface between Denver and Colorado Springs, and about 1.8 km (6,000 ft) below the surface near the Kansas border. Thus, deep drilling is needed to reach the basement, and such drilling is lacking in a large part of the basin. South of the Arkansas River, by contrast, in the area of a buried late Paleozoic uplift, the basement is as little as 0.6 km (2,000 ft) below the surface, and boreholes to the basement are numerous (pl. 1). In western Colorado, the top surface of the basement is strongly deformed and plunges to depths of 7.6–8.2 km (25,000–27,000 ft) below the surface in the Piceance and Sand Wash basins. The basement in the deep parts of these basins has not been reached by the drill.

Phanerozoic igneous rocks scattered unevenly through the State rose from the depths through conduits that perforate the basement-rock surface. The size, shape, and exact locations of many such perforations can only be surmised. The areas of no basement in plate 1 mark the sites of the larger plutons and of calderas. The plutons now exposed in Precambrian settings are outlined accurately, but only generalized outlines are shown for those in sedimentary rocks because of uncertainty over their shape (or location) where they pierce the basement surface. Calderas are considered to be sites at which underlying batholithic bodies breached the surface. (See Lipman and others, 1978, fig. 11.) Thus, except for stoped blocks, basement rocks probably are absent to considerable depths beneath them. No attempt is made in plate 1 to show the perforations associated with many small intrusive bodies.

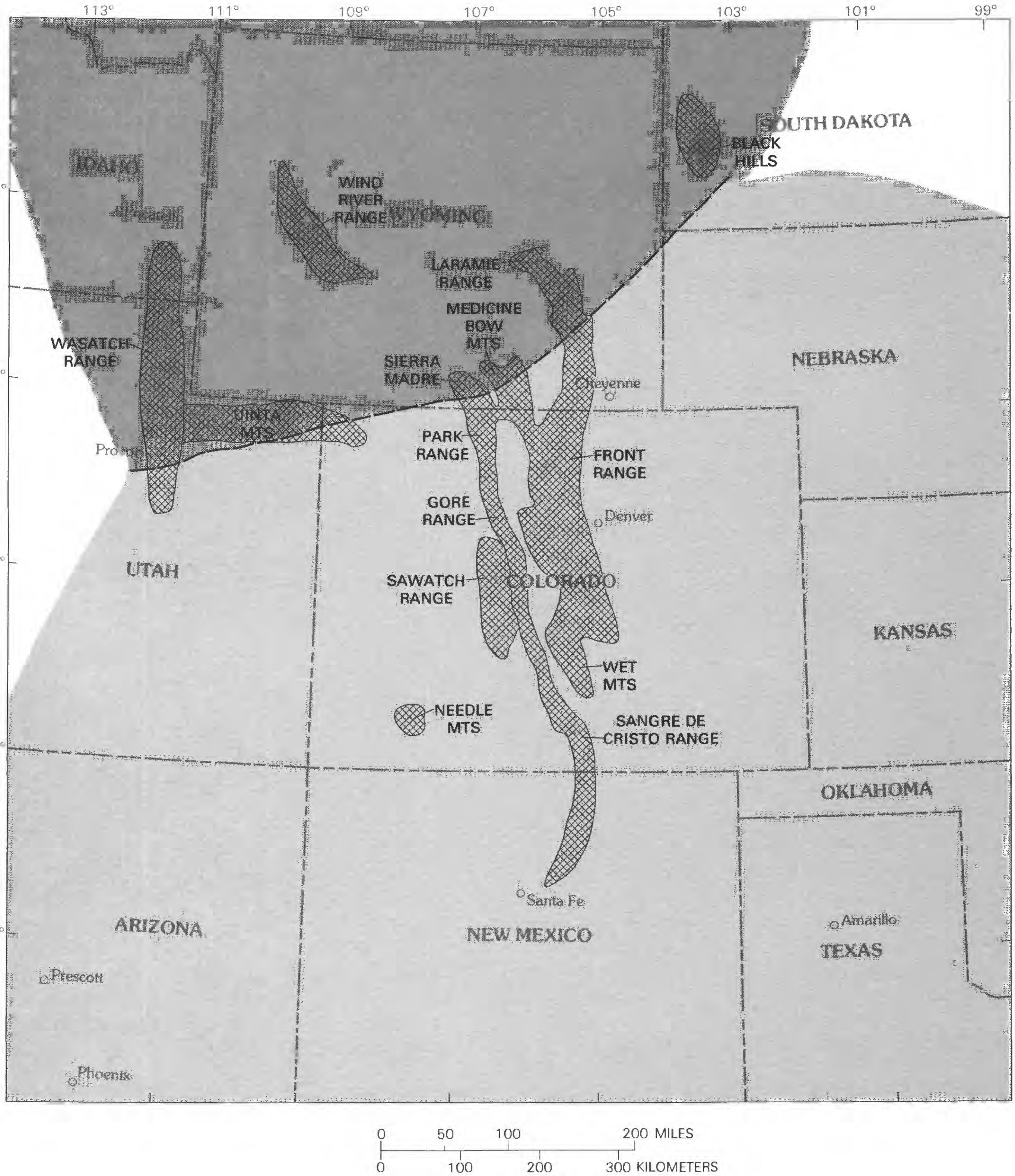


FIGURE 1.—Map of Proterozoic and Archean basement age provinces in Colorado and neighboring States. Proterozoic province, light shading; Archean province, dark shading. Province boundary (heavy dashed line) from Hills and Houston (1979) and Condie (1982).

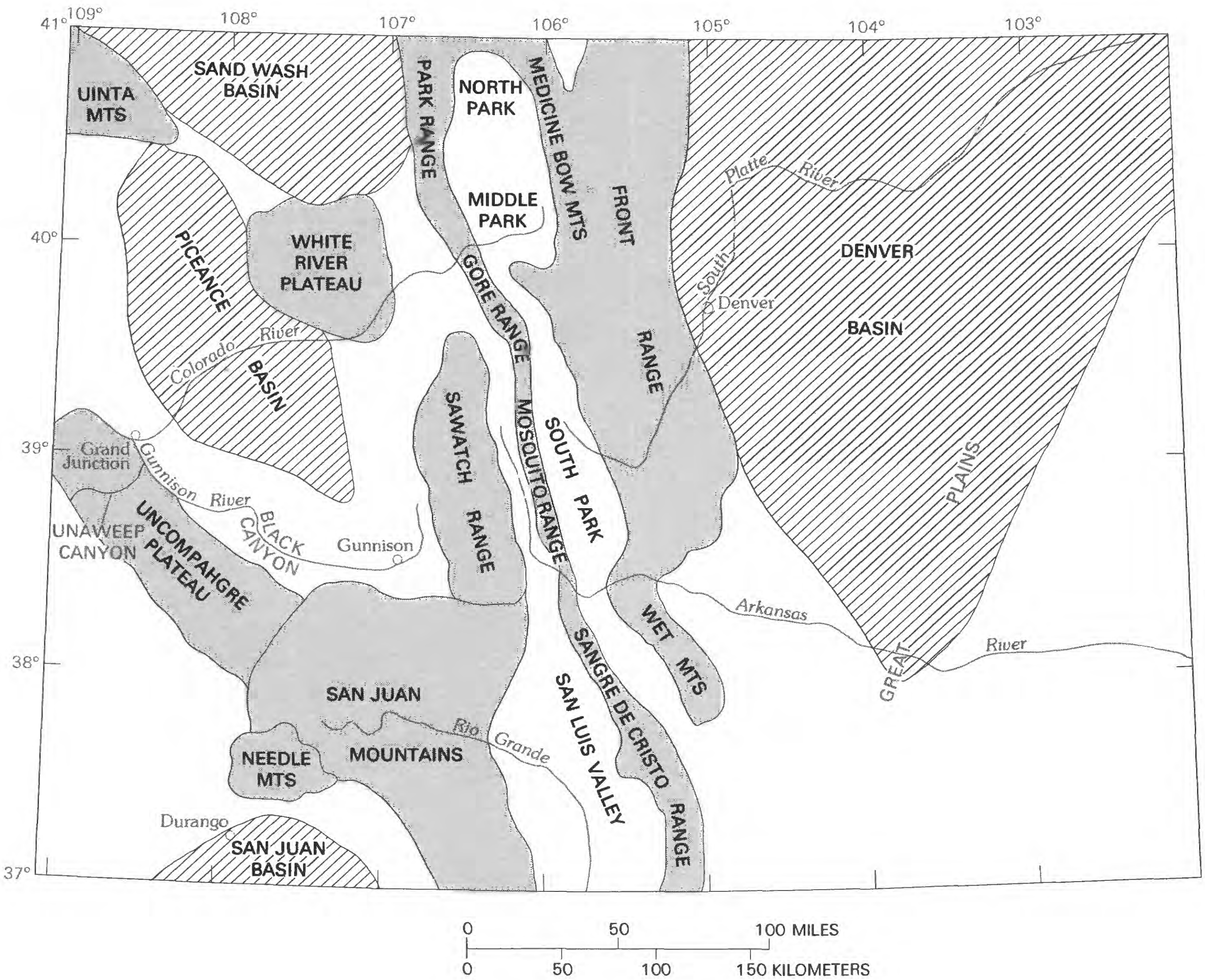


FIGURE 2.—Location map of major tectonic and geographic features in Colorado.

HISTORY OF INVESTIGATIONS

Precambrian rocks were discussed extensively in the reports of the territorial surveys in the 1870's (for example, Marvine, 1874) and in the accounts of later investigations until about 1910. These early studies were summarized by Van Hise and Leith (1909, p. 800–833). Users of any of this early literature must remain aware that age assignments were subjective and inconsistent. The rocks commonly were assigned either to the Archean or to the Algonkian, an obsolete analog of the Proterozoic. However, the distinction was based largely on the degree of metamorphism and deformation rather than on real age differences. All the rocks that were then called Archean in Colorado are now known to be Proterozoic.

The first detailed studies of the Precambrian rocks in Colorado were made by Cross (1894) and Mathews (1900) in the Pikes Peak region, by Cross and colleagues in the Needle Mountains (Cross, Howe, and others, 1905; Cross, Howe, and Ransome, 1905), and by Ball (1906) and Spurr and others (1908) in the central Front Range. A study of the Precambrian rocks of the Gunnison River region made in 1910–1912 (Hunter, 1925) brought to an end an early period of constructive field studies. By that time, several of the major units of Precambrian rocks in Colorado had been recognized. Only incidental studies of Precambrian rocks were made for many years after 1912.

In the 1930's, studies of the Precambrian granites were begun (for example, Boos and Boos, 1934) as was reconnaissance mapping of Precambrian rocks of the

Front Range (Lovering and Goddard, 1938; 1950) and the Sawatch Range (Stark and Barnes, 1935). Detailed studies that led to the present understanding of Precambrian structure, petrology, and tectonic history in the Front Range were begun in the early 1950's by several workers under the leadership of P. K. Sims. The advent of radiometric dating in the late 1950's added a revolutionary dimension to the exacting field and laboratory studies then in progress, and opened the way for correlations among the farflung areas of exposure of Precambrian rocks in Colorado and the Rocky Mountain region (Hedge and others, 1986). The generalizations embodied in the present report express not only the contributions of many field geologists through more than a century, but a sharpening made possible by geochronologists.

SOURCES OF INFORMATION

Geology of the exposed Precambrian terranes shown at scale 1:1,000,000 in plate 1 is a reduced and simplified version of what is shown on the 1:500,000-scale geologic map of Colorado (Tweto, 1979). Both maps summarize and codify information from hundreds of local maps and topical investigations, most of which are referred to in the following text. On the 1:500,000-scale Colorado map, the Precambrian igneous-rock units are informal age groups. In this report the two principal age groups are given formal stratigraphic status as the Routt and the Berthoud Plutonic Suites. Only the suites are identified in plate 1, but discussion in the text is in terms of the various formally named and unnamed rock units that compose the suites.

Geology of the buried basement, a principal feature of plate 1, is shown as inferred from: (1) the character of the Precambrian rocks in bordering areas of exposed basement, (2) the character of basement rocks penetrated by boreholes, (3) the character of inclusions of basement rocks in Tertiary plutons, (4) interpretations of aeromagnetic and gravity data, and (5) concepts of genetic processes and settings. Some of these guides are merely suggestive, and others, though definite in character, can be misleading. Boreholes provide samples of basement rocks, but experience with the exposed basement teaches that the rocks may differ markedly in distances little larger than the diameter of a borehole. For these reasons, and because all the data are sparse, the subsurface part of the map is speculative and purports to show no more than the inferred predominant rock types in general areas. Because the boundaries of these areas are unknown or can be located only within broad limits, contacts between map units are dashed in the subsurface parts of plate 1. The limitations cause an appearance of simplicity in the subsurface rock patterns, but if data were sufficient, the patterns doubtless would be as complex as in the exposed Precambrian terranes. Thus, the areas mapped as gneisses might expectably contain many small granite bodies, and the areas mapped as granites probably contain bodies of gneisses.

The principal source of information on the buried basement is from boreholes made in search of oil and gas. Such test holes are fairly numerous in some areas and sparse or absent in others, depending on the factors that influence exploration. The locations of 204 boreholes that reached the Precambrian basement are shown in plate 1. Records for about 20 percent of these holes are too poor to permit classification of the Precambrian rocks. For many of the remaining boreholes, petrographic data of Edwards (1966) were interpreted in terms of the map units recognized in this report. The rocks of many other boreholes were classified as interpreted from commercial lithologic well logs. Data on a few boreholes were obtained from brief notes on lithology of dated samples (Goldich, Lidiak, and others, 1966; Muehlberger and others, 1966; Hedge, 1967; Wrucke and Sheridan, 1968). Lithologic data on several boreholes in eastern Colorado were obtained from well logs made by J. C. Maher in support of subsurface stratigraphic investigations (Maher and Collins, 1949; 1952). Other sources of data used in classification were Barb (1946), Ryder (1977), Wood and others (1948), and Birch (1950).

Inclusions of basement rocks in Tertiary plutons provide clues to basement composition in some areas. Many plutons contain such inclusions, and closer attention to their compositions than has been paid in the past probably would provide much new information. From the available records, inclusions serve to distinguish Uncompahgre Formation from Irving Formation in the subsurface of the western San Juan Mountains. They also help to define the extent of a gabbro body at the Wyoming border north of Steamboat Springs, and to indicate the presence of granites or gneisses in various localities. Inclusions of pegmatitic materials found in many plutons are not indicators of any particular map unit, though they occur most commonly in biotite gneiss terranes.

Aeromagnetic data (Zietz and Kirby, 1972) are useful in some parts of the State as indicators of rock compositions in the Precambrian basement, but are inapplicable in other parts. Among the rock units having a distinctive magnetic signature, the Pikes Peak batholith is an excellent example. As exposed in the Front Range, the batholith correlates with a pronounced magnetic low. This magnetic anomaly continues eastward beyond the mountain front into the Great Plains region. On the basis of the anomaly, an extensive part of the batholith is interpreted to lie beneath the sedimentary rocks of the plains. Elsewhere on the plains, magnetic highs are inferred to outline bodies of granitic rocks of the Routt

Plutonic Suite, the presence of which is known from a few boreholes. Other correlations between magnetic features and particular rock bodies can be made in several places in the State. In many other places, however, a correlation is not evident, either because rocks of a given class are varied in their magnetic properties or because they give way at shallow depth to rocks with different properties. Granitic rocks of the Berthoud Plutonic Suite, about 1,400 m.y. in age, are particularly varied in magnetic expression, a fact noted years ago by Case (1966). The gneisses also have a wide range of magnetic properties and thus have no consistent magnetic expression, as might be expected from their range in compositions. In many areas in the mountains, the effects of Tertiary volcanic and intrusive rocks dominate the magnetic expression and thus preclude the identification of basement rocks from magnetic features.

Gravity data (Behrendt and Bajwa, 1974) have proved to be only of limited use in identifying rock units in the buried basement, although some general correlations are evident. A gravity trough of low amplitude in the Great Plains region corresponds with the part of the Pikes Peak batholith identified by magnetic data. Farther south in the Great Plains region, a gravity nose of low amplitude corresponds with a belt of metamorphosed sedimentary rocks (Las Animas Formation) in an otherwise granitic subsurface terrane. In most of the State, however, features other than the compositions of the basement rocks dominate the gravity expression. Among these features are: (1) a huge negative Bouguer gravity anomaly extending across the entire State, evidently an isostatic effect of the Rocky Mountains, (2) a tangential gravity valley trending northeastward across the mountain province, related to a high-level granitic batholith beneath the Colorado mineral belt (Tweto and Sims, 1963; Tweto and Case, 1972), (3) a deep negative anomaly of large area in the San Juan Mountains region, related to a shallow batholith that vented through 15 or more calderas (Steven and Lipman, 1976), and (4) various negative anomalies associated with thick sedimentary fill in Phanerozoic basins.

Isotopic studies of the leads in rocks and ores in the San Juan Mountains provide general information on the Precambrian basement beneath the Tertiary volcanic rocks covering that large region (Lipman and others, 1978; Doe and others, 1979). The lead data indicate a lithically complex basement similar to that in the surrounding region, but the information is too general to serve as a basis for mapping the basement. Moreover, the lead data indicate the compositions of rocks only at some indeterminate depth beneath the volcanics. Those rocks might or might not extend to the basement-rock surface.

STRATIGRAPHIC CLASSIFICATION

Accords reached in recent years divide Precambrian time into Archean and Proterozoic Eons separated by a time boundary at 2,500 m.y. before the present (James, 1978). The Precambrian rocks of Colorado are all of Proterozoic age (table 1) except in a small area in the subsurface near the northwest corner of the State, where rocks of Archean age are inferred to be present. Within the Proterozoic, the record begins at 1,800 m.y. ago or a little before and extends with interruptions to about 900 m.y. ago. This time span of 900 m.y. constitutes the latter part of the Early Proterozoic and all of the Middle Proterozoic as currently recognized (Harrison and Peterman, 1982, fig. 1). For cartographic purposes (pl. 1), major time divisions (eras) of the Precambrian are assigned letter designations according to the following scheme.

Time boundary (m.y. before present)	Era	Symbol
570		
	Late Proterozoic	Z
900		
	Middle Proterozoic	Y
1,600		
	Early Proterozoic	X
2,500		
	Late Archean	W
3,000		

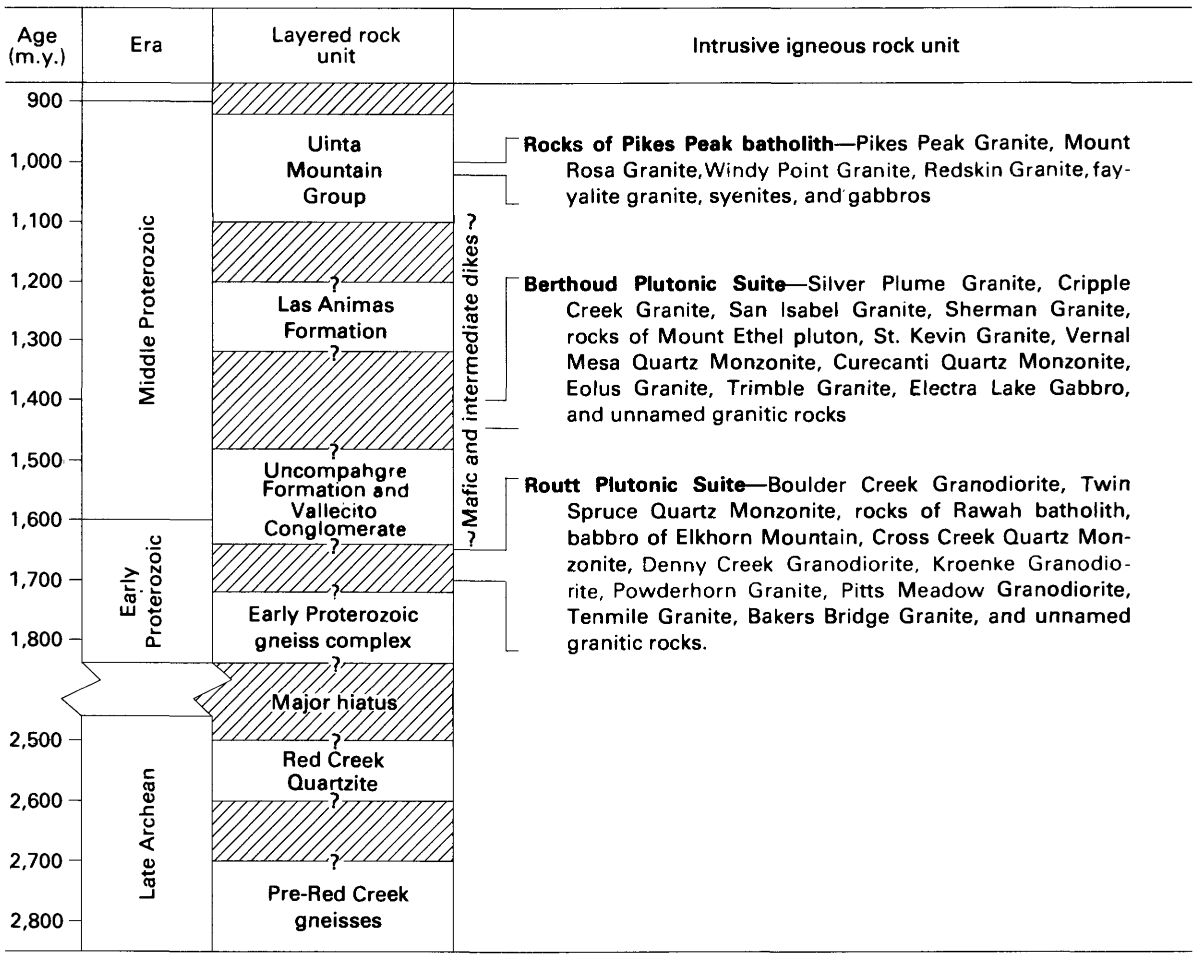
The oldest Proterozoic rocks are gneisses that constitute part of the floor on which younger Precambrian sedimentary units rest and the matrix into which the igneous rocks were intruded. The gneisses were derived from a varied sedimentary and volcanic protolith. They are complexly deformed and are of great but untold thickness. Stratigraphic names have been applied to them, or to facies of them, in various localities, but the distribution and age relations of lithic facies within them are as yet too poorly defined to warrant a formal stratigraphic nomenclature. Consequently, the gneisses are referred to only informally as the Early Proterozoic gneiss complex. A distinction is made on the geologic map (pl. 1) between biotitic gneisses largely of sedimentary origin and hornblendic and felsic gneisses largely of volcanic origin; except locally, these lithologies are not known to bear any fixed stratigraphic relation to each other.

Status of a controversial unit, the Vallecito Conglomerate, is changed herein. The Vallecito originally was classed as being younger than the gneisses, but later it was interpreted as being a lower element of the gneiss complex. On the basis of recent findings, the Vallecito is herein returned to its original status, stratigraphically above the gneiss complex and associated with the Uncompahgre Formation (table 1).

Concerning the intrusive rocks, two collective units are defined herein to embrace the numerous granitic

TABLE 1.—*Major Precambrian rock units in Colorado*[1]

[Queries indicate uncertainty in age limits of rock units. Diagonal pattern indicates hiatus]



[1]Not shown are igneous rocks straddling the Proterozoic-Cambrian age boundary (570 m.y.), known only in the Iron Hill stock south-southwest of Gunnison.

rocks or rock bodies of two different age groupings. In recent practice, the rocks of these two age categories were designated as the 1,700-m.y. and the 1,400-m.y. age groups (Tweto, 1977, 1979). The two age groups have distinguishing lithic characteristics, and as provided in the newly revised North American Stratigraphic Code (North American Commission on Stratigraphic Nomenclature, 1983) the rocks of the two groups are designated plutonic suites. They are given the names Routt Plutonic Suite and Berthoud Plutonic Suite, respectively. A third age group, about 1,000 m.y., is mapped and discussed under the heading, "Rocks of the Pikes Peak batholith."

ROCK TERMINOLOGY

The various reports on which this synthesis is based reveal some inconsistencies in the terminology of igneous rocks, both among authors and through time. In many reports in the past, the term granite was applied loosely to rocks in a range of compositions. Later, distinctions commonly were made between granite in a restricted sense and quartz monzonite and granodiorite. Many of the Precambrian batholiths in Colorado contain rocks of all these compositions, but for convenience of reference, a single term often is chosen. Authors have differed in their choices, and this has led to differing designations in the formal stratigraphic nomenclature for the same rock unit in the same area, such as Boulder Creek Granite, Boulder Creek Quartz Monzonite, or Boulder Creek Granodiorite. Most of the rocks that have been called quartz monzonites in Colorado are granites in the classification recently adopted by the International Union of Geological Sciences (Streckeisen, 1976). Use of that classification would eliminate nomenclatorial awkwardness for the rock units of concern here in the granite-quartz monzonite range. On the other hand, a distinction between the two compositions, if strictly observed, is more precise. The petrology of the granitic rocks not being the aim of this report, the terminology of the original authors is followed throughout, except a few changes that are explicitly noted.

Inconsistencies are evident also in the use of mineral modifiers for both igneous and metamorphic rocks. If there are two or more modifiers, many authors list them in order of increasing abundance. In this usage, "muscovite-biotite granite" means that the rock contains more biotite than muscovite. But some authors evidently have followed an opposite system. Whatever the persuasion, most authors do not state explicitly what system they are using. For this reason, the terminology of the original authors is followed throughout this report.

RADIOMETRIC AGES

Radiometric ages noted in this report have been recalculated as necessary to conform with the isotopic ratios and decay constants recommended by the International Union of Geological Sciences Subcommission on Geochronology (Steiger and Jäger, 1977).

ARCHEAN ROCKS

Rocks of Archean age are exposed in Colorado only in an area of about a square kilometer (0.4 mi^2) at the Utah border on the north side of the Uinta Mountains. From that locality eastward to about long 108° W., Archean rocks are inferred to be present in the subsurface in a narrow strip along the Wyoming border (pl. 1). This strip lies along the southern margin of the Wyoming province, a major crustal unit of Archean rocks in the northern Rocky Mountain region. In the region south of the Wyoming province, including almost all of Colorado, the oldest rocks known are of Proterozoic age. The boundary between the two age provinces is well established in southeast Wyoming (Hills and Houston, 1979), where it extends southwest across the Laramie and Medicine Bow Ranges and then, just north of the Colorado border, turns generally westward across the Sierra Madre (fig. 1). West of the Sierra Madre, points on or near the boundary are known only at the locality near the Colorado-Utah border and at lat 40° N. on the west side of the Wasatch Range in Utah (Hedge and Stacey, 1980). Thus, the boundary trends generally west-southwest from the Sierra Madre to the Wasatch front, and for about 100 km (60 mi) it lies in Colorado (fig. 1). Its precise location in the Uinta Mountains area is not known because all the exposed Precambrian rocks have been displaced northward in the upper plate of a Laramide thrust fault that extends along the north flank of the mountains. A minimum horizontal displacement of 5 km (3 mi) on this fault has been proved by drilling.

On the north side of the Uinta Mountains, at the Colorado-Utah border and westward, the Middle Proterozoic Uinta Mountain Group lies unconformably upon strongly deformed metamorphic rocks assigned to the Red Creek Quartzite by Hansen (1965, pl. 1) and predecessor workers beginning with Powell (1876). More recently, Graff and others (1980) found that within some of the areas previously mapped as Red Creek are fault slivers or blocks of gneisses that are older and more highly metamorphosed than the quartzite and mica schists of the Red Creek. They interpreted these gneisses as upfaulted segments of a pre-Red Creek basement, and they referred to the gneisses as the Owiyukuts Complex.

As described by Graff and others (1980) and Sears and others (1982), the pre-Red Creek gneisses are exposed in a canyon at the Colorado-Utah border and in scattered small fault-bounded areas to the west. The gneisses are mainly of granitic composition and igneous origin. They include granite gneiss, quartzofeldspathic gneiss, garnet gneiss, biotite gneiss, migmatite, and garnet amphibolite. These rocks are interleaved with feldspathic quartz mylonites, and they are separated from the metaquartzites of the Red Creek by zones of quartz mylonite. The gneisses display the sillimanite-potassium feldspar association of the upper amphibolite facies in contrast to assemblages of the lower amphibolite facies in the Red Creek. Age of granitic gneiss from secs. 12 and 13, T. 11 N., R. 104 W., Moffat County, Colo., was determined to be 2,700 ± 50 m.y. by the Rb-Sr (rubidium-strontium) method by C. E. Hedge of the U.S. Geological Survey (Graff and others, 1980, p. 24). Thus, the gneisses clearly are of Archean age and are an element of the Archean Wyoming province.

The Red Creek Quartzite consists of metaquartzite, subordinate kyanite-, staurolite-, and garnet-bearing quartz-muscovite and chlorite schists, and minor marble, all intruded by small mafic sills metamorphosed to amphibolite (Hansen, 1965; Graff and others, 1980). The rocks are tightly folded and extensively faulted. On the basis of structural reconstructions, Hansen (1965, p. 23) estimated the exposed thickness of the Red Creek, including the older gneisses, to be about 6 km (20,000 ft). Graff and others (1980, p. 47) computed a thickness of 4 km (13,000 ft), exclusive of the older gneisses. The metaquartzite and schist of the Red Creek are metamorphosed in the almandine amphibolite facies and contain assemblages of the staurolite-quartz and kyanite-muscovite-quartz subfacies. Muscovite from the schist was dated as 2,390 m.y. in age by the Rb-Sr method and 1,533 m.y. by the K-Ar (potassium-argon) method (Hansen, 1965, p. 31). Four amphibolite samples yielded K-Ar ages ranging from 930 to 1,613 m.y. (Graff and others, 1980, p. 57). The scatter in the age data suggests that the 2,390-m.y. Rb-Sr muscovite age probably is only a minimum metamorphic age for the Red Creek. Thus, the age of the formation could lie on either side of the 2,500-m.y. time boundary between the Archean and the Proterozoic Eons. In keeping with some recent practice (Tweto, 1979) it is here classed as Late Archean, but an Early Proterozoic age has also been suggested on the basis of lithologic resemblance to parts of the Libby Creek Group in the Medicine Bow Mountains of Wyoming, inferred to be of Proterozoic age (Graff and others, 1980, p. 63; Karlstrom and others, 1981, p. 90; Hansen, 1965, p. 31).

The mylonite zones separating the Archean gneisses from the quartzites of the Red Creek are evidence of an early tectonic episode. According to the interpretation of Sears and others (1982), the miogeoclinal Red Creek Quartzite was thrust northward over the margin of the Archean craton in this episode. In the Medicine Bow Mountains of Wyoming, Early Proterozoic rifting along the boundary zone was followed by strike- or oblique-slip faulting (Houston and others, 1979; Hills and Houston, 1979). Middle Proterozoic rifting along the boundary zone in the area of the Uinta Mountains is inferred to have produced the trough in which the Uinta Mountain Group accumulated. (See section on Uinta Mountain Group.) Thus, movements of various kinds occurred in segments of the boundary zone at various times in the Proterozoic, and they recurred in the Phanerozoic. A similar history of episodic movements at various places along a major crustal boundary in the Great Lakes region has been documented by Sims and others (1980).

The character of the buried Archean rocks in Colorado is problematic. As judged from the exposures in the Uinta Mountains and the Sierra Madre in Wyoming (Houston and Ebbett, 1977; Divis, 1976), the rocks are most likely to be feldspathic gneiss, amphibolite, and metaquartzite. The strip of Archean rocks corresponds with a linear aeromagnetic low (Zietz and Kirby, 1972) in an area of subdued magnetic topography beneath thick sedimentary cover in the Sand Wash basin. The low possibly may reflect a preponderance of quartzitic or felsic rocks.

PROTEROZOIC LAYERED ROCKS

GNEISS COMPLEX

In most of Colorado, the oldest rocks belong to a complex of metasedimentary and metavolcanic gneisses of late Early Proterozoic age. The gneiss complex is part of an extensive province of Early Proterozoic rocks that extends southward from the Wyoming Archean province through Colorado into New Mexico and Arizona and beyond (fig. 1). The complex consists of many varieties of metamorphic rocks, reflecting varied parent materials, differences in the degrees of metamorphism and penetrative deformation, and wide ranges in the degree of modification by igneous processes or by retrograde metamorphism. Gneisses of different kinds occur both as large discrete bodies and closely interlayered or intertongued with each other. In this report as on the State map (Tweto, 1979), the gneisses are divided broadly into biotitic varieties, largely derived from sedimentary rocks, and felsic and hornblendic varieties, largely derived from volcanic and related intrusive rocks.

A sharply defined pattern of distribution of the two groups of gneisses is evident in plate 1. The biotitic gneisses occupy a broad belt extending east-west through the central part of the State, and the felsic and hornblendic gneisses occur principally in bordering belts to the north and south. In the areas of exposed Precambrian rocks, the biotitic gneisses are mainly in the central and north-central parts of the Front Range, the Gore and Tenmile Ranges, the northern Sawatch Range, and the Gunnison River area (fig. 2). The biotitic gneisses also occupy extensive areas in the subsurface in west-central Colorado as indicated by scattered small exposures, by boreholes, and by xenoliths in Tertiary intrusive bodies. In the Denver basin in eastern Colorado, boreholes establish the presence of biotitic gneisses from the Fort Collins area eastward at least to long 104° W., but no information is available for a large area farther south except in a strip near the Kansas border. In this strip, several boreholes indicate that a mica schist-quartzite facies dominates the biotitic gneisses (pl. 1).

Felsic and hornblendic gneisses occur in many small bodies within the belt of biotitic gneisses but are far more widespread to the north and south. In the northern part of the State, felsic and hornblendic gneisses form the framework between granitic bodies in the northern Front and Park Ranges and in the subsurface to the east and west. In the southern part of the State, these gneisses are the principal Precambrian rocks of the Wet Mountains, Sangre de Cristo Range, southern Sawatch Range, and the Needle Mountains (pl. 1). The gneisses are also widespread as a matrix between granite bodies in the subsurface as shown by several boreholes in the southeast and a few, together with xenolith occurrences, in the southwest.

The gneiss complex is of great but unknown thickness. No indication of its base has been recognized, nor is there any knowledge of where the oldest or the youngest exposed parts may lie. Igneous intrusion obliterated the complex in many areas and deformed the gneisses that remain. Erosion in Precambrian and later times removed enormous amounts of the gneisses throughout the State. Estimates of exposed thicknesses have been made only in local areas. For example, Sims and Gable (1967, p. 7) estimated an exposed thickness of about 5 km (3 mi) in the Central City 7½-minute quadrangle in the Front Range, but this estimate takes no account of what may be unexposed, or, as compared with other areas, of how much may be missing due to erosion. Until the major elements of the structure and stratigraphy of the gneisses are worked out, the thickness will remain unknown. Seismic investigations (Jackson and others, 1963; Prodehl and Pakiser, 1980) indicate that the siliceous upper crust is about 28 km (17 mi) thick in Colorado. The gneisses constitute some part of that thickness, but the data do not allow a distinction between the gneisses and the abundant accompanying siliceous igneous rocks.

The gneisses were metamorphosed and deformed in a first period of folding and were undergoing a second period of folding when the intrusion of igneous rocks began about 1,700 m.y. ago. Metamorphism occupied some unestablished period of time and probably was not everywhere synchronous but seems to have peaked about 1,740 m.y. ago (Hedge and others, 1967, 1968; Hansen and Peterman, 1968; Silver and Barker, 1968; Stern and others, 1971). Ages determined on selected metavolcanic materials are thought to be indicators of the age of the parent rocks of the gneisses, but always with the reservation that the values may have been affected by metamorphism. U-Pb (uranium-lead) ages of $1,765 \pm 5$ and $1,740 \pm 5$ m.y. were obtained from zircons in metarhyolites from two localities in the Gunnison region (Bickford and others, 1982). Earlier, the felsitic gneisses in the Needle Mountains had yielded a U-Pb age of about $1,753 \pm 20$ m.y. (Silver and Barker, 1968), and the felsitic and basaltic rocks had yielded a Rb-Sr isochron age of about $1,770 \pm 35$ m.y. (Barker, 1969a, p. 19). Ages obtained by the recently introduced Sm-Nd (samarium-neodymium) and Lu-Hf (lutetium-hafnium) methods differ little from the U-Pb ages. Felsic and basaltic gneisses in both the Front Range and the Needle Mountains yielded Sm-Nd ages of $1,800 \pm 90$ m.y. (DePaolo, 1981). The Lu-Hf method yielded similar ages ($1,800 \pm 90$ m.y.) for the felsic and basaltic gneisses and two pelitic rocks in the Front Range (Patchett and others, 1981). Stratigraphic relations among the dated rocks in the different localities are undetermined, and so also are the stratigraphic positions of the dated rocks within the gneiss complex as a whole. The data presently available suggest 1,800 m.y. as an approximate average age of the protolith of the gneiss complex.

BIOTITIC GNEISSES

The rocks grouped as biotitic gneisses are predominantly biotite-quartz-plagioclase gneisses and schists, but include subordinate amounts of several other varieties of metasedimentary rocks. Many different stratigraphic names have been applied to the biotitic gneisses, mostly to facies occurring in local areas (Tweto, 1977). Names applied most widely are the Idaho Springs Formation in central Colorado and the Black Canyon Schist in the Gunnison River region. The term Idaho Springs first was applied by Ball (1906) to biotite schist or gneiss, quartz gneiss, and associated calc-silicate rocks in the Idaho Springs area and Georgetown quadrangle in the central Front Range. The name later was applied to all biotitic gneiss and schist throughout the Front Range (Lovering and Goddard, 1950, pl. 1), and some authors have applied the term even more widely. Thus, the term has lost stratigraphic significance because the biotitic rocks occur repeatedly in stratigraphic sections of great, though unknown, thicknesses. Moreover, because so general a unit serves no purpose in detailed mapping, the term has fallen into disuse, and further use of it has been discouraged (Tweto, 1977). Lithic terminology for units of the biotitic rocks has been used in most mapping in recent years. The term Black Canyon Schist was applied by Hunter (1925) to gneiss or schist and associated rocks in the Black Canyon of the Gunnison River and the area to the southeast. The term was not used in a later detailed study of the Black Canyon area (Hansen, 1971), but it was used in a study nearby (Olson and Hedlund, 1973). Hunter (1925) also distinguished a fine-grained quartz-muscovite schist in the Black Canyon as the River Portal Mica Schist, a term that subsequently was ignored and finally was abandoned (Tweto, 1977).

The biotitic gneisses have been studied more intensively in the central Front Range than elsewhere because of their widespread occurrence in the several mineralized areas in that part of the range (Sims and

Gable, 1964, 1967; Moench, 1964; Braddock, 1969; Sheridan and others, 1967). As constituted there, the gneisses are varied in lithology. They have a layered structure, and foliation is generally parallel to the layering. Many different lithologies and related transitional compositions are present among the interlayered rocks. In addition to varieties of biotite gneiss, the biotitic gneisses locally contain layers or lenses of other rocks of metasedimentary character, such as quartzite or quartz gneiss, calc-silicate gneiss, hornblende gneiss, and marble. Additionally, the biotite gneiss is migmatized in many places. Where the granitic fraction predominates in migmatitic rocks, or where abundant pegmatite is present in irregular bodies, the rock has been classed as "granite gneiss and pegmatite" (Harrison and Wells, 1956, p. 50; Sims and Gable, 1964, p. 31).

Biotite gneiss typically consists mainly of plagioclase (calcic oligoclase or sodic andesine), quartz, and biotite, but any of several other minerals may be present in minor amounts. Varieties characterized by sillimanite, garnet, or cordierite are widely distributed. The sillimanitic rocks generally contain muscovite, which in some rocks is mainly primary but in others is mostly a product of retrograde reactions. In some of the gneisses, sillimanite and potassium feldspar are present in equilibrium assemblages, indicating metamorphism in the upper amphibolite facies. Somewhat more intense metamorphism is indicated by the occurrence of cordierite and garnet along with potassium feldspar and sillimanite in some gneisses, or of cordierite and gedrite in accompanying, more magnesian, gneisses (Gable and Sims, 1970). Magnetite is present in lieu of biotite in a few local layers of gneiss. Hornblende is present with biotite in zones of transition into hornblendic rocks. Potassium feldspar, generally microcline, is a common minor constituent of many varieties of biotite gneiss, and it is a major component in migmatitic varieties. In some localities, lenses of migmatitic biotite gneiss contain a few percent of the rare-earth minerals xenotime and monazite. Young and Sims (1961) interpreted these minerals as products of the migmatizing fluids, mobilized from the surrounding rocks. Disseminated uraninite present in migmatitic gneiss in another locality (Young and Hauff, 1975) is of the age of nearby granite of the 1,400-m.y. Berthoud Suite (Ludwig and Young, 1975).

Mica schist is interlayered with biotite gneiss in some places, especially near the mountain front just north of Golden. As described by Sheridan and others (1967, p. 7), the schist consists largely of quartz, muscovite, and biotite, which are accompanied in places by sillimanite, andalusite, or garnet. Some quartz-rich varieties grade to quartzite or quartz gneiss, and some varieties containing plagioclase grade to biotite gneiss. Lenses and thin layers of conglomerate and of calc-silicate rocks are intercalated in the schist. Pebbles in the conglomerate are mostly less than 3 cm (1.2 in.) in diameter and consist of quartzite and quartz-tourmaline rock. Some of the calc-silicate rock contains abundant calcite and grades to impure marble.

Quartzite or quartz gneiss occurs as lenses and layers in the biotite gneiss in many places. A large body of quartzite is near the mountain front at Coal Creek, between Golden and Boulder, where it lies between the Idaho Springs-Ralston shear zone on the southeast and the 1,700-m.y. Boulder Creek batholith on the northwest (fig. 3). The quartzite once was thought to be a younger stratigraphic unit than the biotitic gneisses (Lovering and Goddard, 1950, p. 23), but later it was found to intertongue to the west with the biotite gneisses and therefore was classed as a facies of the parent rocks of the gneisses (Wells and others, 1964). As described by Wells and others (1964), the quartzite is in four thick layers separated by layers of fine-grained mica schist. The individual quartzite layers pinch and swell, reaching thicknesses of as much as 1,100 m (3,600 ft). The quartzite is foliated but bedding is preserved and crossbedding is visible locally. Conglomerate beds occur throughout the quartzites. Pebbles in them are well rounded, generally less than 2.5 cm (1 in.) in diameter, and are mainly quartzites. Most of the quartzite in all four of the layers is micaceous and consists almost entirely of quartz and muscovite. Sillimanite, andalusite, plagioclase, microcline, chlorite, and tourmaline are minor constituents present locally. The schist interlayered with the quartzite consists mainly of muscovite, biotite, and quartz, but locally contains andalusite, cordierite, garnet, sillimanite, staurolite, or plagioclase. The staurolite occurs only as relict inclusions in andalusite. Thus, the key metamorphic minerals in the quartzite and mica schist are the same as in the biotite gneiss.

In other areas, quartz gneisses of various mineralogies are found. Most of the gneisses contain some biotite, muscovite, or plagioclase, and some are characterized by epidote, magnetite, or garnet (Hawley and Moore, 1967; Sheridan and others, 1967). An unusual rutile-bearing sillimanitic topaz-quartz gneiss occurs in layers and lenses in a belt between Idaho Springs and Morrison. As described by Sheridan and others (1968) and Marsh and Sheridan (1976), the content of the titanium mineral rutile ranges from a trace to about 5 percent, and the content of the fluorine-bearing mineral topaz ranges from 23 to 67 percent. In one locality, a rare magnesium fluophosphate mineral, wagnerite, is present together with corundum and rutile in sillimanitic biotite gneiss (Sheridan and others, 1976). The

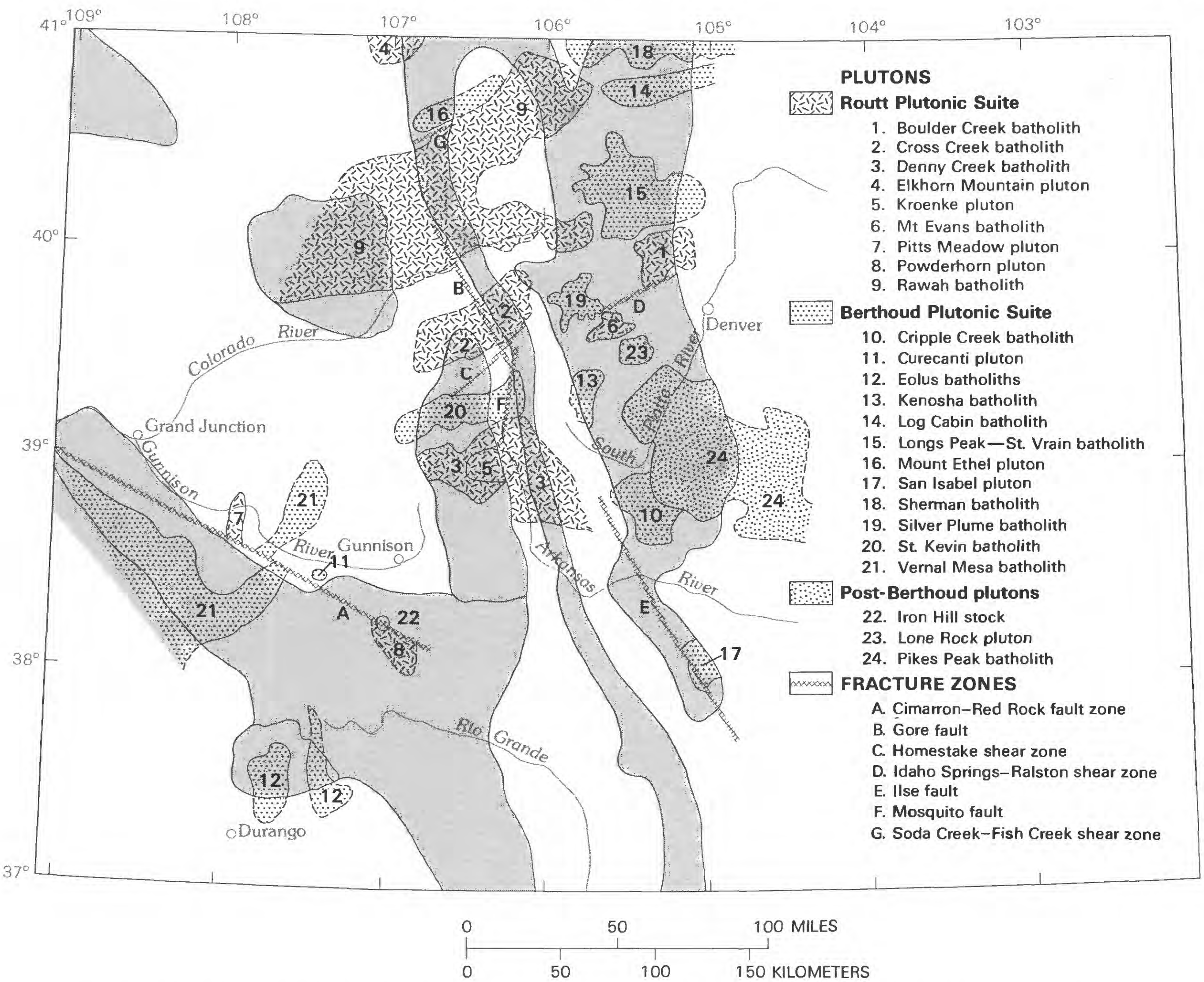


FIGURE 3.—Principal Precambrian plutons and some major fracture zones in Colorado. Boundaries dashed where covered by younger rocks. Outlines of mountain uplifts shaded as in figure 2.

gneisses containing these minerals of possible economic interest are in an interlayered sequence of sillimanitic and cordieritic biotite gneisses, calc-silicate gneisses, hornblende gneisses, and felsic gneisses. The parent rocks of the rutile-, topaz-, and wagnerite-bearing gneisses are thought to have been clays derived from the weathering of volcanic rocks of intermediate to mafic compositions, and the fluorine of the topaz and wagnerite is postulated to be of volcanic origin and to have been adsorbed by the clays (Marsh and Sheridan, 1976; Sheridan and others, 1976).

Calc-silicate and related rocks are interlayered with the biotitic gneisses in many places, but commonly are in small bodies (Sims and Gable, 1964; Moench, 1964; Sheridan and others, 1967). The rocks are highly varied, consisting of many different combinations of amphiboles, clinopyroxenes, members of the epidote group, calcic plagioclase, quartz, magnetite, garnet, scapolite, and vesuvianite. In a few localities, the tungsten mineral scheelite, or a molybdenum-bearing variety of it, is one of the metamorphic minerals (Tweto, 1960). Some of the calc-silicate rocks contain calcite, and the rocks grade to marbles with increasing content of that mineral. Most of the marbles are impure, containing scattered grains of calc-silicate minerals or quartz, but lenses or pods of clean marble occur locally. Rarely, marbles retain chert from the original limestones. Hornblende gneiss generally accompanies the calc-silicate rocks and marble. The gneiss of this occurrence commonly contains calcic minerals such as epidote and diopside and grades to calc-silicate rock. It is in small bodies and is distinct from the hornblendic metavolcanic rocks, which are in much larger bodies.

Biotitic gneisses in other parts of the State are generally similar to those of the central Front Range. In the Thompson River area, west of Loveland, the metamorphic grade is lower than in most other places. Some of the rocks are greenschist-facies phyllites and schists. These grade into biotite gneisses characterized in successive metamorphic zones by garnet, staurolite, sillimanite-muscovite, and finally the sillimanite-microcline of the upper amphibolite facies (Braddock, 1965; Braddock, Nutalaya, and others, 1970). Graded bedding is preserved in many places in the lower grade rocks, and the area is one of the few places in Colorado where the direction of the "tops" or stratigraphic succession in the gneisses is clearly established.

In the Gore Range and the northern Sawatch Range, the biotitic gneisses are mainly migmatites (Tweto and others, 1970; Tweto, 1974; Tweto and Lovering, 1977). The migmatites show abundant evidence of plastic flow and are structurally disorganized. They surround and grade into lenses of contaminated, foliated Cross Creek Quartz Monzonite and evidently are products of partial melting of biotite gneiss. The biotite gneiss component of the migmatites contains sillimanite in apparent equilibrium with microcline, and locally it is cordieritic or garnetiferous. Where present as streaks or xenoliths in the granitic rock, the biotite gneiss commonly is reduced to a refractory mixture of sillimanite, quartz, and retrograde muscovite.

Biotitic gneisses in the Black Canyon of the Gunnison area consist of quartzitic gneiss, mica schist, and migmatitic gneiss as classified by Hansen (1971). Earlier, Hunter (1925) had classified the same rocks as schists of several varieties. Migmatitic gneiss is the most abundant facies. The quartzitic gneiss is in thick layers parallel to bedding. Some of it grades to micaceous quartzite, and some contains rare lenses of conglomerate. The mica schist consists of predominant quartz and muscovite and subordinate biotite. All the rocks locally contain sillimanite and garnet, and the mica schist contains staurolite and andalusite.

In the subsurface in eastern Colorado, at and near the Kansas border, mica schist and quartzite are the predominant basement rocks. The rocks resemble the mica schist and quartzite near the mountain front north of Golden, discussed previously, and they are interpreted to be of the same general age and origin as the rocks north of Golden. Sparse borehole data (pl. 1) suggest that the schist and quartzite intertongue with both biotite gneiss and felsic gneiss and that they were invaded by a body of granitic rock of the 1,700-m.y. Routt Suite. The schist and quartzite extend eastward into Kansas, where they occur in scattered bodies in a predominantly granitic terrane (Bickford and others, 1979). They also extend into southwestern Nebraska (Lidiak, 1972). The metamorphic grade of the schist and quartzite is not uniform and is somewhat lower than that of the gneisses in the exposed terranes. Some of the schist has the chlorite-muscovite association of the greenschist facies (Edwards, 1966), but varieties containing biotite and garnet also occur, and Lidiak (1972, p. 19) identified some with the sillimanite-muscovite association of the lower amphibolite facies.

FELSIC AND HORNBLENDIC GNEISSES

Rocks classed as felsic gneisses have a wide range of compositions, but in general are light colored and consist largely of quartz, plagioclase, and potassium feldspar. They are fairly uniform in lithology throughout any given layer or body. Hornblendic gneisses are also varied in composition, ranging from black amphibolites consisting almost entirely of hornblende and plagioclase to banded dark-gray gneisses containing quartz, biotite, or potassium feldspar as well as hornblende and plagioclase. The felsic and hornblendic gneisses are intimately interlayered in some places and are in separate large bodies in others. Both are interlayered or intertongued with biotitic gneisses.

Various stratigraphic and lithologic names have been applied to the gneisses. In the central Front Range, rocks that herein are grouped as felsic gneiss once were interpreted as igneous rocks intrusive into the biotitic gneisses. Accordingly, they were called quartz monzonite gneiss and gneissoid granite in the pioneer work by Ball (Spurr and others, 1908) and were called granite gneiss, gneissic aplite, and quartz monzonite gneiss by Lovering and Goddard (1950). Detailed studies later established that these rocks are not intrusive but are interlayered with the biotitic gneisses. The nongenetic terms microcline-quartz-plagioclase-biotite gneiss (often abbreviated microcline gneiss) or feldspar-rich gneiss have been applied to them in recent studies (for example, Sims and Gable, 1967; Braddock, 1969; Taylor, 1975a). Hornblendic gneisses in a large body near Montezuma were named the Swandyke Hornblende Gneiss and assigned a post-Idaho Springs age by Lovering (1935, p. 10). Later, Lovering and Goddard (1950, p. 20, pl. 2) extended the Swandyke to include all hornblende gneisses in the Front Range mineral belt and classed it as being in part of the same age as the Idaho Springs and in part younger, though the younger age was never proved. Just as with the Idaho Springs, the Swandyke thus lost stratigraphic significance, and it was later abandoned (Tweto, 1977). In the Gunnison region, where metamorphism and deformation were less intense than in the Front Range, felsic volcanics, mafic volcanics, and various types of sedimentary rocks are interlayered in a sequence of great but unknown thickness. The mafic and predominantly hornblendic volcanic rocks were named the Dubois Greenstone by Hunter (1925, p. 28). That name was retained on some later geologic maps (Olson and Hedlund, 1973; Hedlund and Olson, 1975), whereas lithologic names were substituted on other maps (for example, Olson, 1976a). The felsic volcanic rocks of the Gunnison region were classed as granite porphyry and as a facies of the Powderhorn Granite by Hunter (1925), but the studies of Olson and Hedlund referred to above established their volcanic origin as lavas and tuffs. In the Needle Mountains, to the southwest of the Gunnison region, a diverse but predominantly amphibolitic assemblage of dark metamorphic rocks was named the Irving Greenstone by Cross and Howe (Cross, Howe, and others, 1905). Barker (1969a) changed the designation to Irving Formation because of the heterogeneous character of the unit. The Irving intertongues to the west with a felsic gneiss that was called the Twilight Granite by Cross and Howe. This unit was recognized to be of metamorphic origin and to be an element of the ancient gneiss complex by Barker (1969a, p. 16), who changed the designation to Twilight Gneiss.

Felsic gneiss of the central Front Range is a leucocratic, granitic-appearing, layered and foliated metamorphic rock. It occurs interlayered on many different scales with both biotitic and hornblendic gneisses. The principal occurrence is in three or four massive layers 300–900 m (1,000–3,000 ft) thick separated by similar thicknesses of other gneisses (Sims and Gable, 1967, p. 7; Braddock, 1969, p. 7). Felsic gneiss also is a component of mixed units consisting of closely interlayered gneisses of differing compositions (Sheridan and others, 1967; Taylor, 1975a). In almost all occurrences, the felsic gneiss is associated with hornblendic gneisses. The major layers of felsic gneiss contain thin layers and lenses of black amphibolite and are bordered in many places by amphibolite or other hornblende gneiss. Mixed gneisses consist most commonly of closely interlayered felsic and hornblendic gneisses.

The composition of the felsic gneiss differs from body to body, but all varieties consist principally of oligoclase, quartz, and microcline, which are accompanied by subordinate biotite and magnetite and locally by hornblende or garnet (Sims and Gable, 1964, table 5; 1967, table 2; Braddock, 1969, table 2). Most of the felsic gneiss has the chemical composition of a siliceous granodiorite, but some has the composition of quartz diorite, or of quartz monzonite. The question of an igneous or a sedimentary parentage of the felsic gneiss has been debated. Sims and Gable (1964, p. 29; 1967, p. 26) argued that the gneiss was of metasedimentary origin because of its layering, its heterogeneity, and its intimate interlayering and gradational contacts with other rocks. They concluded that the gneiss was derived from arkose. However, Braddock (1969, p. 13) showed that normal arkose could scarcely qualify as the protolith and concluded that the source rocks could have been either igneous or sedimentary. Evidence from other areas, discussed in following paragraphs, strongly indicates that the source rocks were tuffs. Tuffs range widely in purity, depending upon the degree of reworking and contamination by sedimentary processes. Many of the features thought by Sims and Gable to indicate a sedimentary parentage of the felsic gneisses more likely reflect either impurity of original tuffs or a fine-scale alternation of tuffs and sediments. The compositions of the felsic gneisses suggest that parent tuffs were principally rhyodacites and quartz latites but ranged from dacite to low-silica rhyolite. The tuff parent of some highly siliceous gneiss probably was contaminated by detrital quartz (Braddock, 1969, p. 13).

Extensive areas of felsic gneiss are present along with interlayered hornblendic gneisses in the northern Park Range (Segerstrom and Young, 1972; Snyder, 1980a, b). The felsic gneiss is leucocratic, pink to pale red, and fine to coarse grained. It is composed of quartz, microcline, oligoclase, and minor amounts of muscovite and biotite. The content of microcline (33–41 percent) and thus of K_2O (4.3–5.0 percent) is more than twice as high as that in the felsic gneiss in the central Front Range. This together with an SiO_2 content of about 76 percent (Segerstrom and Young, 1972, table 1) suggest a rhyolitic parent. Layering within the gneiss, the intimate association with amphibolitic rocks, and a combination of abrupt and gradational contacts with other gneiss all suggest that the rhyolitic rock was mainly tuff.

Other areas provide clear evidence that felsic gneiss was derived from tuffs. Near the Arkansas River in the vicinity of Salida and Howard, interlayered metasedimentary and metavolcanic rocks are extensively exposed (Taylor, Scott, and Wobus, 1975; Boardman, 1976). Primary structures are preserved in both kinds of rocks. The metavolcanic rocks were derived from tuffs, breccias, and flow units ranging in composition from felsic to mafic. The metatuffs display flattened pumice fragments, lithic fragments, and compactional banding that indicate welded tuffs, and associated rocks show features such as graded bedding, crossbedding, and primary breccia structure (Boardman, 1976). The metatuffs are of rhyodacitic composition. They and associated rocks contain staurolite and andalusite and are metamorphosed in the lower amphibolite facies. These rocks grade to more highly deformed and metamorphosed gneisses to the north and to the south. The resulting felsic gneisses in the north closely resemble those of the central Front Range. In the south, metatuff grades to a distinctive gray to pink, even-banded gneiss that persists southward along the Sangre de Cristo Range to the New Mexico border and beyond, suggesting wide distribution of the parent tuff. Another facies, which has been called granodiorite gneiss in the Blanca Peak area of the Sangre de Cristo Range (Johnson, 1969), is coarse grained and strongly foliated but not banded.

Metavolcanic rocks in the Gunnison region are little deformed, and like those of the Salida area, they preserve primary structures. The felsic rocks, present in large bodies, are predominantly metatuffs as shown by eutaxitic structure, recrystallized pumice lapilli, and crystal fragments or phenocrysts (Olson, 1976a; Hedlund and Olson, 1975). The metatuffs have the appearance of fine-grained felsites and felsite porphyries and are mainly quartz latites, though compositions range from dacite to rhyolite. Metabasalt associated with the felsic rocks has pillow structure as well as vesicles and amygdules. Phyllitic rocks interlayered with the volcanics are moderately metamorphosed graywackes and siltstones, some of which retain their clastic textures and show graded bedding. Quartzite in thin beds is inferred to be a recrystallization product of chert (Olson, 1976a; Olson and Hedlund, 1973).

The metavolcanic Twilight Gneiss of the Needle Mountains (Barker, 1969a) is the felsic counterpart of the generally amphibolitic Irving Formation. The two units intergrade, and the Twilight contains many interlayers of amphibolite. The Twilight is a homogeneous leucocratic foliated rock that is characterized by quartz ovoids or "eyes" that probably are recrystallized quartz phenocrysts. The gneiss ranges in composition from quartz diorite to quartz monzonite, and much of it is trondhjemitic. Chemical data are given by Barker (1969a, table 2). Parent rocks of the gneiss were tuffs, flows, and possibly volcanic sills of predominantly trondhjemitic and dacitic compositions. These rocks were subordinate components of a volcanic pile that was largely basaltic, probably in a submarine continental-margin environment (Barker and others, 1969; Barker, Arth, and others, 1976).

Hornblendic gneisses are intimately associated with felsic gneisses as indicated in preceding paragraphs. In the central Front Range, most bodies of hornblendic rocks are small and amphibolitic, but moderately extensive bodies of hornblende gneisses are present in the Montezuma area (Lovering, 1935; Wahlstrom and Kim, 1959). The gneisses of that area are finely banded and consist of various combinations of andesine, hornblende, diopside, quartz, biotite, and epidote. These gneisses are interlayered with smaller amounts of biotite gneiss and quartz gneiss and are locally garnetiferous. Lovering (1935, p. 11) concluded that the hornblende gneisses were derived from lavas and breccias of andesitic composition, but Wahlstrom and Kim (1959, table 1) presented chemical analyses indicating a basaltic composition and postulated a pyroclastic protolith.

Amphibolite and other hornblendic gneisses are widespread in the northern Park Range, where they are interlayered with felsic and biotitic gneisses. The amphibolite is basaltic in composition (Segerstrom and Young, 1972, table 1). Some of it is highly garnetiferous, and varieties characterized by the orthoamphiboles gedrite or anthophyllite also occur, as do staurolitic and cordieritic varieties (Snyder, 1980a). A conglomeratic or agglomeratic facies occupies an area of 2–3 km^2 (1 mi^2) on Farwell Mountain (T. 10 N., R. 84 W., pl. 1). Clasts in the rock are flattened and stretched in the plane of foliation and layering. They consist of very fine grained felsite and are in either a hornblendic or a biotitic matrix.

In the Salida area, some of the hornblendic rocks interlayered with felsic and biotitic rocks have features that indicate sedimentary deposition, such as bedding, crossbedding, and graded bedding. Their composition, however, is basaltic, and the parent sediments evidently were derived from basaltic rocks (Boardman, 1976, p. 92). Other hornblendic rocks are amphibolites, some of which were derived from basalt flows and some from gabbro sills (Van Alstine, 1971). Diabasic and amygdaloidal textures are preserved in some of the rocks, as are remnants of diopsidic pyroxene. Relict breccia structure is well preserved in some of the basaltic flow rocks (Boardman, 1976, fig. 7).

Hornblendic gneisses and related rocks in the Gunnison region, some of which are known as Dubois Greenstone, include amphibolite, hornblende schist, chlorite schist, metabasalt, and greenstone (Olson and Hedlund, 1973; Hedlund and Olson, 1975; Olson, 1976a). Interlayered among these rocks are felsic metatuffs, graywackes, siltstones, and metacherts. This entire suite is at the border between a province of felsic and hornblendic metavolcanic rocks to the southeast and a province of pelitic biotite gneisses to the northwest (pl. 1). The metabasalts locally exhibit pillow structure, amygdules, open vesicles, and primary breccia structure. The amphibolites apparently were derived from volcanic flows, epiclastic sediments, and mafic sills and dikes. Massive greenstones were derived from small bodies of gabbro and diorite that probably were intruded into the volcanic pile as it grew. On the basis of several chemical analyses, Condie and Nuter (1981) classified the entire assemblage of metaigneous rocks as a bimodal sequence of tholeiitic basalts and felsites.

The largely metavolcanic Irving Formation of the Needle Mountains includes many kinds of rocks: amphibolites, mica schist, metagraywacke, meta-andesite, greenstones, metasiltstone, quartzite, conglomerate, and banded iron-formation (Barker, 1969a, p. 5). Amphibolite of basaltic composition is predominant. Some of it has pillow structure, indicating a volcanic origin, but some probably originated as sills and dikes. The conglomerate consists of pebbles and cobbles of quartz and metavolcanic rock in a hornblendic matrix. The iron-formation is a laminated magnetite-quartz rock in beds 10 cm (4 in.) to 15 m (50 ft) thick scattered through a great thickness of other rocks (Steven and others, 1969, p. F43). The iron content of the magnetic rock averages a little less than 15 percent, a grade that is not of commercial interest in deposits of this size. The Irving is the principal component of a metamorphosed volcanic pile that has a western felsic facies, the Twilight Gneiss. In the zone of transition between the two units, a plagioclase-quartz aplite that intrudes amphibolite layers apparently was produced by partial melting of the felsic layers during metamorphism (Barker, 1969a, p. 20).

Near Westcliffe in the Wet Mountains (pl. 1), felsic and hornblendic gneisses reached the threshold of the granulite facies of metamorphism. This area constitutes the principal occurrence in Colorado of rocks of so high a metamorphic grade, the only others being aureoles around small gabbroic bodies in the Front Range. In the Westcliffe area, charnockitic felsic gneisses, characterized by hypersthene, are closely associated with microcline-rich felsic gneisses (Brock and Singewald, 1968; Christman and others, 1959). Interlayered hornblendic gneisses generally contain a pyroxene, either augite or hypersthene, and locally contain both. Some, however, contain biotite, and the suite of rocks as a whole falls a little short of the granulite facies. As in the other areas just discussed, these gneisses probably were derived from volcanic and epiclastic sedimentary rocks, though no direct evidence of the original character has survived the intense metamorphism. The suite of rocks with granulitic affinity is truncated on the east by the Ilse fault (fig. 3) and has not been found anywhere east of the fault.

Metamorphosed mafic intrusive rocks in scattered small bodies in the Sangre de Cristo Range and Wet Mountains are interpreted to be equivalents of the mafic volcanic components of the metavolcanic gneisses. Thus, they were part of the gneiss protolith rather than having been intruded into the gneisses. As described by R. B. Taylor (oral commun., 1982), the rocks are metamorphosed to an amphibolitic composition but they contain remnants or alteration products of pyroxenes and olivine that indicate an original gabbroic composition. The contacts with enclosing felsic and hornblendic gneisses are simple in plan but are slightly blurred by the metamorphic recrystallization of both rocks. In order to distinguish the intrusive-appearing mafic bodies from the metavolcanic gneisses that surround them, they are designated as a metagabbro component of the felsic and hornblendic gneisses in plate 1.

The Precambrian gneisses contain cogenetic mineral deposits in many places. Most of the deposits are small, but their widespread occurrence suggests that undiscovered ore deposits of greater significance may exist in the extensive bodies of felsic and hornblendic gneisses (pl. 1). The known deposits are diverse in character. The most productive known deposits are facies of amphibolites or amphibole gneisses that contain copper and zinc minerals, accessory galena, and small amounts of silver and gold. The largest deposit of this type is at the Sedalia mine in Chaffee County (sec. 18, T. 50 N., R. 9 E.) (Lindgren, 1908; Van Alstine, 1974). A somewhat similar deposit was mined at Cotopaxi in Fremont County (Salotti, 1965), and less productive deposits of similar type occur elsewhere in Fremont, Chaffee, Gunnison, Jefferson, and Custer Counties (Sheridan and Raymond, 1977; Salotti, 1965; Lovering and Goddard, 1950; Lindgren, 1908).

The principal ore minerals in the deposits in gneiss are chalcopyrite and marmatitic (iron-bearing) sphalerite, which are accompanied in several of the deposits by gahnite, the zinc spinel (Sheridan and Raymond, 1977). Magnetite, pyrrhotite, and pyrite are present in many of the deposits, as well as various silicate minerals. Some deposits in the Wet Mountains contain the rare magnesium-iron-aluminum silicate sapphirine. The deposits are in amphibolite pendants in granite, but the sapphirine, gahnite, and sulfide minerals are concluded to have been products of the regional metamorphism, which long preceded the emplacement of the granite (Raymond and others, 1980). A similar regional metamorphic origin has been proposed for various other deposits (Tweto, 1960; Salotti, 1965; Sheridan and Raymond, 1977). By this interpretation, the ultimate source of the metals was in the parent rocks of the gneisses. Lindgren (1908) interpreted the Sedalia and Cotopaxi deposits as magmatic segregations in mafic intrusive rocks that were later metamorphosed, but Salotti (1965, p. 1204) found such an origin to be incompatible with the mineralogy and paragenesis. Lovering and Goddard (1950, p. 64) classed many of the Precambrian deposits as of hypothermal replacement origin and inferred a relationship to nearby granites. Granites may have caused recrystallization and redistribution of elements in some deposits, but as indicated in the more recent studies just cited, the source of the metals probably was in the protoliths of the gneisses, not in the granites.

Mineral deposits of other kinds are also present in the gneisses. In the St. Louis (or Milliken) lead-zinc deposit southwest of Fraser in Grand County, galena and dark sphalerite are disseminated in diopsidic hornblende gneiss that is interlayered with biotite gneisses (Lovering and Goddard, 1950, p. 70). In southwestern Gunnison County, the metals in gold- and gold telluride-bearing replacement veins in the so-called Gunnison gold belt are thought to have been mobilized from deposits related in origin to the Dubois Greenstone wall rocks, into which the veins grade (Hedlund and Olson, 1975; Drobeck, 1981). In some Precambrian tungsten deposits, scheelite is a metamorphic component of hornblende gneiss (Tweto, 1960), and scheelite also accompanies the sulfide minerals in the Cotopaxi deposit (Salotti, 1965). The metals in a nickel deposit at the town of Gold Hill in Boulder County may have been introduced from a magmatic source during or immediately following metamorphism. The deposit, which consists of an amphibolite layer containing nickeliferous pyrrhotite and chalcopyrite, is bordered by a small body of quartz gabbro. The gabbro, which is older than the 1,700-m.y. Boulder Creek Granodiorite, was interpreted to be the source of the metals by Lovering and Goddard (1950, p. 70). Both the gabbro and the nickel deposit are cut by an unmineralized diabase dike (the "Iron Dike") of younger Precambrian age. (See section on mafic and intermediate dikes.)

VALLECITO CONGLOMERATE AND UNCOMPAHGRE FORMATION

In the Needle Mountains of southwestern Colorado, a thick body of quartzite, slate, and phyllite forming the Uncompahgre Formation overlies the Irving Formation and Twilight Gneiss of the Early Proterozoic gneiss complex. Nearby, an area between two bodies of granite is occupied by conglomerate and quartzite composing the Vallecito Conglomerate (pl. 1). Stratigraphic and structural relations between the Vallecito and the Irving have been interpreted differently by different workers, and this has led to differences in age assignment and stratigraphic classification of the Vallecito. In defining the formations, Cross and coworkers (Cross, Howe, and others, 1905; Larsen and Cross, 1956) interpreted the Vallecito-Irving contact everywhere as a fault, and they recognized a compositional similarity in the Vallecito and Uncompahgre Formations. Therefore, they regarded the Vallecito as related to but older than the Uncompahgre, and placed it stratigraphically above the Irving. They also established the Needle Mountains Group, consisting of the Uncompahgre Formation above and the Vallecito Conglomerate below, though the two formations are nowhere in contact and the assumed stratigraphic relation was conjectural.

Later, Barker (1969a, p. 4) interpreted the Irving-Vallecito contact as depositional in most places. He also interpreted crossbedding in the Vallecito as indicating that the tops of beds face toward the Irving. Consequently, he concluded that the Irving stratigraphically overlies the Vallecito, making the Vallecito the oldest unit in the Precambrian of the Needle Mountains. This assignment voided the concept of a Needle Mountains Group because it placed the Irving Formation and major granitic intrusions stratigraphically between the two formations of the group. Therefore, the term Needle Mountains Group was abandoned (Tweto, 1977). Subsequently, after the most intensive study of the Vallecito yet made, Burns and others (1980) concluded that, contrary to Barker, the Vallecito stratigraphically overlies the Irving, as indicated from crossbedding and from the occurrence of conglomerate of the Vallecito in channels cut into the top of the Irving. Other considerations support this conclusion as noted in the discussion of the Vallecito.

In this report, the Vallecito is classed as being younger than the Irving Formation and as an equivalent or near-equivalent of the Uncompahgre Formation. This is a reversal of the changes in classification made by Barker (1969a) and returns to the original classification of Cross and coworkers, except that a Needle Mountains Group is not recognized. Such a group would be valid only if the Vallecito were proved to occupy a different stratigraphic position than the Uncompahgre. From what is known at present, however, the Vallecito is as likely to be a lateral facies of the Uncompahgre as it is to be an underlying unit. Until the stratigraphic relationship may be clarified, reinstatement of the Needle Mountains Group is inappropriate.

VALLECITO CONGLOMERATE

The Vallecito Conglomerate is a massive unit of conglomerate, pebbly quartzite, and quartzite at least 1,525 m (5,000 ft) thick. The highly resistant rocks of the formation are exposed in an area of about 50 km^2 (20 mi^2) in the southeastern Needle Mountains in the vicinity of the boundary between Hinsdale and La Plata Counties (pl. 1). Burns and others (1980) recognized four lithic divisions within the formation. From the base upwards, these are (1) a foliated conglomerate unit, characterized by deformed clasts and a schistose matrix in which foliation parallels the bedding; (2) a boulder conglomerate unit containing boulders as large as 1 m (39 in.) in diameter; (3) a pebble conglomerate unit; and (4) a quartzite and pebbly quartzite unit that contains local beds of phyllite and schist. All the conglomerate units contain beds and lenses of quartzite or pebbly quartzite. Clasts in the conglomerates are principally vein quartz, quartzites of several varieties, chert, jasper, and banded iron-formation. Also present in conglomerates near the base of the formation are amphibolite, schist, gneiss, and greenstone. The matrix of the conglomerates is a sericitic quartzite that locally contains andalusite and sillimanite, which are attributed to contact metamorphism by adjoining bodies of Eolus Granite.

Bedding units in the Vallecito are thick and commonly are separated by scour surfaces or local unconformities. Trough crossbedding is visible in almost all the quartzite beds, where it is outlined by hematite films. Burns and others (1980) concluded from the distribution and orientations of compositional and sedimentational features that the Vallecito is an alluvial fan deposit that prograded southward from a source area immediately north of the present area of exposure.

In their detailed studies, Burns and others (1980) found the Vallecito to be in fault contact with the Irving in most places, but they found the contact to be clearly depositional in one area. According to them, crossbedding in that area unequivocally indicates that the tops of beds in the Vallecito face away from the Irving. They also found (Burns and others, 1980, p. 30) that conglomerate of the Vallecito fills scour channels in the top of the Irving. Despite Barker's (1969a) conclusions to the contrary, they concluded that the Vallecito lies on and is younger than the Irving. Other considerations support this conclusion. The basal conglomerate of the Uncompahgre Formation contains the same assortment of lithologies among its clasts as the Vallecito, suggesting that the two formations were joint products of a single sedimentational episode. The Irving Formation is varied in composition, and it could have been the source of all the varieties of clasts found in the Vallecito. Larsen and Cross (1956, p. 24) thought the Vallecito to have been "clearly derived from the Irving" and hence to be the younger of the two. The Vallecito is less deformed than the Irving, as noted by Burns and others (1980, p. 75). Barker (1969a, p. 29), having interpreted the Vallecito as being older than the Irving, postulated an inverted anticline to account for map relations between the two formations. Such extreme deformation is inconsistent with structure observed in the body of Vallecito rocks, and the need for it is unnecessary if the sequence concluded by Burns and coworkers is correct.

In summary, though contradictory interpretations of some features have been made, various lines of evidence indicate that the Vallecito is younger than the Irving Formation and that it probably is closely related in age and origin to the Uncompahgre Formation. As with the age of the Uncompahgre, the age of the Vallecito is known only within broad limits: it is younger than the Irving Formation (about 1,800 m.y.) and older than the Eolus Granite (about 1,440 m.y.). The stratigraphic relationship between the Vallecito and the Uncompahgre also remains undetermined. The Vallecito might be the older, or it might be a local and contemporary near-source facies of the Uncompahgre. A first requirement in appraising these possibilities is a detailed structural study of these formations together with the Irving Formation, yet to be made.

The possibility that the Vallecito might contain fossil placer deposits of gold or uranium has been investigated. The results were negative or nearly so. On the basis of six composite samples from various localities, Barker (1969b) reported that, within the limits of detectability, gold was absent from the Vallecito. Burns and others (1980) reported an average uranium content of 2.9 ppm in 502 samples from the Vallecito and the basal conglomerate of the Uncompahgre Formation. The maximum concentration found in the Vallecito was 11 ppm. Uranium content greater than 2.2 ppm was limited largely to the pebble conglomerate facies.

UNCOMPAHGRE FORMATION

The Uncompahgre Formation is a thick sequence of quartzites, slates, and phyllites exposed in scattered localities in the vicinity of the western San Juan Mountains in southwestern Colorado. The principal area of exposure is in the Needle Mountains, but extensive exposures occur also south of Ouray in the canyon of the Uncompahgre River, the type area of the formation. Smaller occurrences are in an upthrown fault block at Rico and in the canyon of the Piedra River northwest of Pagosa Springs (pl. 1). In the type area, where neither the base nor the top of the formation is exposed, a thickness of about 2,650 m (8,700 ft) is visible in the nose of a large steeply plunging anticline (Kelley, 1946, p. 296; Cross, Howe, and Ransome, 1905). In the Needle Mountains, at least 2,440 m (8,000 ft) is preserved in an arcuate synclinorium about 35 km (22 mi) long (Barker, 1969a; Cross, Howe, and others, 1905). In that area the Uncompahgre lies unconformably upon gneisses of the Irving and Twilight Formations. [Editor's note: An article published by B. J. Tewksbury (Revised interpretation of the age of allochthonous rocks of the Uncompahgre Formation, Needle Mountains, Colorado: Geological Society of America Bulletin, v. 96, p. 224–232, 1985) subsequent to Ogden Tweto's death in November 1983, indicates that the contact between the Uncompahgre and the Irving and Twilight Formations in the Needle Mountains area is a structural discontinuity, probably a thrust fault.] The unconformity also truncates dikes of a granite that is dated as about 1,695 m.y. in age (Barker, 1969a, p. 22). The formation is cut by Eolus Granite, about 1,440 m.y. in age (Silver and Barker, 1968; Bickford and others, 1969), and thus its age is bracketed between 1,440 and 1,695 m.y. Disregarding the Vallecito Conglomerate, no other supracrustal rocks in this age bracket are known in Colorado or nearby in adjacent States except possibly the Texas Gulch Formation in central Arizona (Anderson and others, 1971).

The Uncompahgre Formation consists of thick units of light-colored quartzite alternating with thinner units of dark slate and phyllite. A basal conglomerate present in most places in the Needle Mountains has a maximum thickness of 6 m (20 ft) but generally is much thinner. Clasts in the conglomerate resemble those in the Vallecito Conglomerate and have a maximum diameter of about 10 cm (4 in.) (Barker, 1969a, p. 11). In some places a fault zone subparallel to the bedding separates the Uncompahgre from underlying rocks. This zone is marked by quartz-sericite phyllonite in which foliation parallels the bedding in adjacent rocks (Barker, 1969a, p. 15). The quartzites of the Uncompahgre in the Needle Mountains and at Rico are rather coarse and uneven grained and contain scattered beds of pebbly quartzite or conglomerate (Barker, 1969a, p. 11; McKnight, 1974, p. 10). Those of the Uncompahgre River area are finer and more uniform in grain size, though they also contain a few pebbly layers (Luedke and Burbank, 1962). The slates and phyllites commonly are finely banded or laminated and show graded bedding. As described by Barker (1969a, p. 12), those in the Needle Mountains show progressive increase in grade of metamorphism eastward toward a batholith of Eolus Granite that truncates the synclinorium of Uncompahgre rocks. In the west, the pelitic rocks are quartz-sericite slates containing accessory pyrite and carbon. Eastward, these rocks become phyllitic and then schistose, and successive metamorphic assemblages follow characterized by biotite, garnet, andalusite, staurolite, and finally sillimanite. Staurolite and andalusite have been noted also in the Uncompahgre River area (Cross, Howe, and Ransome, 1905, p. 3).

The pyritic black slates of the Uncompahgre have been investigated for their metal content, and the conglomeratic quartzites have been investigated for gold. The contents of metals and other trace elements in the slates are typical of that in many black shales (Steven and others, 1969, p. 50). Gold was not detected in the conglomerates (limit of detection 0.1 ppm) nor in the slates (limit of detection 0.02 ppm) (Barker, 1969b, p. 10–14). On the basis of numerous samples, the basal conglomerate contains a maximum of 5.1 ppm uranium, and the black slates contain as much as 330 ppm (Burns and others, 1980). Carbon content of the slates ranges from 0.15 to 6.4 percent and in general decreases with increasing metamorphic rank. The carbon in the more highly metamorphosed rocks is in the form of graphite and is isotopically heavier than that in the slates (Barker and Friedman, 1969).

The extent of the Uncompahgre Formation in the subsurface is inferred from the scattered exposures, from a few boreholes, and from the occurrence of its rocks as xenoliths in Tertiary intrusive bodies. Insofar as it is known, the formation is preserved in three areas alined in a belt trending northwest (pl. 1). A subcircular northwestern area, defined by xenoliths and one borehole, borders the area of exposure along the Uncompahgre River. A central area consists of the synclinorium in the Needle Mountains, a belt extending westward from the synclinorium to or beyond Rico, and the body of Vallecito Conglomerate, which here is interpreted as an equivalent of the Uncompahgre. A southeastern area is defined by the small exposures along the Piedra River and by boreholes near the New Mexico border. This body of the Uncompahgre is inferred to extend only a few kilometers into New Mexico. The separation of the Uncompahgre into the three areas is due in part to intrusion by the Eolus Granite, but it is due mainly to erosion before the Phanerozoic covering units were deposited. The erosion left the Uncompahgre preserved only in downwarps, and thus the original thickness and extent of the formation are unknown. The preserved bodies lie along the southwest side of the late Paleozoic and early Mesozoic Uncompahgre highland, which was deeply eroded before a cover of Triassic or Jurassic rocks was deposited over it (Tweto, 1983a). Thus, the Uncompahgre Formation could have existed earlier at the site of the highland. On the other hand, the coarse and pebbly quartzites and the conglomerates suggest a local source of sediment. The evidence afforded by the Vallecito Conglomerate indicates that the source was to the north (Burns and others, 1980) in the direction of the Uncompahgre highland of later time. Until information to the contrary may be obtained, it is assumed that the Uncompahgre-Vallecito sedimentation occurred in a subsiding synclinal basin of northwest trend. The basin may have been asymmetric, with an axis near the northeast border, in the area of the preserved bodies of the formation.

LAS ANIMAS FORMATION

The Middle Proterozoic Las Animas Formation, defined by Tweto (1983b), is a thick sequence of moderately metamorphosed sedimentary and volcanic rocks that lies beneath the plains of southeastern Colorado (pl. 1). The presence of these rocks beneath Paleozoic rocks was first noted by Heaton (1933, p. 136) and was later discussed briefly by Maher and Collins (1949). Knowledge of the formation comes from several boreholes, two of which penetrated deeply into the sequence. As defined by the boreholes, the formation occupies a west-trending belt about 130 km (80 mi) long and as much as 50 km (30 mi) wide that lies near the crest of the buried Apishapa highland of Pennsylvanian age. A deep borehole near the west end of the belt (Lange 1 Government, sec. 10, T. 29 S., R. 62 W.) penetrated 1,270 m (4,165 ft) into the formation without reaching the base. A borehole 60 km (37 mi) to the east (Phillips 1 Haskins, sec. 23, T. 29 S., R. 56 W.)

penetrated 431 m (1,415 ft) into a different and stratigraphically higher part of the formation. Total thickness of the formation is unknown but probably much exceeds the 1,700 m (5,575 ft) penetrated in the two boreholes combined.

At the site of the Lange 1 Government borehole, the Las Animas Formation consists of dark-gray and black slate, phyllite, fine-grained graywacke, and chert. These rocks are in thick units that show little variation through many meters. The slates and graywackes are markedly siliceous, and both grade in places into chert. The slates give way downward to phyllites, and the phyllites become coarser and also biotitic and hornfelsic with depth. At the site of the Haskins borehole, the Las Animas contains volcanic and carbonate rocks as well as slate, phyllite, and graywacke, and the upper part of the explored section is dark red or maroon. The volcanic rocks are basalt and andesite. They occur as scattered layers 2 or 3 m (6–10 ft) thick among the sedimentary rocks in the red part of the section. The carbonate rocks are mainly gray limestones in beds 0.5–3 m (1.5–10 ft) thick in the gray lower part of the explored section, but pink to red dolomite is present in thin beds in the red part of the section. In other boreholes nearby and farther east, latite, andesite, basalt, and dacitic to rhyolitic devitrified tuffs were encountered (Edwards, 1966) as were red, pink, and buff quartzites.

Age of the Las Animas Formation is established directly only within broad limits, but indirect evidence suggests narrower limits. The map pattern of Precambrian rocks in subsurface southeast Colorado (pl. 1) indicates that the formation lies on granite of the 1,400-m.y. Berthoud Suite and on older gneisses. Therefore, the Las Animas is younger than about 1,400 m.y., and, as indicated by its position below rocks as old as Late Cambrian, it is older than Late Cambrian. The Las Animas closely resembles the Tillman Metasedimentary Group of Ham and others (1964) in the Wichita Mountains province of southern Oklahoma, and it probably is of the same age as the Tillman. The Tillman has been interpreted as the basal fill in a deep structural trough (Ham and others, 1964) or aulacogen (Hoffman and others, 1974), though more recent findings suggest a broader but deep structural basin (Brewer and others, 1981). The Las Animas Formation is interpreted as a filling in a trough of rift origin in a terrane of 1,400-m.y. mesozonal granites. Though not connected or alined, the troughs of the Las Animas and the Tillman have similar trends, and they were almost certainly responses to the same tectonic event. This being so, the age of the Las Animas probably is similar to that of the Tillman, which, on the basis of indirect evidence, is thought to be at least 1,200 m.y. and less than 1,400 m.y. (Brewer and others, 1981, p. 574).

UINTA MOUNTAIN GROUP

The Middle Proterozoic Uinta Mountain Group is a thick sequence of clastic rocks that forms the core of the Uinta Mountains of Colorado and Utah and also occupies a substantial area in the subsurface to the east of the mountains (pl. 1). Rocks of the group lie upon highly metamorphosed and deformed Red Creek Quartzite and Early Proterozoic and Archean gneisses, but are themselves essentially unmetamorphosed and only locally deformed. Because of their lack of metamorphism, they were thought to be younger than Precambrian by early workers (for example, Powell, 1876). Hansen (1965) has summarized the history of stratigraphic classification and described the group as it is exposed in the eastern Uinta Mountains near the Colorado-Utah border. In that area, the group consists of more than 7 km (4 mi) of quartzite and siliceous sandstone accompanied by thick wedges of conglomerate and subordinate shale, all predominantly dull red. Hansen (1965, p. 35–36) thought that clasts of metaquartzite and amphibolite in conglomerates in the lower part of the group were derived from the Red Creek Quartzite. He interpreted the conglomerates as alluvial fan deposits that thinned southward and westward from sources to the north. Farther west, in the western Uinta Mountains, arkosic and deltaic units are similarly oriented, indicating sources to the north (Crittenden and others, 1967; Wallace and Crittenden, 1969; Wallace, 1972; Sanderson, 1978). Rocks of the group are of fluvial origin in the east. To the west, in Utah, they are of shallow-marine character south of a west-trending strand line (Wallace and Crittenden, 1969; Sanderson, 1978).

Age of the Uinta Mountain Group ranges from about 925 to 1,100 m.y. as determined by the Rb-Sr whole-rock method on samples of shale from several localities in Colorado and Utah (Chaudhuri and Hansen, 1980). This finding is consistent with an earlier determination of 952 m.y. for shale near the top of the group (Crittenden and Peterman, 1975). The age determinations establish that the Uinta Mountain Group is correlative with the upper part of the generally similar Belt Supergroup of Montana and Idaho, with much of the upper part of the Grand Canyon Supergroup in Arizona, and with much of the Big Cottonwood Group of the Wasatch Mountains in Utah (Harrison and Peterman, 1984). According to paleomagnetic data, however, the Uinta Mountain Group is younger than most of the Belt Supergroup (Bressler, 1981; Elston and McKee, 1982).

The geographic extent of the Uinta Mountain Group beyond the areas of exposure is problematic. A few boreholes indicate that the group extends eastward in the subsurface at least to the longitude of Craig. Boreholes also establish that the group is present beneath the sedimentary rocks on the south flank of the Uinta Mountains. Otherwise, the subsurface extent of the group is open to conjecture. On the basis of aeromagnetic data, Steenland (1969) concluded that the Uinta Mountain Group is absent in the Uinta and Green River basins, south and north, respectively, of the western Uinta Mountains in Utah (fig. 1), and that the group is confined to the area of the present mountains. He postulated further (p. 59) that the group was "deposited in a very narrow basin, probably a rift." Farther east, in the area described by Hansen (1965), the conglomerate wedges of alluvial fan origin suggest that the north border of the depositional area was not far from present exposures on the north flank of the Uinta Mountains, and that the border may have been a fault scarp. No direct evidence of the location of a corresponding southern border is known except in the subsurface at a locality some 80 km (50 mi) east of the Uinta Mountains.

The body of rocks of the Uinta Mountain Group in the subsurface east of the Uinta Mountains is only poorly delineated. On the north, this body is inferred to be in fault contact with Archean rocks from the point where those rocks enter the State near long 108° W. westward to the Uinta Mountains (pl. 1). At the mountains, the line of fault contact disappears beneath the upper plate of the Laramide thrust fault that borders the north flank of the mountains. This thrust fault places the Uinta Mountain Group and older rocks over various Phanerozoic formations and has been proved by drilling to have a horizontal displacement of at least 5 km (3 mi). On the south side of the subsurface body, a fault boundary between the Uinta Mountain Group and older Proterozoic gneisses is also inferred. The location of such a fault boundary is closely fixed at a locality near the southeast corner of Moffat County. A borehole in sec. 14, T. 4 N., R. 90 W., the Miami Oil Producers No. 1 O'Brien, penetrated 140 m (462 ft) of sandstone, shale, and siltstone of the Uinta Mountain Group beneath the Paleozoic rocks, whereas a borehole in sec. 35, T. 4 N., R. 89 W., the Pacific National-Southern Union Production No. 26–35 Pagoda Unit, entered gneiss below the Paleozoic rocks. A point between the two holes is on the line of projection of a major northeast-trending Precambrian fault zone in the Park Range, the Soda Creek-Fish Creek mylonite zone (Snyder, 1980b). This fault zone underwent recurrent movements in the Precambrian (Snyder, 1978, p. 2), and its extension to the southwest is inferred to have served as the southern boundary of the trough in which eastern elements of the Uinta Mountain Group accumulated.

Location of the east end of the eastward-tapering body of the Uinta Mountain Group is uncertain. Originally, the group probably thinned eastward to a depositional margin near the west side of the present Park Range. Later, the margin migrated westward by erosional truncation following late Precambrian and late Paleozoic uplifts. The easternmost proved occurrence of the group is in the Miami Oil Producers well in R. 90 W., mentioned previously. In plate 1, the erosional edge is speculatively drawn through the town of Mount Harris in R. 87 W. A margin was drawn farther east by Edwards (1966, pl. 1) on the assumption that the Uinta Mountain Group was present in the Texaco No. 1 Colvert well, sec. 7, T. 6 N., R. 86 W., and the Texaco No. 1 Peavey well, sec. 27, T. 8 N., R. 86 W. Both wells are in an area of late Paleozoic uplift and the erosional truncation of pre-Pennsylvanian Paleozoic formations. Green shale and gray quartzite in which the Colvert well bottomed are interpreted here as a downwarped remnant of Devonian or Cambrian rocks. Quartzite in which the Peavey well bottomed is notable for a content of as much as 10 percent of sphene (Edwards, 1966, p. 337) and is accompanied by fragments of gneiss, schist, and arkose. It is interpreted as a part of the gneiss complex, probably in a fracture zone.

The Uinta Mountain Group evidently was deposited in a subsiding fault-bounded trough of east-west trend, and the principal source of sediments was from the north. As noted in the section on Archean rocks, the trough coincided approximately in location and trend with an extensive segment of the boundary between the Archean Wyoming province and the Proterozoic age province of the region to the south. In one of the periods of recurrent tectonism along this boundary, rifting produced the structural trough in which the Uinta Mountain Group accumulated. This trough was referred to as an aulacogen by Sears and others (1982).

PROTEROZOIC INTRUSIVE ROCKS

GROUPINGS OF IGNEOUS ROCKS

Igneous rocks of various kinds and ages constitute about 50 percent of the exposed Precambrian basement in Colorado and an approximately equal proportion of the buried basement (pl. 1). The rocks are principally granitic, but they range in composition from granite to gabbro. The major varieties and most of the major bodies have been dated radiometrically by one or more methods, and others are classified by their geologic relationship to the dated units. The age determinations fall statistically into three populations centered near 1,700, 1,400, and 1,000 m.y. The rocks of each of these age categories are products of different stages in the development of the Precambrian basement in Colorado, and, therefore, they differ from one another in character and habit. The rocks that are about 1,700 m.y. in age are catazonal and synorogenic in habit. They are concordant, commonly foliated, and among the granitoid facies, mainly granodioritic to quartz monzonitic. Rocks that are about 1,400 m.y. in age are mesozonal and anorogenic in habit. They are discordant in some part, unfoliated or only locally foliated, and, among the granitoid facies, mainly quartz monzonitic to granitic in composition. Rocks that are about 1,000 m.y. in age are anorogenic, discordant, unfoliated, and highly differentiated.

The rocks of many of the major igneous bodies bear formal stratigraphic names. None of the names is properly applicable as a collective term for all the rocks in an age category. In the absence of collective terms, phrases such as "igneous (or granitic) rocks of the 1,700-m.y. age group" have been used (Tweto, 1977, 1979). A collective term is desirable, not only for convenience of reference but also to discourage the application of established names to rocks that may not be equivalent. Accordingly, and as provided in the revised North American Stratigraphic Code (North American Commission on Stratigraphic Nomenclature, 1983, Art. 35), the term Routt Plutonic Suite is applied in a following section to the rocks of the 1,700-m.y. age group. The term Berthoud Plutonic Suite is applied to the rocks of the 1,400-m.y. age group. The rocks of the 1,000-m.y. age group are all in a single batholith and one outlier. They are referred to as "rocks of the Pikes Peak batholith."

ROUTT PLUTONIC SUITE

The numerous bodies of igneous rocks about 1,700 m.y. in age in Colorado (pl. 1) are related in character, habit, structure, and tectonic setting as well as in age. They are the oldest rocks intrusive into the gneisses, and together they constitute a compound lithodemic unit for which the name Routt Plutonic Suite is here proposed. Rocks of the suite are predominantly granodiorite and quartz monzonite, though compositions range from granite to gabbro. The suite takes its name from Routt County, Colo., where rocks characteristic of the unit are exposed extensively in the Park Range and northern Gore Range (fig. 2). These ranges provide a transect through the middle part of the northeast-trending Rawah batholith (fig. 3). This large batholith exposed in part in the Park and Gore Ranges and Medicine Bow Mountains serves as the type area for the Routt Suite, but other plutons of somewhat different character, discussed in following pages, must also be considered reference localities.

Igneous bodies of the Routt Plutonic Suite were introduced into the Proterozoic gneiss complex during and immediately following the folding and metamorphism of the complex. Two generations of major folds are recognized in the gneisses of the Front Range, and in some places minor folds of a third generation are evident. Granodioritic and quartz monzonitic igneous bodies of the Routt Suite were emplaced late in the second period of folding (Gable and Sims, 1970, p. 11; Taylor, 1975a). Diorite, gabbro, and hornblendite in small bodies accompany the granodiorite. Most bodies of these rocks in the central Front Range are younger than at least some of the granodiorite (Harrison and Wells, 1959, p. 15; Taylor and Sims, 1962), but some bodies are older than the granodiorite (Gable, 1980a, p. 20–22). A comparatively large body of gabbro in the northern Park Range predates the granodioritic rocks (Snyder, 1980a) as do many other bodies of dioritic and gabbroic rocks elsewhere (Tweto and Lovering, 1977, p. 13).

The Routt Plutonic Suite consists of the formally named rocks of several intrusive bodies and the unnamed, or informally named, rocks of numerous other bodies. Currently recognized named lithodemes contained in the suite are the Boulder Creek Granodiorite (or Granite) and Twin Spruce Quartz Monzonite of the central Front Range; the Cross Creek Granite (or Quartz Monzonite) of the northern Sawatch Range and Gore Range; the Denny Creek Granodiorite of the Sawatch and Mosquito Ranges; the Kroenke Granodiorite of the central Sawatch Range; the Powderhorn Granite and Pitts Meadow Granodiorite of the Gunnison region; and the Bakers Bridge and Tenmile Granites of the Needle Mountains.

ROCKS OF THE RAWAH BATHOLITH

A large body of granitic rocks about 1,700 m.y. old in the Medicine Bow Mountains and adjoining west slope of the Front Range was named the Rawah batholith by McCallum and others (1975) and was further described by McCallum and Hedge (1976). As shown in plate 1, the body in the Medicine Bow area is only a small part of a much more extensive body of granitic rocks that is inferred to extend southwestward at least 200 km (125 mi) to the Grand Hogback north of Rifle. This entire body is here called the Rawah batholith (fig. 3). Exposures in the Park and northern Gore Ranges provide a transect across the full width of the batholith in its middle part, and small areas of exposure on the White River Plateau are in the southwestern part. Continuity of the batholith between areas of exposure is established by several boreholes in North Park and in the synclinal area between the Park Range and the White River Plateau. As interpreted from exposures and boreholes (pl. 1), a major arm of the batholith branches eastward from the southeast side in the Park Range. This arm extends across Middle Park and into the flank of the Front Range. A hook-shaped body of granodiorite and the Boulder Creek batholith lie to the east on the same trend (pl. 1). Thus, these intrusives seem to be distal expressions of the east-trending arm of the Rawah batholith, and at greater depth they might be continuous with the arm.

The Rawah batholith is compositionally and texturally varied as seen in the Medicine Bow Mountains (McCallum and Hedge, 1976; Pearson and others, 1982; Griswold, 1980). Composition of the rocks ranges from quartz diorite to granite but is predominantly quartz monzonite. Textures range from coarse-grained porphyritic and pegmatitic to aplitic. Most of the rock has a primary flow foliation expressed by the orientation of microcline phenocrysts, and a parallel or subparallel secondary foliation expressed by the orientation of biotite and planes of cataclasis. Inclusions as much as hundreds of meters (yards) in length are distributed widely in the granitic rocks. They consist of wall-rock gneisses and of prebatholithic igneous rocks. The wall-rock gneiss inclusions are amphibolites, sillimanitic biotite gneisses, and quartz gneisses. The igneous rocks are in two age groups, an early mafic group consisting of metadiorite, metagabbro, and metapyroxenite, and a younger but prebatholithic group consisting of quartz diorite, tonalite, and trondhjemite (Griswold, 1980).

Rocks of the Rawah batholith in the Park and northern Gore Ranges differ little from those in the Medicine Bow Mountains. They have been described by Snyder (1978, p. 12–13; 1980c), who referred to them as the quartz monzonite and granodiorite of Buffalo Mountain. The principal facies is a moderately foliated coarse-grained porphyritic quartz monzonite containing abundant pink prismatic potassium feldspar crystals 3–5 cm (1.2–2 in.) long. Modal and chemical data are in a report by Snyder (1978, table 5). Among the subordinate facies are a cataclastic or augen gneiss, an even- and medium-grained variety, migmatitic varieties, and various aplitic, pegmatitic, and alaskitic varieties. As in the Medicine Bow Mountains, inclusions are abundant. Inclusions mapped by Snyder (1980c) consist of wall-rock gneisses and prebatholithic mafic igneous rocks. Among the gneiss inclusions, the largest are metamorphosed marbles consisting of combinations of calc-silicate rocks and amphibolite. The mafic igneous inclusions are peridotite, hornblendite, gabbro, and diorite. Hornblendite, gabbro, and diorite occur also as small dikes or irregular bodies cutting the granitic rocks.

Rocks of the batholith exposed in the deep canyons of the White River Plateau are of the same character as those in the mountain ranges to the northeast. They have been examined only in reconnaissance and are principally a rather dark granodiorite that is medium to coarse grained and weakly to strongly foliated. A lighter gray porphyritic quartz monzonite and a few lenses of dark-gray quartz diorite are also present. All varieties contain scattered inclusions of wall-rock gneisses and are in concordant contact with bordering gneisses.

The Rb-Sr age of the granitic rocks in the Rawah batholith is 1,665–1,675 m.y. as determined by C. E. Hedge from samples from the Medicine Bow Mountains (McCallum and Hedge, 1976), the Park Range (Snyder, 1980c), and the White River Plateau (C. E. Hedge, written commun., 1975). Some of the subordinate mafic rocks are older than the granitic rocks as determined from geologic relations, and some are younger.

ROCKS OF THE BOULDER CREEK BATHOLITH

The Boulder Creek batholith has been studied intensively and constitutes an important reference locality for the Routt Plutonic Suite. The batholith occupies an area of about 375 km^2 (145 mi^2) west of the mountain front near Boulder (fig. 3) and an area of unknown extent beneath the sedimentary rocks east of the front. The batholith consists of two main rock units, the Boulder Creek Granodiorite and a subordinate and slightly younger facies, the Twin Spruce Quartz Monzonite. The granodiorite was originally described as Boulder Creek Granite Gneiss by Boos and Boos (1934, p. 305) and as Boulder Creek Granite by Lovering and Goddard (1938, 1950) and Lovering and Tweto (1953). Later, it was described more accurately as Boulder Creek Granodiorite (Wells, 1967; Gable, 1980a). The quartz monzonite was first mapped by Lovering and Goddard (1938, 1950), who included it with felsic gneiss of the metamorphic complex in the unit they called "granite gneiss and gneissic aplite." The rock was referred to simply as quartz monzonite by Wells (1967) and by Sheridan and others (1967) and was named the Twin Spruce Quartz Monzonite by Gable (1980a, p. 26).

The Boulder Creek Granodiorite is coarse grained, and where locally coarsely porphyritic, it contains chunky prismatic potassium feldspar crystals 1–5 cm (0.4–2 in.) long. Biotite is fairly abundant and commonly is in clots or packets of flakes. Hornblende is a minor constituent in much of the rock but is abundant locally. Sphene is a characteristic and fairly conspicuous accessory mineral. The rare-earth mineral allanite, more or less altered to epidote, is also a common accessory mineral. According to one interpretation, the allanite was introduced into the rock during the magmatism that produced the Berthoud Plutonic Suite (about 1,400 m.y.) (Hickling and others, 1970, p. 1981), but a more conventional interpretation is that most of it crystallized from the Boulder Creek magma, though some probably is secondary (Gable, 1980a, p. 82). Numerous modes and chemical and spectrographic analyses of the Boulder Creek and the associated Twin Spruce Quartz Monzonite may be found in a comprehensive report on the batholith by Gable (1980a, tables 1–5, 10–13).

The Boulder Creek Granodiorite near the margins of the batholith is strongly foliated and is contaminated with partly assimilated wall-rock gneiss. The foliation parallels the batholith contact and the foliation in bordering gneisses. The rock in the interior of the batholith is only weakly foliated. Rocks at the south end of the batholith have a cataclastic foliation in and near the Idaho Springs-Ralston shear zone (fig. 3) (Tweto and Sims, 1963).

The Twin Spruce Quartz Monzonite forms numerous irregular bodies in the southern part of the Boulder Creek batholith (Gable, 1980a, pl. 1). The quartz monzonite is more leucocratic and finer grained than the Boulder Creek Granodiorite. Most of it is slightly younger than the Boulder Creek, but some seems to have been emplaced before the Boulder Creek solidified (Gable, 1980a, p. 27).

Mafic rocks related to the Boulder Creek Granodiorite include diorite, gabbro, hornblendite, and pyroxenite. They occur with granodiorite in numerous small satellitic bodies in the gneisses bordering the batholith, and also in a few small bodies within the batholith. The diorites are mostly hornblendic, and some contain clinopyroxene. The hornblendites commonly contain clinopyroxenes, and most of the gabbro bodies contain both clino- and orthopyroxenes. Modes and chemical and spectrographic analyses of the mafic rocks are given by Gable (1980a, tables 7, 12). The largest body of gabbro, described by Taylor and Sims (1962), is in the Central City area southwest of the batholith.

Age of the Boulder Creek Granodiorite has been firmly established by means of numerous radiometric determinations and serves as a standard for comparison among units of the Routt Plutonic Suite. On the basis of U-Pb determinations on zircon in samples from six localities and facies in the batholith, Stern and others (1971) fixed the age of emplacement at 1,700 m.y. On the basis of Rb-Sr determinations on several samples from the batholith and on granodiorite and diorite in satellitic bodies, Peterman and others (1968) calculated an age of $1,665 \pm 40$ m.y. They also reported a mineral isochron that indicates a subsequent thermal disturbance at about $1,315 \pm 70$ m.y. Within the limits of discrimination, the Rb-Sr age of the Twin Spruce Quartz Monzonite as determined by C. E. Hedge is the same as that of the Boulder Creek Granodiorite (Gable, 1980a, p. 35).

ROCKS OF THE MOUNT EVANS BATHOLITH

Foliated granodiorite in a small batholith in the Mount Evans area, about 16 km (10 mi) southwest of the Boulder Creek batholith (fig. 3), closely resembles and is assigned to the Boulder Creek Granodiorite. The rock has had a complex history of nomenclature. It has been called Boulder Creek Granodiorite (or Granite, or Granite Gneiss) by several authors (Bryant and Hedge, 1978, p. 448; Boos, 1954, p. 119; Lovering and Goddard, 1950, p. 27), but it has also been called "Archean quartz monzonite" (Ball, 1906), "Mount Evans Quartz Monzonite" (Boos and Aberdeen, 1940; Göbel and Hutchinson, 1971), and "Pikes Peak Granite" (Lovering, 1929). The history of nomenclature and the confusion caused by the misidentification as Pikes Peak Granite have been reviewed by Tweto (1977, p. 14) and by Bryant and Hedge (1978, p. 447–448).

A Rb-Sr age of $1,660 \pm 60$ m.y. of the granodiorite (Bryant and Hedge, 1978, fig. 3) is almost identical with the $1,665 \pm 40$-m.y. Rb-Sr age of the Boulder Creek Granodiorite. A Lu-Hf age of $1,700 \pm 20$ m.y. also is reported (Patchett and others, 1981, table 1). Considering the composition of the rocks and the proximity to the Boulder Creek batholith, most of the rock in the Mount Evans batholith logically can be called Boulder Creek Granodiorite. Coarse-grained granite and quartz monzonite of the same age as, but intrusive into, the granodiorite was referred to as the granite of Rosalie Peak by Bryant and Hedge (1978). This rock bears the same relation to the granodiorite as does the Twin Spruce Quartz Monzonite in the Boulder Creek batholith, and it reasonably might be referred to the Twin Spruce.

GABBRO OF ELKHORN MOUNTAIN

Gabbro in a small batholith on the west side of the Park Range and Sierra Madre at the Wyoming border (fig. 3) was referred to as the gabbro of Elkhorn Mountain by Snyder (1980a). Gabbro in a smaller pluton 10 km (6 mi) to the southeast also was assigned to this unit by Snyder. The gabbro of the batholith is exposed in an area of about 130 km^2 (50 mi^2), and it is present in the subsurface in a somewhat larger area as indicated by boreholes on each side of the Colorado-Wyoming boundary, by inclusions in Tertiary intrusives, and by a pronounced magnetic anomaly. As described by Snyder (1980a), the rock of the batholith is a dark hornblende gabbro that in most places contains clinopyroxene or biotite or both. The gabbro is cut by dikes of basalt and by many dikes and irregular bodies of granitic rock called the quartz monzonite of Seven Lakes by Snyder (1980a). Near its northeast corner, the batholith contains a small pluton of a gabbro that is older than the main body. This pluton, about

4 km^2 (1.5 mi^2) in area, consists of an olivine-two pyroxene-amphibole gabbro. It is cut by a few small dikes of the hornblende gabbro and by numerous dikes and pods of amphibole peridotite.

On the basis of a highly concordant U-Pb determination on zircon, age of the gabbro of Elkhorn Mountain is about 1,755 m.y. (Snyder, 1980a). This makes the gabbro the oldest major unit in the Routt Plutonic Suite. The quartz monzonite of Seven Lakes is dated by the Rb-Sr method as about 1,665 m.y. in age (Snyder, 1980a) and is also a unit of the Routt Suite. In addition to the occurrences within the gabbro, it forms scattered small concordant bodies eastward to North Park (Snyder, 1980a). The basalt dikes that cut the gabbro are reported by Snyder to be both older and younger than the quartz monzonite.

CROSS CREEK QUARTZ MONZONITE

Quartz monzonite and granodiorite resembling those of the Boulder Creek and Rawah batholiths constitute parts of the Gore and northern Sawatch Ranges. These granitic rocks were formally named Cross Creek Granite by Tweto and Lovering (1977) following many years of informal usage (Lovering and Tweto, 1944). The name is here changed to Cross Creek Quartz Monzonite to express more accurately the composition of a principal facies as well as an empirical average of compositions that range from quartz diorite to granite. As interpreted from aeromagnetic data (U.S. Geological Survey, 1968; Tweto and others, 1970, p. 33), the exposures in the Gore and Sawatch Ranges are parts of a larger body here called the Cross Creek batholith. This batholith, about 60 km (40 mi) long, trends northeast subparallel to the Rawah batholith to the northwest and to a major Precambrian tectonic feature, the Homestake shear zone, on the southeast (fig. 3). The Homestake shear zone is a wide belt of shear zones in which movements occurred repeatedly in the Precambrian (Tweto and Sims, 1963; Tweto, 1974). The shear zone marks the southern limit of the Cross Creek batholith. Border facies of the batholith lie within or along the shear zone, indicating that the shear zone predated the Cross Creek Quartz Monzonite.

The Cross Creek Quartz Monzonite is intimately associated with migmatites, shows abundant evidence of reaction with and assimilation of biotite gneiss, and in many places occurs in lenses alternating with lenses or septa of migmatite. Inclusions of migmatite and diorite are abundant near the borders of the batholith and commonly are somewhat granitized. The inclusions and lenses of migmatite display continuity in lithology and orientation through considerable distances in the igneous matrix, which consists of lenses of granodiorite, diorite, and quartz monzonite. The complex relations indicate that both magma and its wall rocks flowed and crumpled as a unit. The present level of exposure of the Cross Creek batholith, particularly in the Gore Range, clearly represents a different and presumably deeper crustal level than do the Boulder Creek and Rawah batholiths.

Where largely free of contaminating migmatitic material, the Cross Creek Quartz Monzonite is gray to pinkish gray, medium to coarse grained, slightly to markedly foliated, and irregularly porphyritic. The porphyritic facies contain prismatic pink perthitic microcline crystals as much as 5 cm (2 in.) long. These crystals locally constitute as much as 50 percent of the rock, and where they are abundant the rock grades to granite. Modal and chemical data may be found in the report by Tweto and Lovering (1977, tables 2, 3). The microcline in the porphyritic facies formed by replacement after plagioclase and biotite locally had been somewhat cataclasized and altered. The crystals occur also in some bordering migmatites, and they thus are porphyroblasts rather than phenocrysts, though various lines of evidence indicate that they were products of the Cross Creek magmatism and not of some younger event (Tweto and Lovering, 1977, p. 10). Dioritic facies of the Cross Creek are mainly biotite tonalites and quartz diorites. These rocks occur both as inclusions in the quartz monzonite and as lenses or streaks seemingly contemporaneous with the quartz monzonite. Hornblende or hornblende-biotite diorites occur in dikes and pluglike bodies younger than the quartz monzonite.

Age of the Cross Creek Quartz Monzonite is about 1,675 m.y. as determined from a six-point Rb-Sr isochron (C. E. Hedge, written commun., 1974; Tweto and Lovering, 1977, p. 14). Thus, within the margin of uncertainty, the Cross Creek is of the same age as the Boulder Creek Granodiorite and the rocks of the Rawah batholith.

DENNY CREEK GRANODIORITE

Gneissic granodiorite in the central Sawatch Range was named Denny Creek Granodiorite Gneiss, and a small body of trondhjemitic rock that intrudes it was named Kroenke Granodiorite by Barker and Brock (1965). Bodies of these two rocks (fig. 3) are part of a large and irregular body of intrusive rocks of the Routt Suite that occupies much of the area from the west edge of the Sawatch Range to the mountain front northeast of Canon City (pl. 1). This large body differs in character from area to area and probably is a cluster of plutons of various sizes. For many years, rocks in most of the large body were erroneously called Pikes Peak Granite (Stark and Barnes, 1935; Stark, 1935; Stark and others, 1949; Lovering and Goddard, 1950;

Dings and Robinson, 1957). The misapplications of the name stemmed from the mistaken assumption that a gneissic granodiorite with which the Pikes Peak is in contact was a part of the batholith rather than a wall rock of the batholith. No valid name to replace the misnamed rocks northeast and west of Canon City and near Eleven Mile Reservoir in southeast South Park has yet been proposed. The granodioritic rocks in the Mosquito Range and southwest South Park are Denny Creek Granodiorite, as are those of the central Sawatch Range. The name Trout Creek Augen Gneiss was applied to the granodiorite in the Mosquito Range by Hutchinson and Hedge (1967) and Hutchinson (1972) but is invalid (Tweto, 1977, table 1). In the interest of consistency, the term "gneiss" is here abandoned as part of the name of the Denny Creek, which is no more foliated than other units of the Routt Plutonic Suite.

As described by Barker and Brock (1965) and Brock and Barker (1972), the Denny Creek Granodiorite is medium to dark gray, coarse grained, well foliated, and contains unevenly distributed augen of pale-pink microcline perthite. Modal and chemical analyses may be found in the report by Barker and Brock (1965, table 1). Where microcline is abundant the rock grades to quartz monzonite, and where sparse the rock grades to biotite quartz diorite. Inclusions of hornblende diorite or of mica schist are locally abundant, and in places the granodiorite is in sheets interlayered in biotite gneiss or migmatite. The Denny Creek is about the same age as the Cross Creek and the Boulder Creek rocks. It is cut by Kroenke Granodiorite, which is about 1,665 m.y. in age (Barker and others, 1974). Augen gneiss of the Denny Creek (the so-called Trout Creek) in the Mosquito Range was dated by C. E. Hedge (Hutchinson and Hedge, 1967, p. 27) as about 1,665 m.y. Thus, the Denny Creek probably is only slightly older than the Kroenke, and its age may be in the range 1,665–1,675 m.y. Quartz monzonite in the southern Front Range, in the eastern part of the large body of which the Denny Creek is an element, has been dated at about 1,745 ± 66 m.y. on the basis of an eight-sample Rb-Sr whole-rock isochron plot (Vera and Van Schmus, 1974).

KROENKE GRANODIORITE

The leucocratic and trondhjemitic Kroenke Granodiorite occupies a pluton about 130 km^2 (50 mi^2) in area and a few smaller outlying bodies in the central Sawatch Range (fig. 3). The pluton is discordant and in most places is bordered by a brecciated zone in the wall rocks (Barker and others, 1974). Barker and Brock (1965) named the Kroenke and described it as quartz diorite, granodiorite, and quartz monzonite. These rocks are white to buff, fine to medium grained, and foliated to massive. Biotite is sparse and occurs as uniformly scattered flakes in the massive varieties. In the foliated varieties it is concentrated in laminae that give the rock a faintly banded appearance. Swarms of amphibolite inclusions are present locally. An early quartz diorite facies of the Kroenke was fractured and then invaded by the predominant trondhjemitic granodiorite facies. Detailed chemical and isotopic data on the Kroenke and discussions of the origin and tectonic significance of trondjhemitic rocks in Colorado and New Mexico are in a report by Barker and others (1976). As noted, the Kroenke is younger than the Denny Creek Granodiorite and is about 1,665 m.y. in age (Barker and others, 1974).

GRANITIC ROCKS IN TAYLOR PARK-GUNNISON REGION

Rocks assigned to the Routt Plutonic Suite occupy several areas in the Taylor Park-Gunnison region, on the west side of the Sawatch Range and southwestward (pl. 1). Some of these areas are single plutons; others consist of multiple intrusive bodies, some of which are questionably classified. Although a few invalid names have been applied in the past, none of the rocks in these areas bear formal names except the Powderhorn Granite, south of Gunnison.

In the discussion that follows, references are made to Rb-Sr age determinations made by Wetherill and Bickford (1965) on several of the granitic rocks in the region. A single isochron of 1,617 ± 35 m.y. applies to all the numerous rocks sampled. Although that result indicates that all the rocks are of about the same age, possibly excepting two that are nearly devoid of rubidium, the age is approximate and probably is low. This is so because many of the analyzed rocks were cataclasized, some were somewhat altered, and they came from many different intrusive bodies and migmatites in an extensive region.

Extensive bodies of granitic rocks in the vicinity of Taylor Park, northeast of Gunnison (pl. 1), consist of several different rock units, some of which present problems in classification that as yet are unresolved. Some of the rock on the east side of Taylor Park is Denny Creek Granodiorite, and some in scattered small bodies on both sides of the park is Kroenke Granodiorite. Porphyritic granodiorite and augen gneiss in a large body bisected by the Taylor River southwest of Taylor Reservoir has characteristics of the Routt Suite, being foliated and somewhat cataclasized and containing clotted biotite, prismatic or ovoid potassium feldspar crystals, and prominent accessory sphene. On the basis of one sample, Wetherill and Bickford (1965, p. 4673) suggested that the rock might be somewhat older than the other rocks represented in their 1,617 ± 35-m.y. isochron. However, on the basis of several analyses, C. E. Hedge and R. E. Zartman found both the Rb-Sr and the U-Pb systems too disturbed to yield meaningful age data (C. E. Hedge, written commun., 1976). The rocks are classified as a unit of the Routt Plutonic Suite because of their character and habit, but further investigation of their metamorphic and deformational histories is needed. Some of them, notably a dark granodiorite augen gneiss south of Taylor Park, might be related to the protolith of the gneiss complex. A facies of the assemblage on the west, near Almont, was classified as biotite tonalite by Urbani and Blackburn (1974).

As in the Taylor Park area, a body of granitic rocks in the Monarch Pass–Pitkin area consists of several different rock units, some of which are uncertainly classified. Three units mapped by Dings and Robinson (1957) as gneissic granite, "Pikes Peak Granite," and hornblende diorite have the characteristics of the 1,700-m.y. intrusive rocks and are assigned with confidence to the Routt Plutonic Suite. The gneissic granite has a cataclastic foliation and possibly is a deformed facies of the biotite granite that was erroneously assigned to the Pikes Peak. As seen in most places, this latter granite is foliated, coarse grained, and porphyritic. It resembles the Denny Creek Granodiorite but has a larger content of prismatic potassium feldspar crystals, which makes the rock a granite. In some places the rock is only vaguely foliated or massive (Dings and Robinson, 1957, p. 7). The hornblende diorite occurs with the gneissic granite but lacks the cataclastic foliation, indicating that the diorite is younger than the granite. A moderately large body of rock called Silver Plume(?) Granite by Dings and Robinson (1957, pl. 1) is difficult to classify and may consist of two or more intrusive units. Some of the rock is slightly gneissic, has biotite in clots, contains dioritic schlieren, and where porphyritic, contains prismatic potassium feldspar crystals. This rock is tentatively classed as a unit of the Routt Plutonic Suite though younger than the gneissic granite and the coarse biotite granite of the same area. Closely intertongued with the rock just described is a pink to light-gray texturally heterogeneous granite that contains the tabular potassium feldspar crystals characteristic of many of the granites in the Berthoud Plutonic Suite. Rb-Sr data from a sample of one of the two units, described as medium-grained granodiorite, is included in the 1,617-m.y. isochron plot of Wetherill and Bickford (1965, fig. 2, no. CO–55), but the rock is so poor in rubidium that the data are not meaningful to the isochron. The entire granitic body in the Monarch Pass–Pitkin area is classified as a unit of the Routt Plutonic Suite in plate 1, though it probably contains some rock of the Berthoud Plutonic Suite.

Coarse-grained porphyritic pink granite in a partly discordant pluton northwest of Doyleville was described by Staatz and Trites (1955) and was referred to as the granite of Woods Gulch by Olson (1976b). Porphyritic quartz monzonite and tonalite nearby probably are related to the granite. Three samples of the granite are represented in the 1,617-m.y. whole-rock Rb-Sr isochron plot of Wetherill and Bickford (1965), who reported cataclastic textures in all three samples. The granite is one of the country rocks in the Quartz Creek pegmatite district. Early radiometric dating led to a conclusion that the pegmatites and the granite were both about 1,325 m.y. in age (Aldrich and others, 1956), but Wetherill and Bickford (1965, p. 4676) later established that the pegmatites are significantly younger than the granite, and that the isotopic systems in the pegmatites were disturbed in a subsequent metamorphic event, creating apparent mineral ages of about 1,325 m.y.

Granitic rocks in several small plutons southwest and southeast of Gunnison are assigned to the Routt Plutonic Suite. Rocks in a pluton a short distance southwest of Gunnison, mainly in T. 49 N., R. 1 W., were called granite of South Beaver Creek by Hedlund and Olson (1974) and Hedlund (1974). The pluton consists principally of granite and quartz monzonite but contains granodiorite and quartz diorite. Inclusions of migmatite and hornblende schist are abundant. The pluton is concordant and has a marked circular structure in its exposed part. The various compositional facies and the inclusions, many of which are elongate septa, are arranged concentrically about a body of dioritic rocks at the center of the pluton. A sample of the granite is represented in the plot of the 1,617-m.y. Rb-Sr isochron of Wetherill and Bickford (1965, fig. 2, no. CO 73), and more recently the rock was dated by U-Pb measurements on zircon as $1,720 \pm 5$ m.y. (Bickford and others, 1982). The granite has been quarried as building stone and was used to construct the Colorado State Capitol.

A pluton southeast of the one just discussed, mainly in T. 48 N., R. 1 E., consists of a trondhjemite called "quartz diorite to quartz monzonite of Gold Basin" by Hedlund and Olson (1974), Olson (1976a), and Olson and Steven (1976). As described by those authors, the rock contains about 65 percent of sodic oligoclase and only 2–5 percent of potassium feldspar. The rock is light pinkish gray and medium grained. Contacts of the pluton parallel the foliation in bordering gneisses in most places, but they are sharply discordant locally. Though the rock has not been dated, the composition indicates affiliation with the family of old trondhjemitic rocks that includes the approximately 1,700-m.y. Kroenke and Pitts Meadow Granodiorites and the slightly older Twilight Gneiss (Barker, Arth, and others, 1976).

Immediately southeast of the trondhjemite pluton, across an east-trending Precambrian fault zone, is a pluton of granitic rocks called the quartz monzonite of Cochetopa Creek by Olson and Steven (1976) and briefly alluded to as Cochetopa granite, an informal unit, by Hutchinson (1981). The granite or quartz monzonite is pink, medium to coarse grained, and weakly to moderately gneissic. Sphene is a conspicuous accessory mineral. As mapped in part by Hutchinson (1981), the pluton contains numerous inclusions and long septa of gneiss oriented parallel to the foliation in bordering gneisses and to the contacts of the generally concordant pluton. The rock of the pluton has a pronounced cataclastic foliation on the north side of a Precambrian shear zone of east-northeast trend that cuts across the southern part of the pluton. Rock in a belt on the south side of the shear zone was mapped as gray granodiorite by Olson and Steven (1976) and as metadiorite by Hutchinson (1981). One sample from the pluton, described as granite, is represented in the 1,617-m.y. Rb-Sr isochron plot of Wetherill and Bickford (1965).

POWDERHORN GRANITE AND RELATED ROCKS

Granitic rocks in several grossly concordant bodies near the village of Powderhorn, southwest of Gunnison, were originally designated units of the Powderhorn Granite Group by Hunter (1925). With recognition that some of the rocks in Hunter's Powderhorn Group were metavolcanic rather than intrusive, the Powderhorn later was reduced to formational rank. As recognized in mapping by Olson and Hedlund (1973), Olson (1974), and Hedlund and Olson (1975) (fig. 3), the Powderhorn Granite is a pink to gray, medium- to coarse-grained, gneissic porphyritic biotite granite. Near contacts with the younger Precambrian or Cambrian alkalic stock of Iron Hill, the granite is fenitized (altered to a syenitic composition). An elongate concordant arcuate body of granite northwest of Powderhorn was included in the Powderhorn Granite by Hunter (1925), but was distinguished separately as the granite of Tolvar Peak by Olson and Hedlund (1973) and Hedlund and Olson (1975). This granite differs substantially from the Powderhorn, being trondhjemitic and containing scattered euhedral quartz phenocrysts. It also contains abundant inclusions of hornblende schist of the Dubois Greenstone wall rocks. Analyses of the granite of Tolvar Peak and the Powderhorn Granite are in a report by Hedlund and Olson (1981, table 2). The granite of Tolvar Peak is dated as $1,754 \pm 5$ m.y. in age by the U-Pb method on zircon (Bickford and others, 1982). Although this figure indicates that the granite is older than most of the Routt Suite, the inclusions establish that the granite is younger than, and not an element of, the gneiss complex.

A body of metagabbro and hornblende quartz diorite exposed along Cebolla Creek about 10 km (6 mi) south of Powderhorn probably is closely related in age to the Powderhorn Granite. As described by Hunter (1925, p. 56–61), the metagabbro is dark green, medium grained, slightly porphyritic, and massive to slightly gneissic. It is bordered by belts of gneissic hornblende quartz diorite. The diorite intrudes the gabbro, but the two rocks are judged to be comagmatic and to be about the age of the Powderhorn Granite. Hunter (1925, p. 41) noted that although age relations between the gabbro and the granite were not well defined, the gabbro probably was the younger. Olson (1974), however, mapped dikes of Powderhorn cutting the gabbro. His map also suggests that the quartz diorite cuts the Powderhorn. Thus, the sequence from oldest to youngest seems to be (1) gabbro, (2) Powderhorn Granite, and (3) quartz diorite. None of these rocks has been dated, but their character and habit classify them as units of the Routt Plutonic Suite.

PITTS MEADOW GRANODIORITE

Trondhjemitic granodiorite and quartz diorite exposed near the northwest end of the Black Canyon of the Gunnison, southeast of Delta (pl. 1, fig. 3), were first described by Hansen (1968) and were named by him Pitts Meadow Granodiorite. Rocks of the unit are gneissic and contain many mafic inclusions, most of which show evidence of reaction and partial assimilation by the trondhjemitic magma. Foliation is parallel to contacts and to foliation in bordering gneisses. A predominating light-colored facies of the granodiorite contains 45–55 percent oligoclase, 25–35 percent quartz, 2–18 percent microcline, 5–15 percent biotite, and 1–5 percent hornblende. A subordinate dark facies contains about 15 percent each of biotite and hornblende. Several chemical analyses of the rocks are available (Barker, Arth, and others, 1976, tables 2, 3). Extent of the Pitts Meadow beyond the canyon is unknown. Granitic rock in several boreholes near the canyon (pl. 1) probably is trondhjemite of the Pitts Meadow, but at some unknown location farther west this gives way to normal granodiorite or quartz monzonite.

Age of the Pitts Meadow was reported by Hansen and Peterman (1968) to be $1,695 \pm 190$ m.y. as determined from a seven-sample Rb-Sr isochron plot. The large calculated uncertainty reflects low Rb/Sr ratios. Thus, the age is similar to that of the trondhjemitic Kroenke Granodiorite (Barker and others, 1974) and to that of other units of the Routt Plutonic Suite.

TENMILE AND BAKERS BRIDGE GRANITES

Granitic rocks of the Routt Plutonic Suite are sparse in the Precambrian terrane of the Needle Mountains, where they are represented by the Tenmile Granite and the Bakers Bridge Granite, each mainly in a single stock. The Tenmile Granite is in a stock bisected by the Animas River about 12 km (8 mi) south of Silverton (pl. 1) and in small bodies farther north that were called Whitehead Granite by Cross and Howe (Cross, Howe, and others, 1905, p. 8), a term that was later abandoned (Barker, 1969a, p. 22). The Tenmile was first described and named by Cross and Howe (Cross, Howe, and others, 1905, p. 7) and was described further by Barker (1969a). The principal facies of the granite is pink to light gray, medium to coarse grained, faintly to moderately foliated, and locally contains abundant amphibolite xenoliths. A subordinate dark facies intrudes the lighter facies. At the borders of the stock, thin sills of the granite alternate with septa of gneiss of the enclosing Irving Formation. The Tenmile is closely dated at about 1,695±20 m.y. by the U-Pb method on zircon (Silver and Barker, 1968), and about 1,690±55 m.y. by the Rb-Sr method (Bickford and others, 1969).

The Bakers Bridge Granite is in a stock that is partially exposed in the valley of the Animas River about 22 km (14 mi) north of Durango (pl. 1). The granite was named by Barker (1969a, p. 23), who described it as pale red, coarse grained, massive, and homogeneous. It consists largely of microcline perthite and quartz, which are accompanied by small amounts of plagioclase and biotite. Swarms of amphibolite inclusions are present locally in the granite. Chemical analyses of both the Bakers Bridge and the Tenmile Granites are provided by Barker (1969a, table 2). Silver and Barker (1968) reported a U-Pb age of zircon from the Baker Bridge of 1,695±20 m.y., the same age as they reported for the Tenmile. A Rb-Sr age of 1,658±78 m.y. is also reported (Bickford and others, 1969).

GRANODIORITE OF THE GRAND JUNCTION AREA

On the basis of limited exposures at the Utah border and in Unaweep Canyon on the Uncompahgre Plateau, and occurrence in a few boreholes, a batholithic unit of the Routt Plutonic Suite is inferred to occupy a large area in central western Colorado near Grand Junction (pl. 1). Bodies of coarse-grained gneissic granodiorite along the Utah border (Case, 1966) are interpreted to be fingers of this batholith. The granodiorite has a Rb-Sr age of about 1,635±40 m.y. as determined from an isochron plot for four samples (Hedge and others, 1968). In Unaweep Canyon (T. 14 S., R. 100 W. to T. 15 S., R. 103 W., pl. 1), dark-gray coarse-grained gneissic granodiorite intertongues complexly with other varieties of granitic rocks and with felsic gneisses. Precambrian rocks in the canyon are inadequately mapped, and distribution of the granodiorite shown in plate 1 is generalized. Samples of the granodiorite and felsic gneiss evidently were included by Mose and Bickford (1969) in an isochron plot presumed to represent the Vernal Mesa Granite (of the Berthoud Plutonic Suite). The samples of these older rocks doubtless account for a spuriously old age reported by those authors for the Vernal Mesa.

Small exposures of Precambrian rocks in canyons in the Colorado National Monument west of Grand Junction are shown as biotite gneiss in plate 1 because of their predominantly migmatitic character, but they consist in part of rather dark granodiorite and quartz diorite. These exposures are interpreted to be in the border zone of a rib of gneisses within the irregular granodiorite batholith. In the subsurface farther north, a sample of gneissic granodiorite (Edwards, 1966, p. 225) in a borehole in sec. 8, T. 8 S., R. 102 W. was dated as about 1,740±90 m.y. by the Rb-Sr method (Muehlberger and others, 1966, p. 5415).

PLUTONS BENEATH EASTERN PLAINS

Boreholes establish the presence of several bodies of granitic rocks of the Routt Plutonic Suite in subsurface eastern Colorado (pl. 1). Little information is available to define the extent or shape of the bodies, except that the areas proved by boreholes coincide in most places with magnetic highs. A similar correlation was noted by Lidiak (1972, p. 27) in Nebraska. Therefore, magnetic data (Zietz and Kirby, 1972) were used to outline the granitic bodies as shown in plate 1, but at best the outlines can be no more than approximate. The Precambrian rocks of many of the boreholes were described petrographically by Edwards (1966), and they were classified by me from the descriptions.

Rock interpreted as gneissic quartz monzonite or granodiorite was encountered in six boreholes in an area of at least 1,900 km^2 (720 mi^2) in the vicinity of Sterling, near the northeast corner of the State. The area corresponds rather closely with a northeast-trending magnetic high of as much as 500 gammas. The rocks are medium to coarse grained, moderately foliated, and locally porphyritic. Rb-Sr mineral ages of 1,555 m.y. on feldspar and 1,287 m.y. on biotite reported from two of the boreholes (Goldich, Lidiak, and others, 1966) probably reflect modifications by post-crystallization events, including weathering. A single borehole immediately to the southwest of the Routt body (pl. 1) bottomed in granite that clearly is of the Berthoud Plutonic Suite. Geochemical disturbance associated with this granite may be responsible for the low mineral ages. Another body of the Routt Suite lies to the east in the Holyoke area, where it is inferred to occupy about 3,000 km^2 (1,150 mi^2) in Colorado and to extend eastward a short distance into Nebraska. An anomalously low Rb-Sr whole-rock age of about 1,400 m.y. (Goldich, Lidiak, and others, 1966) from one borehole probably reflects weathering, as the borehole penetrated only 2.4 m (8 ft) into the basement beneath Cambrian sandstone.

In central eastern Colorado, mainly in Kit Carson County, a body of granitic rocks of the Routt Suite occupies a subsurface area of at least 2,800 km^2 (1,080 mi^2). This body corresponds with a north-trending magnetic high that has a relief of as much as 500 gammas. Smaller bodies are present in southeastern Colorado in the vicinity of Las Animas and of Fowler (pl. 1). Coarse-grained and porphyritic gneissic quartz monzonite from a borehole in a small body of the Routt Suite between Pueblo and Florence yielded an anomalously low muscovite age of about 1,340 m.y. by the Rb-Sr method (Muehlberger and others, 1966). The rock contains as much as 10 percent of magnetite (Edwards, 1966, p. 318) and thus is itself anomalous. A small subsurface body of gneissic quartz monzonite along the New Mexico border east of long 104° W. is surrounded by granite of the Berthoud Suite (pl. 1), but is assigned to the Routt Suite because of its gneissic character and its association with apparent stringers of schist and gneiss in boreholes on the New Mexico side of the boundary (Foster and Stipp, 1961; Darton, 1928, p. 308).

BERTHOUD PLUTONIC SUITE

In a second period of major igneous intrusion, about 1,400 m.y. ago, many large bodies of granitic rocks invaded the already deformed and metamorphosed Precambrian terrane of Colorado. This magmatism was part of a broad belt of anorogenic igneous intrusion that extended southwestward across the south-central part of the continent from Labrador to California (Silver, Bickford, and others, 1977; Emslie, 1978). In Colorado, most of the intrusions consisted of granite or quartz monzonite, but a few small ones were of syenite or mafic rocks. The name Berthoud Plutonic Suite is here proposed for this family of related intrusive rocks emplaced about 1,400 m.y. ago in Colorado. The suite is a lithodemic unit. It takes its name from Berthoud Pass, which is located within the Silver Plume batholith in the central Front Range (fig. 3). This batholith serves as the type area for the Berthoud Suite, but because no single pluton contains examples of all the rocks in the suite, many of the rock bodies discussed in following pages must be considered reference localities.

The various rock units that constitute the Berthoud Suite have in common a generally discordant habit as well as similarities in character and age. Although they were emplaced long after the major deformation recorded by two periods of early folding and the intrusions of the Routt Plutonic Suite, they were accompanied by deformation in some areas. This deformation was principally by fracture, and it led to development or reactivation of major shear zones and faults, mainly of northeast and north-northwest trends (Tweto and Sims, 1963). In and near shear zones in the Front Range (Moench and others, 1962), incompetent rocks such as biotite gneiss were deformed by a third generation of folds (Gable and Sims, 1970, p. 11; Taylor, 1975a). Time relations between the intrusive rocks and the shear zones or third-generation folds are varied because both intrusion and deformation were spread through an appreciable time interval. Widespread retrograde metamorphism and disturbance of isotopic systems accompanied the intrusion and deformation. As a result, rocks older than the Berthoud Suite commonly yield K-Ar ages nearly identical with those of the Berthoud Suite.

The Berthoud Plutonic Suite consists of the formally named rocks of several intrusive bodies and the unnamed, or informally named, rocks of numerous other bodies. Currently recognized named lithodemes contained in the suite are the Silver Plume Granite (or Quartz Monzonite) of the central and northern Front Range, the Sherman Granite of the northern Front Range, the Cripple Creek Granite (or Quartz Monzonite) of the southern Front Range, the San Isabel Granite of the Wet Mountains, the St. Kevin Granite of the northern Sawatch Range, the Vernal Mesa and Curecanti Quartz Monzonites of central western Colorado, and the Eolus Granite, Trimble Granite, and Electra Lake Gabbro of the Needle Mountains.

SILVER PLUME GRANITE

The Silver Plume Granite (or Quartz Monzonite) as presently recognized constitutes three batholiths and numerous smaller plutons in the central and northern Front Range. The granite was named by Ball (1906) for the town of Silver Plume near the east edge of a batholith of modest size that has come to be known as the Silver Plume batholith (fig. 3). A larger batholith about 24 km (15 mi) to the northeast was named the Longs Peak-St. Vrain batholith (fig. 3) and its rocks correlated with the Silver Plume by Boos and Boos (1934). A minor marginal facies in this batholith earlier had been designated the Mount Olympus Granite by Fuller [later, M. F. Boos] (1924), but was classed as Silver Plume Granite by Lovering and Goddard (1950, pl. 1) and Peterman and others (1968). All these authors as well as Abbott (1976) also assigned to the Silver Plume the rocks of the Log Cabin batholith (fig. 3), a pluton in the northern Front Range about 20 km (12 mi) north of the Longs Peak-St. Vrain batholith that was named and briefly described by Boos and Boos (1933). Granitic rocks in several irregular plutons south and southeast of the Silver Plume batholith have also been assigned to the Silver Plume Granite. Quartz monzonite in a small pluton southwest of Evergreen was assigned to the Silver Plume by Bryant and Hedge (1978, p. 449) and by Boos (1954, p. 119), who thus effectively abandoned an earlier name, Indian Creek Granite (Boos and Aberdeen, 1940). "Indian Creek" was nonetheless used again by Puffer (1972). Boos (1954, p. 120) also applied the outmoded name Mount Olympus to a dike facies in the area. Another pluton of muscovite-biotite quartz monzonite referred to the Silver Plume lies along the east side of northern South Park. This pluton was called the Kenosha batholith (fig. 3) by Boos and Aberdeen (1940, fig. 1) and has been discussed briefly by Hutchinson (1960) and Hutchinson and Hedge (1967).

The Silver Plume and the Longs Peak-St. Vrain batholiths are irregular in outline and have many ragged apophyses extending into the bordering gneisses. They contain or surround island-like bodies of gneisses, and are accompanied in surrounding areas by numerous small satellitic plutons. Some parts of the batholiths contain abundant inclusions, and in parts of the Silver Plume batholith many of these are of granodiorite of the Routt Suite (Theobald, 1965). The contacts of the batholiths are varied in character and predominantly discordant. Some segments of the contacts follow older Precambrian faults or shear zones. Other segments are serrated by protruding discordant flat sheets or dikes. Still other segments are gradational through zones of migmatite and abundant pegmatite. In inconspicuous shear zones within complexly intruded migmatitic areas west of Berthoud Pass (Theobald, 1965), an older facies of the granite has cataclastic foliation, but a younger facies is massive and crosscuts the foliation, indicating that movements occurred along the shear zones during the period of Silver Plume emplacement.

Most Silver Plume Granite is light colored, medium grained to seriate porphyritic, and generally structureless except as the alinement of potassium feldspar phenocrysts may define a primary flow structure. Somewhat darker and finer grained varieties, including the rock that was called Mount Olympus by Fuller (1924), occur locally, principally in border areas. The Silver Plume characteristically contains a few percent of primary muscovite, which has led to its identification only as "biotite-muscovite granite" in some satellitic bodies (Harrison and Wells, 1956, p. 33; Moench, 1964, p. 37; Sims and Gable, 1967, p. 39). Potassium feldspar (perthitic microcline) crystals in the porphyritic varieties are tabular, 5–15 mm (0.2–0.6 in.) long, and commonly display Carlsbad twinning. In places, the crystals have a preferred orientation that outlines primary flow structure. Near contacts, this structure parallels the contact, but it may be highly discordant with the structure in bordering rocks. Except in locally gneissic facies along contacts, fabric of the granite is hypautomorphic granular.

Numerous modes of the Silver Plume Granite in localities near the type area are reported by Braddock (1969, table 9), Sims and Gable (1967, table 23), Moench (1964, table 8), and Harrison and Wells (1956, table 7). Rock described as granodiorite, but actually quartz monzonite, in a small quarry just west of Silver Plume has been designated international geochemical reference rock GSP–1. Numerous analyses consequently are available (Flanagan, 1973, table 1; 1976, table 103), and additional chemical and isotopic data for Rb, Sr, Pb, Th, and U also are available (Peterman and others, 1967). Chemical and petrographic analyses of rocks in the Longs Peak-St. Vrain batholith indicate granite and quartz monzonite compositions slightly more siliceous than in the Silver Plume batholith (Boos and Boos, 1934, table 1, fig. 5). A chemical analysis and brief description of quartz monzonite in the Log Cabin batholith were provided by Eggler (1968, table 1, p. 1549). In mapping the southeastern part of the Log Cabin batholith, Abbott (1976) classed the rock as quartz monzonite and divided it into massive, porphyritic, and dark seriate porphyritic facies.

The Silver Plume Granite is comparatively rich in thorium and the rare earths of the cerium group (Goddard and Glass, 1940; Phair and Gottfried, 1955; Peterman and others, 1967), reflecting the widespread though uneven occurrence in it of accessory monazite, allanite, and compositionally related minerals. Smoky quartz in some parts of the granite probably reflects radioactivity originating in these minerals.

Radiometric ages have been determined for several different bodies and facies of the Silver Plume Granite. The granite in the type area near Silver Plume originally was dated by Rb-Sr as about 1,325 m.y. (Aldrich and others, 1958), but later was determined in two Rb-Sr investigations to be about $1{,}410 \pm 30$ m.y. (Peterman and others, 1967; C. E. Hedge, written commun., 1982). A Lu-Hf age of about 1,420 m.y. also is reported (Patchett and others, 1981). A U-Pb age of about 1,395 m.y. was obtained from zircon and uraninite in outlying bodies (Stern and others, 1971, p. 1625). A Rb-Sr isochron age of about $1{,}420 \pm 30$ m.y. for rocks of the Longs Peak-St. Vrain batholith was reported by Peterman and others (1968, fig. 8). These authors also reported a Rb-Sr age of about $1{,}392 \pm 30$ m.y. for Silver Plume Granite in the Log Cabin batholith. In the 1930's, a primitive attempt at U-Pb dating of cerite in pegmatitic Silver Plume yielded an apparent age of 940 m.y., which was less than the 1,000 m.y. that had been obtained from samarskite in the Pikes Peak Granite (Goddard and Glass, 1940, p. 404). This result contributed to a consensus that the Silver Plume was younger than the Pikes Peak Granite and was the youngest granite in the Front Range. That notion held sway until the late 1950's, when more detailed geologic studies and more advanced methods of radiometric dating (Aldrich and others, 1958) proved otherwise.

CRIPPLE CREEK GRANITE

The Cripple Creek Granite constitutes a batholith about 650 km^2 (250 mi^2) in area in the southern Front Range on the west side of the younger and much larger Pikes Peak batholith (fig. 3). The Cripple Creek was defined by Mathews (1900), who judged it to be younger than the Pikes Peak because of being finer grained, an interpretation that later was proved erroneous. The Cripple Creek closely resembles the Silver Plume Granite except that it is more consistently pink or red. Like the Silver Plume, it contains primary muscovite and has both equigranular and porphyritic facies and a hypautomorphic granular fabric. Phenocrysts are perthitic microcline, locally alined in flow structure. Fluorite is a characteristic accessory mineral. At its north end, the Cripple Creek batholith as mapped in plate 1 consists of pink porphyritic rock that was called the quartz monzonite of Elevenmile Canyon by Hawley and Wobus (1977). This quartz monzonite was included in the Cripple Creek Granite in the original description by Mathews (1900), and it is here classed as a facies of the Cripple Creek. The quartz monzonite is less potassic than most of the Cripple Creek and contains only small amounts of muscovite (Hawley and Wobus, 1977, table 7, fig. 15). Inclusions of gneissic granodiorite of the Routt Suite are common in it, and it is cut by dikes and pods of leucogranite. Stocks of quartz monzonite near this body were called Silver Plume(?) by Hawley and Wobus (1977, fig. 2), but the name Cripple Creek would be more appropriate for reasons both of geography and priority.

The Cripple Creek and Silver Plume Granites have long been regarded as equivalents (Ball, 1906; Boos and Boos, 1934), though they might be of slightly different ages. A Rb-Sr whole-rock age of about 1,400 m.y. reported for the Cripple Creek (Hutchinson and Hedge, 1967, fig. 8) is essentially the same as the age reported for the type Silver Plume. The quartz monzonite facies near Elevenmile Canyon may be a little older than the main part of the Cripple Creek batholith as suggested by a Rb-Sr age of about 1,430 m.y. (Hawley and Wobus, 1977, p. 23).

SAN ISABEL GRANITE

The San Isabel Granite occupies a pluton about 195 km^2 (75 mi^2) in area and a few smaller plutons nearby in the Wet Mountains of south-central Colorado (fig. 3). The name was proposed by Boyer (1962, p. 1056) in a study that included only a part of the main pluton. On the basis of later mapping (Taylor, 1974; Scott and others, 1978) that established the extent of the granite, the San Isabel is here adopted as a formal stratigraphic unit and the pluton is recognized as the type area. The San Isabel is a coarse-grained porphyritic biotite granite or quartz monzonite. Unlike the Silver Plume Granite, it is devoid of primary muscovite, is moderately foliated in many places, and its microcline phenocrysts tend more to prismatic or ovoid than to tabular forms. Although these features are characteristic of the Routt Suite, the San Isabel has the habit and age of the Berthoud Suite. The plutons are generally discordant, and the contacts with bordering gneisses are sharp. Migmatites among those gneisses were thought by Boyer (1962) to be genetically related to the San Isabel Granite, but they preexisted the granite and date back to the time of the Routt Suite or of regional metamorphism. Similarly, inclusions or pendants of highly metamorphosed and mineralized rocks within the pluton (see section on felsic and hornblendic gneisses) were simply enveloped by the granite, not modified by it. Zircon from the granite yielded a U-Pb age of about 1,380 m.y. (T. W. Stern, oral commun., 1976). The San Isabel is abnormally rich in magnetic minerals (Boyer, 1962, table 1), and it coincides with a positive magnetic anomaly of several hundred gammas (Zietz and Kirby, 1972).

SHERMAN GRANITE

The Sherman Granite is in a large batholith and related plutons that lie principally in Wyoming but extend a short distance into Colorado in the Front Range and Medicine Bow Mountains (fig. 3). The granite was described and named for a locality in Wyoming by Blackwelder (1908). In the southern part of the batholith, mainly in Colorado, is an almost circular ring-dike structure about 15 km (9 mi) in diameter described by Eggler (1968). Sherman Granite outside this structure is red, coarse grained, equigranular, and hypidiomorphic-granular in fabric (Eggler, 1968, p. 1548, 1552). The potassium feldspar is perthitic microcline. The plagioclase crystals are large, embayed, and zoned. Quartz is in multigranular pods and as blebs in feldspars. Dark-green hornblende, the principal mafic mineral, is mainly in clots, accompanied by minor biotite. Chemical and petrographic analyses (Eggler, 1968, tables 1, 2, fig. 1; Zielinski and others, 1981, table 6) indicate compositions near the granite-quartz monzonite boundary. The Sherman within the ring structure differs somewhat, being a pinkish-gray porphyritic biotite quartz monzonite that commonly shows flow layering. The quartz monzonite is perforated by a small stock of the Silver Plume Granite of the Log Cabin batholith.

The ring-dike zone, which dips outward, consists of dikes and pods of granite, diorite, andesite, and dacite together with septae of metamorphic rocks. Intrusion in the ring system began with diorite that probably is a hybridized gabbro. Eggler (1968) concluded that the diorite was emplaced in a ring-fault system that probably bordered a caldera at the surface, and that the present core of the ring structure was formed by later upward movement. This would be a deep-level expression of the resurgent doming observed in many calderas.

Fresh Sherman Granite in cores from two boreholes in northernmost Colorado and southern Wyoming yielded a Rb-Sr isochron age of $1,430 \pm 20$ m.y. (Zielinski and others, 1981, fig. 3). An almost identical U-Pb age of $1,422 \pm 15$ m.y. also has been reported (Subbarayudu and others, 1975; Hills and Houston, 1979). Because of the high quality of the samples, the Rb-Sr results supercede previously reported ages of $1,382 \pm 20$ m.y. (Peterman and others, 1968) and $1,375 \pm 30$ m.y. (Hills and others, 1968, table 2). The higher values are consistent with the fact that the Sherman is cut by the Silver Plume Granite of the Log Cabin batholith, for which an age of $1,392 \pm 30$ m.y. is reported (Peterman and others, 1968).

ROCKS OF THE MOUNT ETHEL PLUTON

A long slender body of granitic rocks trending northeast across the Park Range in northern Colorado was named the Mount Ethel pluton by Snyder (1978). The pluton, about 11 km (7 mi) wide, is exposed through a distance of about 27 km (17 mi) in the Park Range and is inferred to extend through at least an equal distance in North Park (pl. 1, fig. 3). In the Park Range, the pluton parallels a major Precambrian fault zone, the Soda Creek-Fish Creek mylonite zone, which lies along its southeast side (Snyder, 1980b). This zone is an ancient, recurrently reactivated zone of movement that was in existence when the pluton was emplaced. The fault zone clearly controlled the trend of the bordering pluton in the Park Range, and the pluton is in effect an extra-wide dike along the fault zone. The fault zone has not been recognized beyond the Park Range but is inferred to continue both to the northeast and the southwest of the exposures in the range (pl. 1). Granite of Mount Ethel pluton is exposed in thrust plates in Delaney Butte and Sheep Mountain in North Park (Tps. 8 and 9 N., R. 91 W., pl. 1), but the thrust sheets do not extend far enough south to reach the mylonite zone. Beyond these exposures, the pluton projects northeastward toward a stock of quartz monzonite of the Berthoud type north of Cowdrey, at the north end of North Park, and beyond that, to a body of Sherman Granite at the Wyoming border (pl. 1). The Mount Ethel pluton and this line of projection are marked by magnetic lows that suggest a belt of granitic intrusion extending to the Wyoming border in the Medicine Bow Mountains (Behrendt and others, 1969, p. 1529, pl. 1). This belt may depart from the Soda Creek-Fish Creek fault zone somewhere beneath North Park.

On the southwest, the Mount Ethel pluton as mapped at the surface terminates against gneisses just short of the flank fault of the Park Range (Snyder, 1980b). However, granite encountered in two boreholes about 8 km (5 mi) to the southwest suggests that the pluton plunges downward from its apparent terminus at the surface and then rises again farther to the southwest. The Soda Creek-Fish Creek fault zone may continue much farther southwest, as indicated in plate 1 and noted in the section on the Uinta Mountain Group.

Rocks of the Mount Ethel pluton have been described in detail by Snyder (1978; 1980b), who recognized several different intrusive units within the pluton, each with sharp contacts. The oldest unit consists of dark-gray medium-grained hornblende-biotite granodiorite and diorite, which occur only as inclusions in other units. Next younger is a red to reddish-gray coarsely porphyritic quartz monzonite that contains tabular euhedral microcline crystals as much as 10 cm (4 in.) long. This rock, characterized by nearly vertical flow structures, forms a semicircular sheath about 3 km (2 mi) wide around the southwest end of the pluton. The next younger unit, called the quartz monzonite of Roxy Ann Lake, constitutes most of the pluton. Rocks of this unit are gray to reddish-gray, fine- to coarse-grained biotite quartz monzonite and biotite granite that locally contain widely scattered microcline phenocrysts. They also contain small amounts of muscovite. In the southwest-central part of the pluton, the rocks of Roxy Ann Lake contain abundant inclusions of the older red porphyritic quartz monzonite. A fine-grained pink muscovite-biotite granite occurs as dikes, pods, and irregular bodies cutting the quartz monzonite of Roxy Ann Lake. Long north-trending dikes of pink aplite cut the southwestern part of the pluton. Modal, chemical, and spectrographic analyses of all these rocks were presented by Snyder (1978, table 5).

The stock north of Cowdrey probably is an outlying extension of the Mount Ethel pluton. The stock is in the upper plate of the Independence Mountain thrust fault (Steven, 1960, pl. 12), and, therefore, the location of its root is uncertain. Steven (1957, p. 365) noted that rocks of the stock closely resemble those of the Mount Ethel pluton (quartz monzonite of Roxy Ann Lake), but he also noted a resemblance to the intrusive body at the Wyoming border to the northeast, which has been assigned to the Sherman Granite (Hills and others, 1968, pl. 1). This would imply that the Mount Ethel pluton is Sherman Granite, but except for the red porphyritic quartz monzonite, rocks of the pluton lack many of the distinctive lithic features of the Sherman. However, as noted in the discussion of the Sherman, and also by Snyder (1978, p. 8), granites of different characters evidently were intruded into the same area at almost the same time.

A Rb-Sr isochron plot by C. E. Hedge indicates an age of about $1{,}440 \pm 50$ m.y. for the Mount Ethel pluton, and also that the red porphyritic quartz monzonite, the quartz monzonite of Roxy Ann Lake, and the fine-grained granite were all emplaced in a short period of time (Snyder, 1978, p. 17). The body of Sherman Granite at the Wyoming border in the Medicine Bow Mountains is represented in an isochron plot that indicates an age of about $1{,}375 \pm 30$ m.y. (Hills and others, 1968, p. 1770), but this figure may be too low. (See discussion of the Sherman Granite.)

ST. KEVIN GRANITE

Granitic rocks in a batholith originally described as being 25 mi long and 12 mi wide in the northern Sawatch Range were named the St. Kevin Granite by Tweto and Pearson (1964). Subsequent mapping proved the batholith to be much more extensive (Tweto and others, 1978; Bryant, 1979, 1971). As now known, the batholith extends across the full width of the Sawatch Range and deep into the Mosquito Range on the east and the Elk Mountains on the west (pl. 1). On the west flank of the Sawatch Range, the batholith trends northeast, parallel to the trend of the Precambrian Homestake shear zone (Tweto and Sims, 1963, p. 1002). On the east flank, the batholith bends eastward, departing from the Homestake zone (fig. 3). The bending is accentuated by faults that displace the batholith contacts southward on their east sides. Major strands of the Homestake shear zone in biotite gneisses on the eastern slope of the range project southwestward into the batholith, where they are apparently absent (pl. 1), suggesting that the batholith post-dates much of the shearing.

Rocks in the part of the batholith exposed along the Mosquito fault near Climax and east of Leadville have been called Silver Plume Granite in the past (Butler and Vanderwilt, 1933; Wallace and others, 1968; Behre, 1953) or, more recently, were designated simply as "of Silver Plume age" (Kuntz and Brock, 1977). In the Sawatch Range, local areas within the batholith were called Silver Plume Granite by Stark and Barnes (1935, pl. 1), and the rocks in a larger area were called Hell Gate Porphyry. Tweto and Pearson (1964) found the Hell Gate to be one of several textural varieties of the St. Kevin and abandoned the name. Stark and Barnes (1935) also mapped large parts of the batholith as "schist and migmatite" because of evidence of assimilation or metasomatism of gneisses. The rocks of the entire batholith are herein assigned to the St. Kevin Granite.

The St. Kevin Granite is an inhomogeneous assemblage of two-mica rocks ranging in composition from granite to granodiorite and varying in texture and fabric. In some parts of the batholith, the granite contains abundant inclusions in various stages of assimilation, and in many places the rocks are heavily contaminated with materials residual from the gneisses, giving them a streaky appearance. Near some contacts, the contaminants outline folds and crenulations that are continuous with structural features in the wall-rock gneisses, indicating that granite replaced gneiss metasomatically. The contact between the batholith and its wall rocks is extremely irregular, giving the batholith a ragged outline (Tweto and Pearson, 1964, fig. 127.2). Large parts of the contact are grossly concordant, but locally they are highly discordant.

Principal varieties of the St. Kevin were called the trachytoid hybrid, the normal, the granodioritic, the fine-grained, and the coarse porphyritic facies by Tweto and Pearson (1964). Each of these facies has several variants, and each occurs in clean and in streaky, or contaminated, forms. Various pairs of the facies intergrade in some places, but they are clearly sequential in others. Where sequence is evident, a facies A may cut a facies B in one place, but in another place B may cut A. The trachytoid hybrid facies consists principally of closely packed parallel crystals of microcline 3 cm (1.2 in.) long together with coarse flakes of muscovite. The crystals lie in a matrix of fine-grained biotite gneiss or its derivatives. The facies occurs in border zones of the batholith or of inclusions, and it may grade to or be cut by other facies. It clearly is a product of reaction at an advancing granite front. The so-called normal facies, the principal rock of the batholith, is a light-gray to light-pink equigranular to markedly porphyritic muscovite-biotite granite or quartz monzonite. In the porphyritic varieties, tabular microcline phenocrysts are oriented in trachytoid or flow structure, which commonly parallels contacts with bordering facies or with gneiss. Mica-rich wisps and streaks, descended from biotite gneiss, are abundant in some localities. Even where the normal facies appears compositionally uniform, it may contain microscopic remnants of gneiss, such as sillimanite, garnet, and biotite. The granodioritic facies, the darkest rock, is fine to medium grained, equigranular to moderately porphyritic or seriate porphyritic, and weakly foliated. It occurs principally in the central interior and southwestern parts of the batholith. The fine-grained facies is a gray, buff-weathering, equigranular, massive to weakly foliated muscovite-biotite quartz monzonite. It occurs in border areas of the batholith and in the central interior where it grades to the granodioritic facies. The coarse porphyritic facies is muscovite-biotite granite or quartz monzonite characterized by abundant blocky microcline phenocrysts about 2 cm (1 in.) long. This is the rock that was called Hell Gate Porphyry by Stark and Barnes (1935, pl. 1), but it is much less widespread than was portrayed by those authors. It occurs principally in layerlike bodies or lenses in gradational contact with layers or lenses of the normal and the granodioritic facies in the central interior of the batholith. All facies in the central interior contain inclusions of dark quartz diorite.

The rock in the southwestern part of the batholith is fairly homogeneous, and as described by Bryant (1979, p. 9) is an equigranular to locally porphyritic muscovite-biotite quartz monzonite or granodiorite. I include also in the batholith a small body of coarse-grained muscovite-biotite granite described by Bryant (1979, p. 9), even though this granite seems to be in gradational contact with a porphyritic quartz monzonite that is assigned to the Routt Suite. An easternmost element of the batholith, east of the Mosquito fault in the Mosquito Range, was described as the Treasurevault stock by Kuntz and Brock (1977). Quartz monzonites of this body, illustrated by Kuntz and Brock (1977, fig. 6), appear identical to varieties that are common in the St. Kevin in the Sawatch Range. Rocks of the granite body at Climax, described by Butler and Vanderwilt (1933, p. 208), consist in major part of elongate potassium feldspar crystals and resemble the trachytoid hybrid facies of the St. Kevin in the Sawatch Range. Rocks in the part of the batholith exposed at the surface and in mine workings at Leadville include some of the trachytoid hybrid facies, but consist principally of intergrading streaks or lenses of the normal and the coarse porphyritic facies. They were described petrographically by Emmons and others (1927, p. 24).

Modes and chemical analyses of the St. Kevin may be found in the reports by Tweto and Pearson (1964, p. 30), Bryant (1979, table 1), and Kuntz and Brock (1977, table 3). The latter report also contains data on feldspar compositions and petrogenesis. Chemical analyses, without modes, were also provided by Butler and Vanderwilt (1933, table 1).

The St. Kevin is one of the most thoroughly dated Precambrian granites in Colorado, many age determinations having been made by K-Ar, Rb-Sr, and U-Th-Pb (uranium-thorium-lead) methods. Its age as determined from a Rb-Sr isochron plot for seven whole-rock samples is about 1,432±60 m.y. (Pearson and others, 1966). The age as determined from a concordia plot of U-Th-Pb ratios in zircons is about 1,400±40 m.y. (Doe and Pearson, 1969). As often is found of Precambrian rocks, K-Ar determinations on micas yielded somewhat younger ages, in the range 1,210–1,358 m.y. (Pearson and others, 1966, table 5). Kuntz and Brock, however, (1977, p. 467) reported a K-Ar age of 1,437±45 m.y. for the Treasurevault stock.

SYENITIC ROCKS IN GUNNISON-SAN JUAN REGION

In the Cebolla Creek–Powderhorn area southwest of Gunnison (Tps. 47–48 N., Rs. 2–3 W., pl. 1), syenitic rocks form numerous plugs and small irregular intrusive bodies. Rocks of this group of intrusives are classified as augite syenite, biotite leucosyenite, hornblende and biotite melasyenites, shonkinite, and minette (Olson and Hedlund, 1973; Hedlund and Olson, 1973; Hunter, 1925). Some of the intrusive bodies consist of only one of these rocks but others consist of two or more. Many of the composite bodies consist of an outer zone of melasyenite and a central zone of more felsic rock, such as leucosyenite (Olson and others, 1977, p. 676). Detailed petrographic descriptions of the syenitic rocks and chemical analyses of an augite syenite and a shonkinite are in the report by Hunter (1925, p. 64–76). K-Ar ages of 1,387 and 1,397±40 m.y. were obtained from biotite from three of the syenitic plugs, and a biotite age of 1,338±36 m.y. came from a minette dike (Olson and others, 1977, table 1). Thus, the syenitic rocks clearly are of the age of other rocks of the Berthoud Suite and much older than the alkalic rocks of Iron Hill (fig. 3) in the same general area (Olson and Marvin, 1971). (See section on latest Proterozoic and Cambrian rocks.) The two episodes of alkalic intrusion in the same area are interpreted to reflect control by the Cimarron–Red Rock fault zone (fig. 3), which evidently predated the Berthoud Suite and extended to lower crustal or mantle depths.

Melasyenite in the San Juan Mountains south of the Gunnison region was described by Larsen and Cross (1956, p. 29) and by Barker and others (1970). The melasyenite is exposed only in small windows through the Tertiary volcanic rocks along Ute Creek, north of the eastern Eolus batholith (fig. 3), in T. 40 N., R. 5 W. The rock is seriate porphyritic and consists principally of hornblende, biotite, microcline microperthite, and oligoclase. Chemical analyses (Barker and others, 1970, table 1) show compositions almost identical with those of augite syenite and shonkinite in the Gunnison area as reported by Hunter (1925, p. 65, 73). Age of the melasyenite is essentially the same as that of the syenitic rocks in the Gunnison area, as indicated by K-Ar determinations on biotite of 1,417±50 and 1,386±50 m.y. (Barker and others, 1970, table 2). The Ute Creek occurrence is on the line of projection of a major east-trending fault in the basement beneath the Tertiary volcanic rocks (pl. 1). This suggests fault control of the intrusion, just as the Cimarron fault localized syenitic intrusion in the Gunnison area.

VERNAL MESA QUARTZ MONZONITE

The name Vernal Mesa was applied by Hunter (1925) to granitic rocks exposed in a belt about 1.6 km (1 mi) wide striking northeast across the Black Canyon of the Gunnison in eastern Montrose County (pl. 1). The distinctive pinkish-gray coarse-grained porphyritic rock of this occurrence later was recognized in scattered exposures on the Uncompahgre Plateau, in westernmost Colorado and bordering Utah (fig. 2), where the name Vernal Mesa was also applied (Shoemaker, 1956, p. 56; Hansen and Peterman, 1968; Hedge and others, 1968). As interpreted herein on the basis of surface exposures, borehole data, and magnetic data, the Vernal Mesa forms a large batholith in the area of the Uncompahgre Plateau and westward into Utah (fig. 3). The batholith has a positive magnetic signature and is marked by numerous high-amplitude magnetic anomalies, though as noted by Case (1966, p. 1437), the Vernal Mesa is not uniformly magnetized. A long linear magnetic high branches north-northeast from the main magnetically high area (Zietz and Kirby, 1972), and at the Black Canyon the linear high coincides with the Vernal Mesa. On the basis of this feature, a narrow dikelike arm of the batholith is interpreted to extend north-northeast about 95 km (60 mi). At the Black Canyon, this arm is grossly concordant with bordering gneisses (Hansen, 1971), but its great length suggests that it follows a major preexisting fault.

The Vernal Mesa in the Black Canyon area is coarsely porphyritic biotite quartz monzonite and granodiorite containing abundant blocky crystals of pink microcline as much as 5 cm (2 in.) long and fairly abundant biotite (Hansen, 1971). These minerals have parallel orientation that defines a flow foliation, which parallels contacts. The contacts, in turn, grossly parallel the foliation in bordering gneisses. Inclusions of the gneisses occur abundantly in the quartz monzonite near the contacts. As exposed in Unaweep Canyon on the Uncompahgre Plateau (fig. 2), the Vernal Mesa is complexly intertongued with older rocks and with the slightly younger Curecanti Granite. The Vernal Mesa there is foliated to massive and is varied in its content of microcline phenocrysts and biotite. The Vernal Mesa farther west, near the Utah border, is more uniform and massive.

Although a major rock unit, the Vernal Mesa is inadequately investigated. Detailed field studies and petrographic and chemical analyses of it are yet to be made, but some age data are available. Samples from the Black Canyon and Unaweep Canyon together define a whole-rock Rb-Sr isochron that indicates an age of 1,450±40 m.y. (Hansen and Peterman, 1968, fig. 4; Hedge and others, 1968, fig. 5). An aberrant age similar to that of the Routt Suite reported by Mose and Bickford (1969) clearly was due to the inclusion of rocks other than the Vernal Mesa among the samples from the complexly intruded eastern part of Unaweep Canyon used to define the Rb-Sr isochron. In a later study, Bickford and Cudzilo (1975) determined a U-Pb zircon age of about 1,422±22 m.y. for a sample from the western part of Unaweep Canyon, where relations among the granites are also complex.

CURECANTI QUARTZ MONZONITE

The Curecanti Quartz Monzonite (or Granite), named by Hunter (1925, p. 49), forms a small pluton in the Black Canyon of the Gunnison in extreme western Gunnison County (figs. 2, 3), and is present also in a body of unknown extent that is partly exposed in Unaweep Canyon on the Uncompahgre Plateau. The pluton in the Black Canyon is of unusual shape, being a flat-lying, slightly domed, markedly discordant sheet that at its west edge plunges downward in a throatlike feeder (Hansen, 1964, figs. 2, 3). Exposures in the walls of the canyon and tributaries indicate that the sheetlike pluton occupies an area of about 18 km^2 (7 mi^2), though much of it is covered by roof rocks of gneiss. Nearby, numerous discordant pods and thin lenses of a facies of the Curecanti are scattered through an area several times the size of the pluton (Hansen, 1971), suggesting a more extensive body of the quartz monzonite at shallow depth. The Curecanti in the pluton is a sodic biotite-muscovite quartz monzonite that grades locally to granodiorite. Euhedral garnet is a distinctive accessory mineral, an unusual occurrence among the Precambrian granitic rocks of Colorado. The Curecanti is light gray to pink, equigranular, and medium grained. Modes and chemical analyses of the rock are available in a report by Hansen (1964, tables 4, 5). A whole-rock Rb-Sr isochron plot of four samples of the Curecanti from the Black Canyon indicates an age of about 1,392±15 m.y. (Hansen and Peterman, 1968, fig. 5).

Garnet-bearing sodic muscovite quartz monzonite near the east end of Unaweep Canyon closely resembles the Curecanti of the Black Canyon (Shoemaker, 1956, p. 56; Hedge and others, 1968) and was called Curecanti-type by Mose and Bickford (1969, p. 1679). Only the top of a pluton of this rock, intruded into Vernal Mesa Quartz Monzonite, is exposed in the canyon walls. Modes of the rock (Mose and Bickford, 1969, table 4) indicate a content of about 10 percent of muscovite and about 1 percent of biotite. The plagioclase is albite. A whole-rock Rb-Sr age of about 1,450±60 m.y.

was reported, but evidence of reaction with the Vernal Mesa in some of the sampled rocks was also noted (Mose and Bickford, 1969, fig. 4, table 4). A garnet-free porphyritic muscovite-biotite granite in the same area was also recognized by Mose and Bickford (1969), who called it Silver Plume-type granite. This rock cuts Vernal Mesa Quartz Monzonite, but its relation to the Curecanti is not established. A whole-rock Rb-Sr isochron age of about 1,443±57 m.y. was reported (Mose and Bickford, 1969, fig. 3). Thus, both muscovite-bearing rocks of the Unaweep area have yielded apparent ages greater than that of the Curecanti in Black Canyon, but considering the large analytical uncertainties, the disparity may not be real. Both areas are in or near the ancient west-northwest-trending Cimarron-Red Rock fault zone (fig. 3). The spotty occurrence of small bodies of Curecanti Quartz Monzonite and of the other two-mica granite along this fault zone suggests that localized tongues of granitic magma rose along the fault zone about 1,400 m.y. ago, but after intrusion of the Vernal Mesa Quartz Monzonite. Several hundred million years later, alkalic and mafic magmas rose along the fault zone. (See section on latest Proterozoic and Cambrian rocks.)

EOLUS GRANITE

The Eolus Granite is exposed in two bodies about 10 km (6 mi) apart in the Needle Mountains northeast of Durango (fig. 3). Together with inferred subsurface extensions, each body is a small batholith. The Eolus was named and described in some detail by Cross and Howe (Cross, Howe, and others, 1905, p. 8) on the basis of its occurrence in the western batholith. The name later was applied to the eastern batholith by Cross and Larsen (1935), who called the two batholiths together the Los Pinos or Pine River batholith. Larsen and Cross (1956) excluded a granodiorite facies in the two batholiths from the Eolus Granite, but Barker (1969a, p. 25) classed all rocks in both bodies as Eolus. Barker's practice is followed here, and in addition, a body of granite to the north, exposed near the head of the Lake Fork of Gunnison River, is assigned to the Eolus. Before faulting, this body may have been connected by a narrow neck to the eastern batholith (pl. 1). Melasyenite in the body on Ute Creek, discussed previously, is in the area of the neck. It is about the age of the Eolus and may be genetically related to the granite (Barker and others, 1970, p. 12).

The batholiths are grossly concordant with foliation in the Irving Formation in some areas, but are discordant in other places and in detail. They are markedly discordant with the Uncompahgre Formation and the Vallecito Conglomerate. The eastern batholith truncates the belt of isoclinally folded Uncompahgre (pl. 1), and small satellitic bodies of the Eolus cut both the Uncompahgre and the Irving. Long straight contacts between the batholiths and the intervening block of Vallecito and Irving rocks suggest control by old faults of near-north trend, although other evidence of such faults is lacking and probably was obliterated by the intrusions.

The Eolus Granite consists principally of pink to brick-red biotite-hornblende quartz monzonite, which is characterized by blocky crystals of pink microcline microperthite 1–3 cm (0.4–1.2 in.) long and by bluish-white quartz (Barker, 1969a, p. 25; Larsen and Cross, 1956, p. 29). Local areas of a granodiorite facies of this rock, containing as much as 30 percent of hornblende and biotite, are present in both batholiths. In the eastern batholith, the granodiorite contains swarms of inclusions rich in dark minerals, probably reflecting an early and more mafic phase of intrusion. Biotite quartz monzonite and biotite granite were intruded into the hornblende-bearing quartz monzonite. They occur in a moderately large body in the southern part of the western batholith and in small bodies elsewhere in both batholiths. These rocks are lighter in color than the biotite-hornblende quartz monzonite, and the perthitic microcline phenocrysts in them have tabular rather than blocky forms. Chemical analyses of the various facies in the two batholiths are reported by Barker (1969a, table 2). Age of the Eolus has been firmly established by a combination of U-Pb and Rb-Sr determinations. Age of quartz monzonite in the western batholith is about 1,438±20 m.y. as dated by U-Pb in zircon (Silver and Barker, 1968). Whole-rock Rb-Sr determinations on samples from both batholiths yielded a similar age, 1,444±27 m.y. (Bickford and others, 1969).

The granite body to the north of the batholiths, near the head of Lake Fork of Gunnison River, is a porphyritic sodic biotite granite that contains secondary muscovite. Larsen and Cross (1956, p. 32) judged the granite to be older than the Eolus in the batholiths to the south because it was more cataclasized and altered, but these features may merely reflect the location on the borders of Tertiary calderas.

TRIMBLE GRANITE

The Trimble Granite has been recognized only in a stock about 13 km^2 (5 mi^2) in area in the center of the western Eolus batholith (fig. 3) in the Needle Mountains. The granite was named and described by Cross and Howe (Cross, Howe, and others, 1905, p. 8) and was described further by Barker (1969a, p. 27). The Trimble is a pink to light-red, fine- to medium-grained, equigranular granite that contains small amounts of muscovite. Near contacts, it contains scattered microcline phenocrysts. A chemical analysis (Barker, 1969a, table 2) indicates a composition similar to the biotite granite facies of the Eolus Granite. Although this similarity and the occurrence as a stock within the Eolus batholith suggest relationship to the Eolus Granite, available age data indicate that the Trimble is considerably younger than the Eolus. A whole-rock Rb-Sr age of about 1,323±50 m.y. is reported for the Trimble (Bickford and others, 1969, fig. 5), in contrast to the Eolus age of about 1,440 m.y.

ELECTRA LAKE GABBRO

The Electra Lake Gabbro is known only in a pluton having an exposed area of about 16 km^2 (6 mi^2) at the west edge of the Needle Mountains, about 35 km (22 mi) north of Durango (pl. 1). The pluton has nearly planar vertical contacts that cut across the structure of the bordering Twilight Gneiss and Irving Formation. The gabbro was first described by Cross (Cross and Hole, 1910, p. 3) and was named and described further by Barker (1969a, p. 27). The Electra Lake Gabbro is an inhomogeneous unit consisting of intergrading facies ranging in composition from olivine gabbro to biotite-augite granodiorite. Augite and hypersthene gabbros are most characteristic. The rocks are dark, massive, and predominantly coarse grained. They are cut by many dikes of pink to red granophyre, alaskite, and pegmatite. A whole-rock Rb-Sr isochron age for the gabbro of about 1,425±50 m.y. (Bickford and others, 1969, fig. 9) is not much different from that of the nearby Eolus Granite. One of the granophyric alaskite dikes cutting the gabbro yielded a U-Pb age of 1,438±20 m.y. (Silver and Barker, 1968). Considering the analytical uncertainties, this is essentially the same as the age of the gabbro as well as of the Eolus Granite.

SUBSURFACE UNITS

In the subsurface, rocks of the Berthoud Plutonic Suite occur widely in the south half of Colorado but are uncommon in the north half except as noted previously (pl. 1). In addition to the widespread Vernal Mesa Quartz Monzonite in the subsurface of southwest Colorado, a unit consisting of micrographic granite occupies the extreme southwest corner of the State. Knowledge of this granite comes from three closely spaced boreholes (pl. 1) and the presence of the granite as xenoliths in Laramide intrusives to the north (Ekren and Houser, 1965, p. 33). These are the only occurrences of micrographic granite known in the Precambrian of Colorado. The micrographic or granophyric texture suggests shallower and probably younger intrusion than that of most granites of the Berthoud Suite. A Rb-Sr feldspar age of 1,810 m.y., reported from one borehole (Fitzsimmons, 1963), is rejected because the rock lacks any of the features of rocks that old. The rock is tentatively included in the Berthoud Suite, though it might be younger.

Granitic rocks of the Berthoud Suite constitute the Precambrian basement in an area of more than 11,700 km^2 (4,500 mi^2) in Las Animas and bordering counties in southeastern Colorado, and they probably underlie much of the Middle Proterozoic Las Animas Formation also (pl. 1). The rocks are part of an extensive granitic terrane that extends from Colorado into bordering parts of New Mexico, Oklahoma, and Kansas. Samples from four boreholes in this terrane yielded whole-rock Rb-Sr ages in the range 1,300–1,330 m.y. (Muehlberger and others, 1966, fig. 1). Considering the inadequacies commonly experienced with borehole samples, these ages are regarded as minimal. The rocks of the region are granite, quartz monzonite, and local granodiorite as interpreted from numerous boreholes. They are characterized by hypautomorphic granular fabric and commonly by zoned plagioclase. Petrographic descriptions of numerous samples may be found in the report by Edwards (1966).

MAFIC AND INTERMEDIATE DIKES

Mafic and intermediate dikes, mostly of about the age of the Berthoud Suite, are present in places in the Front, Park, and Sawatch Ranges. Although the dikes constitute only a miniscule fraction of the basement terranes, they are significant for the insights they furnish regarding Precambrian tectonic and magmatic histories. A sparse swarm of such dikes cuts through the east flank of the northern Front Range and extends into the Laramie Range of Wyoming (Braddock, Calvert, and others, 1970; Braddock, Nutalaya, and others, 1970; Abbott, 1976; Eggler, 1968; Ferris and Kreuger, 1964). Many of the dikes follow or parallel Precambrian faults of north-northwest trend. The dikes consist variously of andesite, basalt, and dacite. The andesites and dacites are commonly porphyritic, and the basalts have diabasic fabrics. Geologic relations bracket the age of the dikes within the time frame of the Berthoud Suite. The dikes cut the Sherman Granite of the Berthoud Suite, but they are cut by Silver Plume Granite (Peterman and others, 1968, p. 2288; Eggler, 1968, p. 1549). Hornblende and biotite K-Ar ages of the dike rocks (Peterman and others, 1968; Ferris and Kreuger, 1964) are consistent with the geologic relations. Rocks of the same character as the dike swarm occur in the ring-dike zone of the circular structure noted in the discussion of the Sherman Granite. This structure and its ring dikes originated during emplacement of the Sherman (Eggler, 1968). The north-northwest-trending dike swarm, presumably derived from the same deep source as the ring dikes, must have been emplaced only shortly after the Sherman.

Southwest of the dike swarm, a remarkably persistent diabase dike, or zone of closely spaced dikes, slices through the Front Range at a small angle to the trend of the range. This dike or zone, long known as the Iron Dike, trends north-northwest parallel to the dike swarm to the northeast and to a system of Precambrian faults. Except for some local gaps, the dike extends from a point near Boulder at least to the Wyoming border on the east side of the Medicine Bow Mountains, a distance of 128 km (80 mi) (pl. 1). A mafic dike of this length must reflect a major fracture deep in the crust. In many places, the Iron Dike is a set of two, three, or four dikes in a narrow zone in and near major Precambrian faults. The individual dikes locally reach thicknesses of as much as 45 m (150 ft) but commonly are 3–15 m (10–50 ft) thick. They dip 60°–80° SW. The character of the dike zone is well illustrated in a map by Gable (1980b). Wahlstrom (1956) described the diabase in detail and supplied several chemical analyses. As described by him, the diabase is chilled at the dike borders and has flow structure expressed by a preferred orientation of elongate grains. This structure everywhere plunges straight down dip, indicating that the diabase magma rose vertically throughout the length of the dike and thus that the deep source was widespread. The diabase consists principally of labradorite and augite but is notable for its high content of titaniferous magnetite, which averages about 8.5 percent. The Iron Dike cuts through the Silver Plume Granite of the Longs Peak-St. Vrain batholith, and near the Wyoming border it also cuts the Sherman Granite. Thus, it is younger than those granites, but its age is otherwise uncertain. Attempts to date it radiometrically have been inconclusive. Although the diabase was long assumed to be of Laramide age (Lovering and Goddard, 1950, fig. 12), its trend is inconsistent with Laramide tectonic forces, and in recent years most workers in the region have interpreted a Precambrian age. A paleomagnetic investigation by E. E. Larsen suggests a late Precambrian age (Gable, 1980b), and the parallels in trend and occurrence with the dike swarm to the northeast suggest an age near that of the Silver Plume Granite.

In the Park Range, Snyder (1978, 1980b) found dikes of fine-grained andesitic porphyry to be both older and younger than the Mount Ethel pluton of the Berthoud Suite. Dikes of the older group occur mainly in the Soda Creek-Fish Creek mylonite zone, where they strike north-northwest, almost normal to the mylonite zone, and are displaced by faults within that zone. Dikes younger than the pluton are more widespread and their predominant trend is northeast. All the dikes are short, few exceeding 1 km (0.6 mi) in length. The rocks of the dikes are andesites but they show considerable range in composition. Chemical analyses are available in the report by Snyder (1978, table 5). Dikes in the Northgate district, at the north end of North Park, may be related to the older set in the Park Range. As described by Steven (1957, p. 364), they strike a little west of north, consist of dacite porphyry, and are older than the quartz monzonite of the Berthoud Suite in that area.

On the northeast flank of the Sawatch Range, scores of metalamprophyre dikes cut the Precambrian rocks in the vicinity of the Homestake shear zone (Tweto, 1974; Pearson and others, 1966, fig. 2). Most of the dikes are in a swarm about 3 km (2 mi) wide that trends east-northeast from an intersection with the northeast-trending Homestake shear zone near the crest of the range. Many of the dikes cut St. Kevin Granite and related pegmatites, but a few seem to be cut by the granite and pegmatite, suggesting that the metalamprophyres are of more than one age. Most of the dikes are less than 3 m (10 ft) thick and less than 600 m (2,000 ft) long, but a few are 10–15 m (30–50 ft) thick and 3 km (2 mi) long. The dikes are weakly foliated parallel to their walls and independent of foliation in bordering gneisses. Although the dike rocks have been almost completely recrystallized, chill zones are commonly evident. The recrystallized rocks consist of abundant hornblende and biotite accompanied by lesser amounts of andesine, potassium feldspar, and quartz. Relics of original pyroxene and a calcic plagioclase are also present, and in some dikes apatite and magnetite or ilmenite are abnormally abundant. The original rocks were lamprophyres such as spessartite, minette, and vogesite (Pearson, 1959). The metalamprophyres are younger than a hornblende diorite that occurs in scattered dikes both among the metalamprophyres and within individual shear zones of the Homestake shear zone. The shear zones also contain scattered pods of hornblendite that probably is a recrystallization product of mafic or ultramafic rocks. Thus, the main group of metalamprophyres seems to be a late manifestation of mafic magmatism that occurred intermittently along the Homestake shear zone. An attempt to date the metalamprophyre by the K-Ar method on biotite yielded an apparent age of $1{,}328 \pm 66$ m.y. (Pearson and others, 1966, p. 1116), but considering the recrystallized state of the dikes, that age can be only a minimum. Although many of the dikes are observed to be younger than the St. Kevin Granite (1,400–1,432 m.y. in age), they may be only slightly younger.

Small lamprophyre dikes occur in many other places but are too sparse and of too uncertain age to have received much attention. Among the few that have received some study are dikes along Clear Creek, west of Golden. These dikes lie close to and parallel to Precambrian faults of west-northwest trend. The dikes are a little less mafic than those in the Sawatch Range, but like the latter, they are foliated parallel to their walls and are inferred to be of about the age of the Silver Plume Granite (Sheridan and others, 1972; Scott, 1972; Taylor, 1975a; Sheridan and Marsh, 1976). Metamorphosed lamprophyre dikes in the Black Canyon of the Gunnison are younger than the Pitts Meadow Granodiorite of the Routt Suite and older than the Curecanti Quartz Monzonite of the Berthoud Suite (Hansen and Peterman, 1968). Unmetamorphosed and presumably younger lamprophyre dikes in the same area are undated.

ROCKS OF THE PIKES PEAK BATHOLITH

About 1,000 m.y. ago, a third period of major Precambrian igneous activity in Colorado produced the Pikes Peak Granite and related rocks. Insofar as they are known at the surface, the rocks of this period are almost entirely limited to the Pikes Peak batholith and a related pluton to the northwest. The batholith (fig. 3) occupies an area of about 3,100 km^2 (1,200 mi^2) in the southern Front Range, and on the basis of magnetic features, it is inferred to extend through a larger area in the subsurface to the east (pl. 1). A 143-km^2 (55-mi^2) pluton of Pikes Peak Granite northwest of the batholith was long known as the Rosalie lobe of the Pikes Peak batholith, but because of geographic and stratigraphic confusion over the name Rosalie, the pluton was renamed the Lone Rock pluton (fig. 3) by Bryant and Hedge (1978, p. 448). Other than the batholith and the smaller pluton, rocks of Pikes Peak age in Colorado are known only in a few syenitic and granitic dikes in the east-central Front Range and along the Gore fault on the west side of the Gore Range (Taylor, 1975a; Sheridan and Marsh, 1976; Barclay, 1968).

The Pikes Peak batholith is an epizonal intrusion that truncates structure in bordering gneisses and in the granitic rocks of the Routt and Berthoud Suites. Despite these features and differences in lithology, the batholith was assumed for many years to include at its south end a body of gneissic granodiorite and augen gneiss. These rocks later were recognized to be much older than the batholith and now are assigned to the Routt Suite. But having been interpreted as part of the batholith, the old rocks were classed in the past as Pikes Peak Granite (Mathews, 1900; Finlay, 1916; Lovering and Goddard, 1950). This led to misassignments elsewhere, as noted previously, and contributed to the erroneous belief that the Pikes Peak was older than the Cripple Creek or Silver Plume Granite.

Structural studies of the batholith (Hutchinson, 1960, 1972, 1976) indicate three areas of concentric and radial orientation of aplite and pegmatite dikes and of steep concentric flow structure. These areas, two in the northern part of the batholith and one in the southern, are each 20–25 km (12–15 mi) in diameter and are interpreted as intrusive centers (fig. 4). They may also mark the underpinnings of former calderas, consistent with the conclusion (Barker and others, 1975) that the batholith was emplaced at shallow depth and may have breached the surface. Segments of circular structure are also evident in the aeromagnetic map, particularly in the southern part of the batholith (Zietz and Kirby, 1972). The exposed part of the batholith coincides with a pronounced magnetic low. This magnetic low continues eastward into the Plains (Zietz and Kirby, 1972), where the Precambrian surface is as much as 6 km (20,000 ft) lower (and more distant from the observation level) than in the mountains. On the basis of this magnetic pattern, the batholith is inferred to extend eastward far beyond the mountain front (pl. 1). If this inference is correct, the batholith is elongate nearly east-west rather than in the northerly direction suggested by surface exposures. The exposed part of the batholith corresponds with a postive anomaly in an Airy-Heiskanen gravity anomaly map (Qureshy, 1958), but it corresponds with a poorly defined gravity trough in a Bouguer anomaly map (Behrendt and Bajwa, 1974).

The Pikes Peak batholith is a composite body consisting of several different rock units, some of which are in discrete plutons (Barker and others, 1975; Barker, Hedge, and others, 1976; Wobus, 1976). The principal component of the batholith is coarse-grained pink Pikes Peak Granite. Local facies and plutons within the body of this granite bear various formal and informal names (Bryant and others, 1981; Scott and others, 1978). A few of these units are older than the Pikes Peak Granite but most are younger. All the units are cogenetic (Barker and others, 1975; Gross and Heinrich, 1965), and all probably were emplaced within a period of less than 20 m.y. (Hedge, 1970). The Rb-Sr age of the batholith was placed at about 1,020 m.y. by C. E. Hedge (written commun., 1982) after minor adjustment of an earlier figure. This age is in agreement with a Lu-Hf determination (Patchett and others, 1981, table 1).

PIKES PEAK GRANITE

Granite constituting the main body of the Pikes Peak batholith was named the Pikes Peak Granite by Cross (1894) and was described in some detail by his as-

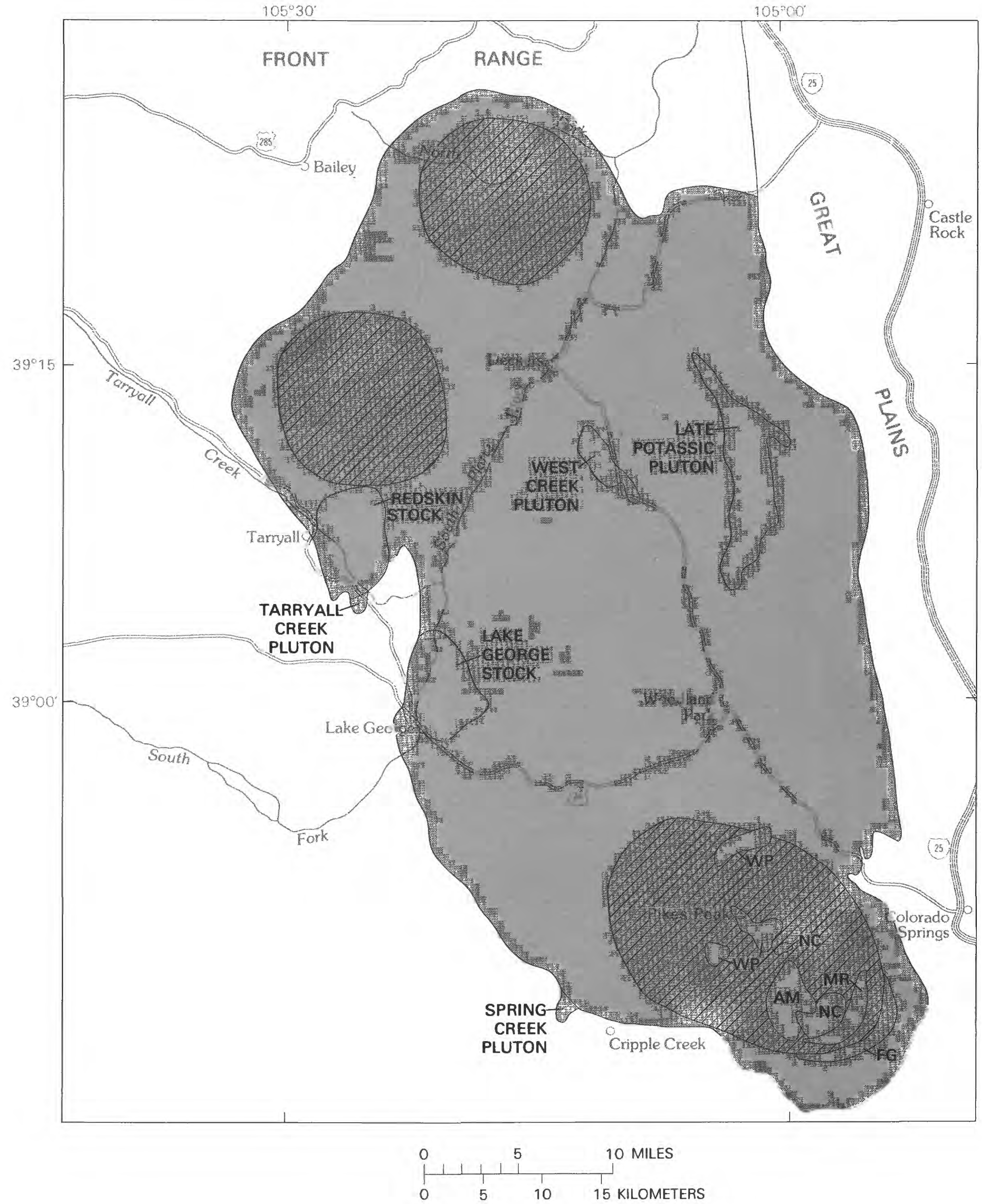


FIGURE 4.—Sketch map of intrusive bodies and centers in exposed part of Pikes Peak batholith (shaded) in southern Front Range. Intrusive centers patterned. FG, arc of bodies of fayalite granite; MR, Mount Rosa Granite; NC, granite of Nelson Camp; AM, granite of Almagre Mountain; WP, Windy Point Granite. Compiled from data in Wobus and Anderson (1978, fig. 1); Hutchinson (1976, fig. 4); Barker and others (1975, fig. 2); and Scott and others (1978).

sociate, Mathews (1900). This granite is pink to pale orange, coarse grained, massive, and homogeneous. It disintegrates readily to a coarse orange-red gruss that is visible along many roads and highways in the area. The granite consists of perthitic microcline, sodic plagioclase, quartz, and a small amount of biotite. A little hornblende and magnetite are present in border facies (Hutchinson, 1960, table 2). Modes and chemical analyses of the granite are available in several reports (Barker and others, 1975, tables 3, 4; Wobus and Anderson, 1978, table 3; Gross and Heinrich, 1965, table 1; Hutchinson, 1960, tables 1, 2). Many other rocks in the batholith have been excluded from the Pikes Peak Granite in recent mapping (Scott and others, 1978; Bryant and others, 1981). These rocks are discussed in following sections.

ASSOCIATED SUBALKALIC ROCKS

An intrusive center in the southern part of the batholith, southwest of Colorado Springs, is defined by steep concentric flow structure (Hutchinson, 1976) and by several varieties of slightly to moderately alkalic rocks (Barker and others, 1975). A prominent feature of this center is a long arc defined by small bodies of fayalite-bearing granite (fig. 4). This granite is interpreted to be older than the surrounding Pikes Peak Granite or to be partly contemporaneous with it (Gross and Heinrich, 1965, p. 1289; Barker and others, 1975, p. 114). Both granites contain inclusions of even older, dark, fine-grained hornblende monzonite and quartz syenite. Pikes Peak Granite bordering the arc is pierced by bodies of three younger granites. In probable order of decreasing age, these units are the granite of Almagre Mountain, the granite of Nelson Camp, and the Mount Rosa Granite.

The fayalite granite, first reported by Gross and Heinrich (1965), is a gray or olive-green medium-grained rock that forms resistant outcrops in the crumbly pink Pikes Peak Granite. Bodies of the fayalite granite range from 1 m (3 ft) to 1.5 km (1 mi) in greatest dimension. These bodies possibly are remnants of the roots of a ring dike (Barker and others, 1975, p. 115). The fayalite granite consists principally of microcline microperthite, quartz, oligoclase, and biotite. The content of fayalite, the ferrous iron end member of the olivine group, is small at about 1 percent, but is important as an indicator of the chemical and physical conditions at the time of crystallization (Barker and others, 1975). Trace amounts of ferrohornblende and hedenbergite are present in some bodies, absent in others. Fluorite is a characteristic accessory mineral. Modes and chemical analyses of the fayalite granite are given by Gross and Heinrich (1965) and Barker and others (1975).

The granite of Almagre Mountain (Barker and others, 1975, p. 120) is in an elliptical pluton about 6 km (4 mi) in length near the center of curvature of the arc of fayalite granite (fig. 4). The rock in the pluton is pink, medium-grained, massive, homogeneous biotite granite. This rock resembles the Pikes Peak Granite, but it lacks the oligoclase and is richer in fluorine than the Pikes Peak. The granite of Nelson Camp (Scott and others, 1978) occurs in two stocks at or near the border of the Almagre Mountain pluton. This rock, earlier called the tan granite (Barker and others, 1975, p. 120) is a light-brown, medium- to coarse-grained biotite granite that contains small amounts of riebeckite and astrophyllite. Flame perthite is the only feldspar in the rock, and the quartz is euhedral to subhedral. The Mount Rosa Granite was named by Finlay (1916) and was described further by Gross and Heinrich (1965). The granite occurs in elongate bodies totaling about 10 km^2 (4 mi^2) in an arc nested against the inner side of the arc of fayalite granite (fig. 4), suggesting repeated intrusion along a ring fracture. The Mount Rosa is a blue-gray, fine- to coarse-grained riebeckite granite composed principally of microcline perthite, quartz, and sodic plagioclase as well as the sodic amphibole riebeckite. The rock also contains a varied suite of accessory minerals, including aegerine, biotite, zircon, ilmenorutile, pyrochlore, monazite, bastnaesite, and astrophyllite among others. Cryolite-bearing riebeckite pegmatites in the nearby St. Peters Dome area (T. 14 S., R. 67 W.) are correlated with the Mount Rosa Granite (Finlay, 1916, p. 5; Steven, 1949, p. 265).

WINDY POINT GRANITE

Fine-grained porphyritic granite in irregular bodies that cut the Pikes Peak Granite was referred to as the "summit type" of granite by Cross (1894) and Mathews (1900). Finlay (1916) applied the name Windy Point to the fine-grained granite and indicated a Precambrian age. Lovering and Goddard (1950, p. 43, pl. 1) assigned an early Tertiary age to the granite, but Aldrich and coworkers (Aldrich and others, 1958, table 9) later established a Precambrian age of about 1,000 m.y. by K-Ar and Rb-Sr methods. The Windy Point is one of the youngest rocks in the batholith but is older than the Mount Rosa Granite (Scott and others, 1978). Wobus (1976, fig. 1) grouped the Windy Point and similar rocks together as "late potassic plutons," which he showed to be scattered throughout the batholith. The largest of the plutons, about 19 km (12 mi) long and 3 km (2 mi) wide, is northwest of Colorado Springs (fig. 4). The rock in this and the other plutons is pink, fine grained, and commonly porphyritic. Like most granites of the batholith, it contains accessory fluorite. Analytical data may be found in the report by Wobus

(1976, tables 1, 2), who noted a close compositional similarity to the Pikes Peak Granite (p. 59) and classed the rock as a "late, rapidly cooled, textural variant" of the Pikes Peak.

REDSKIN GRANITE

The Redskin Granite (Hawley and others, 1966) is a high-level alkali granite that is enriched in fluorine and contains abnormal amounts of several uncommon elements, such as tin, lithium, rubidium, beryllium, zirconium, tantalum, and niobium. Beryllium-bearing greisen deposits are associated with it. The granite forms an oval stock about 50 km^2 (19 mi^2) in area (fig. 4) and two small satellitic bodies at the western margin of the Pikes Peak batholith in Tps. 10–11 S. (pl. 1). The stock is concentrically zoned, consisting of an outer zone of granular granite, an intermediate zone of porphyritic granite, and an inner and youngest zone of fine-grained granite (Hawley and Wobus, 1977, pl. 1, p. 37). The granites in all three zones are red to pink, contain both muscovite and biotite, and are notably rich in fluorite or topaz. The muscovite content of the granites increases inward in the stock. The granular granite contains more microcline microperthite than sodic plagioclase, whereas in the other two granites the plagioclase predominates. Numerous modes and chemical analyses of the granites are available in the report by Hawley and Wobus (1977, table 15). Mineralogy of accessory components of the granites together with related analytical data were reported by Desborough and others (1980).

Veins and small areas of greisen occur in scattered localities within the stock and the two small satellitic cupolas to the west, and also in older rocks near their contacts with the Redskin Granite (Hawley, 1969). Many of the greisen bodies have been prospected, first for molybdenum, then for uranium, and finally for beryllium. A substantial output of ore containing the beryllium minerals beryl, bertrandite, and euclase was made from deposits in one of the cupola bodies of Redskin Granite (Hawley, 1969, p. 24).

The Redskin stock was intruded into two different units of the Pikes Peak batholith. A northern and larger unit, called the Tarryall Mountains batholith by Hawley and Wobus (1977, fig. 18), consists of coarsely porphyritic Pikes Peak Granite. This facies of the Pikes Peak contains noteworthy amounts of most of the minor elements found in the Redskin Granite. The south end of the stock was intruded into a small funnel-shaped body called the Tarryall Creek pluton (Bryant and others, 1981). This pluton consists of biotite-hornblende and muscovite-biotite quartz monzonites rimmed by olivine gabbro. Modes and chemical analyses of these rocks are contained in a report by Wobus (1976, tables 1, 2).

SYENITE AND GABBRO PLUTONS

Small plutons of syenite or of syenite and gabbro are scattered throughout the Pikes Peak batholith (Wobus, 1976, fig. 1). A stock adjacent to the village of Lake George, at the western margin of the batholith (fig. 4), neatly records progressive change in magma composition. The stock, about 6.5 by 8 km (4 by 5 mi) in diameter, is a composite of several different rock units in a concentric arrangement attributed to ring dikes and probably to subsidence of a former caldera (Wobus and Anderson, 1978, fig. 3). An outer ring consists of medium-grained granite that resembles the Windy Point Granite. Granite of this ring was intruded into coarse-grained Pikes Peak Granite and older rocks, and it encloses a ring of younger fine-grained granite. A ring dike of quartz syenite and minor fayalite granite occupies a segment of the contact between the two granite rings, and elsewhere extends into the ring of medium-grained granite. The core of the stock is a body of quartz syenite and syenomonzonite, the youngest rocks. This body contains large xenoliths of alkali gabbro, which is inferred by Wobus and Anderson (1978, p. 84) to be the oldest rock in the stock. Other units of the stock contain inclusions of diorite porphyry and granodiorite that probably are related to the gabbro. Analytical data on the several rock types in the stock are in the report by Wobus and Anderson (1978, tables 2, 3). Pegmatites in the outer ring of the stock are famous as sources of specimen amazonite (green microcline), smoky quartz, and topaz (Eckel, 1961, p. 28).

A pluton about 8 km^2 (3 mi^2) in area in the central part of the Pikes Peak batholith (fig. 4) consists of the quartz syenite of West Creek (Barker and others, 1975, p. 101). This pluton predates the Pikes Peak Granite as shown in several places where the granite crosscuts the syenite. The pluton consists chiefly of pink coarse-grained hornblende quartz syenite. Minor facies are greenish-gray to brown coarse-grained fayalite quartz syenite and a pale-pink medium-grained hornblende quartz syenite. Chemical analyses (Barker and others, 1975, table 1) show only slight differences in these varieties. Other and smaller bodies of syenite and of fayalite granite to the northwest and southeast postdate the Pikes Peak Granite (Wobus, 1976).

Olivine syenite and gabbro constitute a pluton about 5 km^2 (2 mi^2) in area that lies 3 km (2 mi) northwest of the town of Cripple Creek, at the southwest edge of the Pikes Peak batholith (fig. 4). This body, known as the Spring Creek pluton, forms a wartlike protuberance in the border of the batholith. The pluton cuts the Pikes Peak Granite along its east side, but it has yielded a Rb-Sr age about the same as those of the granite and other rocks of the batholith (Barker, Hedge, and others, 1976, p. 45). Rocks of the pluton were described in detail petrographically by L. C. Graton (*in* Lindgren and Ransome, 1906, p. 53–56) and were later studied by Barker and others (1975, p. 121), and Barker, Hedge, and others (1976, p. 45, table 1). The main body of the stock consists of greenish-gray coarsely crystalline olivine syenite that locally grades to clinopyroxene syenite. Fayalite is present locally in the syenite, as also is primary calcite. The syenite body is almost surrounded by a rim of dark fine-grained olivine-clinopyroxene gabbro. The gabbro is reported both to grade to the syenite (Barker, Hedge, and others, 1976, p. 45) and to cut the syenite (Wobus and others, 1976). Dikes of pegmatitic labradorite anorthosite and of alkali diabase and granophyre cut the syenite.

LATEST PROTEROZOIC AND CAMBRIAN ROCKS

After a long period that lacks any record of igneous activity in Colorado, intrusion resumed on a small scale at the close of the Proterozoic and continued into the Cambrian. At that time, the small alkalic and mafic Iron Hill stock and subsidiary dikes invaded the Powderhorn area southwest of Gunnison (pl. 1). The stock evidently rose along the Cimarron fault of the Cimarron–Red Rock fault zone (fig. 3), an ancient fracture zone that probably predated the Berthoud Suite. Tholeiitic diabase or gabbro dikes cut the stock and occur along the fault zone for a long distance to the northwest. The stock is a complex of pyroxenite, the melilite-rich rock uncompahgrite, the nepheline-pyroxene rock ijolite, magnetite-ilmenite-perovskite rock, nepheline syenite, the metasomatic rock fenite, and carbonatite (Larsen, 1942; Temple and Grogan, 1965; Olson and others, 1977; Olson and Hedlund, 1981). These rocks are of potential commercial interest for several mineral commodities. Thorium-bearing veins consisting of complex mixtures of carbonate, iron-oxide, sulfide, and alkali-feldspar minerals are spatially associated with the rocks (Olson and Hedlund, 1981; Olson and Wallace, 1956). Carbonatite in large bodies and dense dike swarms (Olson, 1974) contains appreciable amounts of the niobium-bearing mineral pyrochlore as well as some thorium and rare earths (Temple and Grogan, 1965). The pyroxenite and the magnetite-ilmenite-perovskite rocks contain large tonnages of titanium-iron material of marginal ore grade (Rose and Shannon, 1960), and some production of vermiculite has been made from biotite pockets in the pyroxenite.

On the basis of several determinations by K-Ar, Rb-Sr, and fission-track methods, intrusion at Iron Hill began 579 m.y. ago and continued to 565 m.y. ago (Olson and others, 1977, p. 679). Age of the base of the Cambrian is accepted to be 570 m.y. by most workers in the United States. Thus, the alkalic and mafic rocks of the Iron Hill area seem to bracket the Proterozoic-Cambrian boundary, though uncertainties in both the rock dates and the age of the boundary make close classification problematic. Alkalic and mafic rocks in small plutons in the Wet Mountains, 145 km (90 mi) east of the Iron Hill area, are 50 m.y. younger than the Iron Hill rocks and thus certainly of Cambrian age (Olson and others, 1977). The plutons in the Wet Mountains are shown as areas of no basement in plate 1. The diabase dikes at Iron Hill and northwestward also are probably of Cambrian age. Some of these dikes in the Black Canyon of the Gunnison have been dated as about 510 ± 60 m.y. (Cambrian or Ordovician) in age (Hansen and Peterman, 1968).

TECTONIC INFLUENCES

The interplay between rock-forming and tectonic events has been alluded to repeatedly in the preceding discussions. This interplay and the tectonic evolution of the Precambrian basement are to be the subjects of a report in preparation, but a summary of the relation of rock units to tectonic features is pertinent here.

The paleotectonic setting in which the sedimentary and volcanic protolith of the gneiss complex accumulated is only vaguely discernible. An older continental element, the Archean Wyoming province, now adjoins the province of Proterozoic rocks in Colorado, but whether the two provinces were in the same relative positions 1,800 m.y. ago when the protolith of the gneisses accumulated is uncertain. The petrographic character and isotopic compositions of the gneisses indicate that the protolith was not derived in any substantial part from older continental rocks, such as the Wyoming province, but they do suggest an origin in some sort of oceanic arc environment. A principal feature was an east-trending trough that is outlined by predominantly metasedimentary rocks flanked to the north and south by predominantly metavolcanic rocks. This trough extended the full width of Colorado and unknown distances beyond. Although now masked by tectonic features of many other trends, other evidence exists of a pattern of east-west structural organization in the crust beneath Colorado. The aeromagnetic maps of both Colorado (Zietz and Kirby, 1972) and Utah (Zietz and others, 1976) show a predominant east-west magnetic grain. This same grain is evident also in the Bouguer gravity map of Colorado (Behrendt and Bajwa, 1974), although it is extensively disrupted by other gravity features. The grain in both the magnetic and gravity maps matches the geology poorly, suggesting that it expresses features of the deeper crust. Such crustal features presumably controlled old and long faults of east-west trend that are present in some areas, as in the western San Juan Mountains region, the White River Plateau, and the northern Front Range (pl. 1).

Fault systems began to form at least as early as the time of folding and metamorphism of the gneiss complex, and they substantially influenced igneous intrusion thereafter. Some fault zones of northeast and north-northwest trends existed before intrusion of the Routt Suite, and some may have formed concurrently with intrusion. Northeast-trending fault or shear zones in the Colorado mineral belt (Tweto and Sims, 1963) are the best documented. There, the Homestake shear zone in the Sawatch and Gore Ranges and the Idaho Springs–Ralston shear zone in the Front Range coincide with large parts of the southeastern borders of the Cross Creek and Boulder Creek batholiths, respectively. The two shear zones originated during folding that preceded batholithic intrusion and later were reactivated recurrently (Tweto and Sims, 1963, p. 999, 1004). The two batholiths define a broken belt of intrusion along the shear zones of the mineral belt. Farther north, the large Rawah batholith (fig. 3) also trends northeast, parallel to the mineral belt shear zones. As exposed in the Park Range, the northwest side of the Rawah batholith borders the northeast-trending Soda Creek–Fish Creek shear zone (fig. 3). Like the shear zones in the mineral belt, this zone has a history of repeated movements in various directions (Snyder, 1978), and map relations (Snyder, 1980b) indicate that it predated the batholith. Map relations suggest another ancient northeast-trending tectonic boundary at the southeastern border of the large composite body of Routt Suite rocks near the center of the State (pl. 1). This border coincides with a belt of northeast-trending faults in the Arkansas River region from Cotopaxi through the area north of Canon City. The faults of this belt, transverse to the main faults of the region, are interpreted as products of post-intrusive movements along an ancient crustal flaw. Plutons of the Routt Suite are widespread north of this belt but sparse and small to the south (pl. 1).

Thus, the ancient northeast-trending fault zones, commonly expressed as shear zones, clearly influenced the intrusion of granitic rocks about 1,700 m.y. ago. The intrusive bodies in turn became elements of a fundamental tectonic grain of northeast trend that overprints the older east-west grain and is overprinted by tectonic features of other trends. The shear zones are elements of the Colorado lineament of Warner (1978), a wide belt of Precambrian faults traced by him from Arizona to South Dakota and speculatively to Minnesota. Thus, the northeast-trending shear zones are part of a tectonic zone of near-continental dimension that originated during folding and metamorphism of the gneisses more than 1,700 m.y. ago.

Many Precambrian faults of north-northwest trend are known, but only the Ilse fault in the Wet Mountains (fig. 3) shows evidence of an origin before intrusion of the Routt Suite. Like the northeast-trending shear zones, the Ilse has a complex history of repeated movements of various kinds (Taylor, 1975b; Singewald, 1966). Precambrian terranes on the two sides of the fault differ markedly. On the east side gneisses of upper amphibolite facies are warped into open folds (Taylor, 1974; Taylor, Scott, and others, 1975). On the west side, gneisses at the threshold of the granulite facies are isoclinally folded and refolded (Brock and Singewald, 1968). Small plutons of Routt Suite rocks on the two sides are similar in character and evidently were emplaced after the great fault displacement that is implied by the contrasts in the gneisses (R. B. Taylor, oral commun., 1973).

Plutons of the 1,400–m.y. Berthoud Suite further demonstrate the influence of the ancient fault zones on igneous intrusion. As noted previously, the St. Kevin batholith invaded and obliterated part of the Homestake shear zone. Farther northeast in the mineral belt, the Silver Plume batholith lies in the line of projection of the Homestake zone and is transected by northeast-trending faults that indicate late movements in that zone (pl. 1). In the Park Range and North Park, the Mount Ethel pluton lies alongside the Soda Creek-Fish Creek shear zone, which clearly controlled the location and trend of the pluton. In the Black Canyon of the Gunnison and the Uncompahgre Plateau, plutons of Curecanti Quartz Monzonite and of an unnamed muscovite granite evidently rose in the Cimarron–Red Rock fault zone, as did, much later, the mafic-alkalic Iron Hill pluton. At right angle to the Cimarron–Red Rock zone, a long narrow arm of the Vernal Mesa Quartz Monzonite suggests control by a fault of north-northeast trend.

The plutons of the Berthoud Suite in Colorado are elements of a broad belt of anorogenic granites of 1,400–1,500 m.y. age that extends generally northeastward across much of the continent (Silver, Anderson, and others, 1977; Emslie, 1978). Orogenic features of the scale and age of the belt are lacking, but in Colorado, as has been shown, preexistent orogenic features controlled the locations and trends of many of the plutons. Most pronounced of these features are the northeast-trending zones of dislocation that originated during regional folding and metamorphism prior to 1,700 m.y. ago. Other preexistent orogenic features may have guided intrusion elsewhere in the anorogenic belt, but they could scarcely account for the generation of the belt. As has been noted (Silver, Bickford, and others, 1977), the cause for the generation of the great volumes of magma introduced into the upper crust at that time has not yet been recognized.

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