Geology of the Farmington Canyon Complex, Wasatch Mountains, Utah

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By BRUCE BRYANT

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Rocks at the margin of the Archean continent strongly overprinted by Early Proterozoic metamorphism and plutonism

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GEOLOGY OF THE FARMINGTON CANYON COMPLEX, WASATCH MOUNTAINS, UTAH

By Bruce Bryant

ABSTRACT

The Farmington Canyon Complex, in its largest area of exposure in the Wasatch Mountains between Ogden and Bountiful, Utah, is divided into four principal map units: (1) quartz monzonite gneiss, (2) migmatite, (3) schist and gneiss, and (4) schist, gneiss, and quartzite.

Quartz monzonite gneiss forms three bodies separated by migmatite in the area between Ogden and Weber Canyons and forms exposures of the complex extending along the west side of the Wasatch Mountains 25 km north of the study area. The quartz monzonite gneiss generally has sharp contacts with the migmatite, but the smaller, more southerly bodies have gradational contacts and contain some layered and partly migmatitic rocks.

Adjacent to and south of the quartz monzonite gneiss is a unit of migmatite containing some layers of quartz monzonite gneiss like that in the large bodies to the north. Other rock types in the migmatite unit are gneisses containing various proportions of garnet, hornblende, biotite, feldspar, and quartz, less abundant layers of biotite schist, garnet-biotite schist, and sillimanite-garnet-biotite schist, and even rarer layers of quartzite.

Southward the rocks of the Farmington Canyon Complex gradually lose their migmatitic aspect and grade into a unit of schist and gneiss north of Farmington Canyon. This unit typically contains rather thinly layered biotite-feldspar-quartz gneiss and biotite schist. Layers of quartzite and feldspathic quartzite are common, especially near contacts with the unit of schist, gneiss, and quartzite. Layers of sillimanite-bearing schist are much more numerous than in the migmatite unit to the north.

At the head of Ward and Holbrook Canyons in the southern part of the area, numerous layers of white quartzite, as much as several meters thick, are intercalated with schist and gneiss, forming a distinctive map unit.

Lenses of amphibolite, metadiabase, and metagabbro, typically 1–3 m thick and a few meters long, occur throughout the complex. A few amphibolite bodies are as much as 10 m thick and 500 m long. Textures range from nemanoblastic to diabasic or ophitic with minor recrystallization.

Pegmatites are abundant throughout the complex. They are generally smaller and more sparsely distributed in the quartz monzonite gneiss than in the layered metamorphic rocks. They range from stringers and lenses, a few centimeters thick with sharp to indistinct borders, to sharply bordered crosscutting bodies as much as 200 m thick and 700 m long. They contain various proportions of quartz, plagioclase, and microcline, and, locally, biotite. Some pegmatites contain hornblende or garnet where the country rock contains those minerals.

Mineral composition and chemical data show that the protoliths of most of the layered metamorphic rocks in the complex were arkose, graywacke, quartzite, and shale derived from continental crust. Major and trace element data indicate that the amphibolites, metadiabases, and metagabbros were derived from tholeiitic basalts, and trace element data suggest that they erupted through oceanic rather than continental crust. Thus, the protoliths of the layered metamorphic rocks of the Farmington Canyon Complex were deposited on oceanic crust near a continental margin. Rb-Sr and Sm-Nd isotope ratios show that the Farmington Canyon Complex protoliths are Archean and contain crustal materials 2,800–3,600 m.y. old that were metamorphosed about 2,600 m.y. ago (Hedge and others, 1983).

The main metamorphic assemblage of the Farmington Canyon Complex is of amphibolite facies, ranging from the sillimanite-potassic-feldspar zone in the north to the sillimanite-muscovite zone in the south. The prevalence of hastingsitic hornblende in the northern part of the area suggests that oxygen fugacity was low there during this metamorphism.

Modal and chemical compositions of the quartz monzonite gneiss indicate that it is an early melting fraction, and isotopic studies suggest that it originated by melting of upper crustal rocks about 1,790 m.y. ago (Hedge and others, 1983). This is the approximate age of the Early Proterozoic metamorphism responsible for the main mineral assemblages in the layered metamorphic rocks and the quartz monzonite gneiss. The gneiss lacks many of the features of S-type granitoids because it was derived by melting of crust already dehydrated during a previous metamorphism about 2,600 m.y. ago.

In a few rocks, relict grains of hypersthene suggest an earlier metamorphism under granulite facies conditions. Whether this occurred during an early phase of the Early Proterozoic metamorphism 1,790 m.y. ago or during the Archean metamorphism 2,600 m.y. ago is not known.

The large-scale structural features in the layered metamorphic rocks are obscure because of lack of stratigraphic marker horizons and facing criteria. Minor folds formed during the Proterozoic metamorphism range from isoclinal to open and generally plunge gently to moderately west, but many variations occur. Biotite and hornblende form lineations generally parallel the axes of minor folds.

About 20 percent of the rocks of the Farmington Canyon Complex have been variously sheared and retrogressively metamorphosed under conditions of the chlorite zone. Zones of shearing range from a meter to a few tens of meters thick from Weber Canyon north, and southward they merge into a zone a kilometer or more thick. Variable interplay between cataclasis and recrystallization has resulted in rocks ranging from cataclastic gneiss to mylonite and phyllonite over short distances. Similar shearing and retrogressive metamorphism on Antelope Island west of the Wasatch Mountains took place before late Precambrian time. This similarity and the pervasiveness of shearing in large areas in the Farmington Canyon Complex in contrast with shearing along the Ogden thrust, which formed during the Sevier orogeny of late Mesozoic age, indicate that the shearing and retrogressive metamorphism may be older than the Sevier orogeny. This metamorphic event may have occurred between the main metamorphism and plutonism 1,790 m.y. ago and deposition of unmetamorphosed Proterozoic sequences about 1,000 m.y. ago.

Correlative basement rocks on Antelope Island west of the Wasatch Mountains have a history resembling that in the Wasatch Mountains except that they contain granite gneiss about 2,020 m.y. old (Hedge and others, 1983). Metamorphic rocks near Santaquin correlated with the Farmington Canyon are south of the margin of the Archean

continent and probably Early Proterozoic in age. The Little Willow Formation southeast of Salt Lake City may also be Early Proterozoic.

The Ogden thrust, which outlines the northern part of Farmington Canyon Complex in the Wasatch Mountains, formed during the Sevier orogeny in the late Mesozoic. One branch that passes north and east of, and tectonically above the complex may connect with faults at Durst Mountain and Hardscrabble Creek and may be older than the Absaroka and Mt. Raymond thrusts. The other, tectonically lower branch passes into the Farmington Canyon Complex on the west flank of the Wasatch Mountains and may die out southward. Although the Farmington Canyon Complex is parautochthonous in relation to the far-traveled Willard thrust sheet overlying it to the north, regional interpretations of geology and seismic data seem to require a major thrust beneath the Farmington Canyon.

Neogene high-angle faults and fractures bound and cut the Farmington Canyon Complex. The Wasatch fault separates the Wasatch Mountains block from the Salt Lake valley and the Farmington graben on the west. A complex system of faults separates the Wasatch Mountains from the Morgan valley on the east and dies out to the south where the Morgan basin loses its topographic expression.

Quartz veins of Precambrian age containing chalcopyrite, galena, or sphalerite, and reported to contain some gold and a little silver, have been locally mined on a small scale. Some of them cut phyllonite zones, but not the Paleozoic rocks, and may be of Middle Proterozoic age.

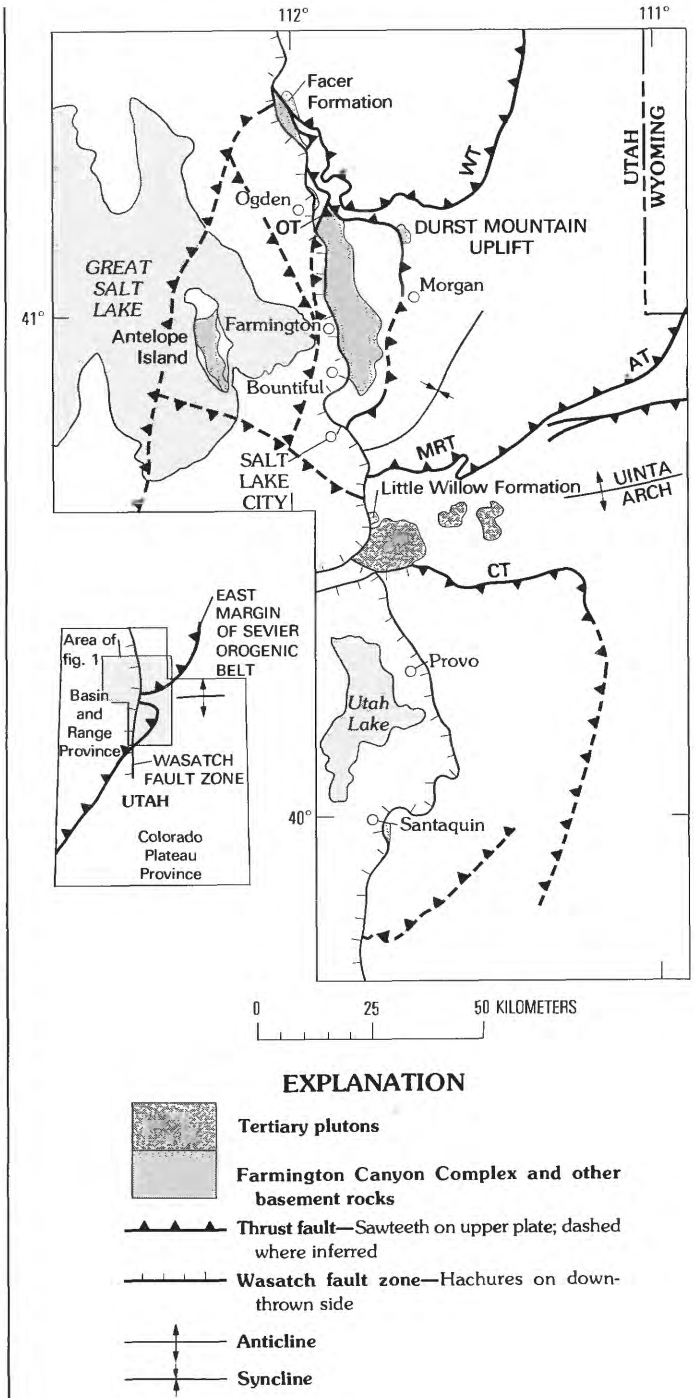


FIGURE 1.—Location and tectonic setting of the Farmington Canyon Complex. OT, Ogden thrust; WT, Willard thrust; MRT, Mt. Raymond thrust; AT, Absaroka thrust; CT, Charleston thrust. Major inferred faults not named. Modified from Stokes (1963) and Crittenden (1972, 1974).

INTRODUCTION

Crystalline basement rocks of the Farmington Canyon Complex (Eardley and Hatch, 1940) make up a 45-km-long segment of the Wasatch Mountains between Ogden and Bountiful, Utah (figs. 1 and 2), on the east margin of the Basin and Range province. The Farmington Canyon Complex also forms much of the lower west flank of the Wasatch Mountains from 25 km north of Ogden to Brigham City. Other areas of basement rock within the region have been called the Farmington Canyon Complex; these are on Antelope Island in the Great Salt Lake, 30 km west of the Wasatch Mountains; in the Durst Mountain uplift, 15 km east of the Wasatch Mountains; north of the mouth of Little Cottonwood Creek, 25 km to the south; and at Santaquin, some 100 km to the south of the type locality of the complex in Farmington Canyon and at Bountiful Peak. The outcrop area of basement rocks in the Wasatch Mountains is more than 300 km^2 and by far the largest area of basement rock exposed in north-central Utah.

HISTORY OF PREVIOUS INVESTIGATIONS

Captain Howard Stansbury (1853) first recognized crystalline rocks in the region during his exploration of the Great Salt Lake and visit to Antelope Island. The exposures of crystalline rocks in the Wasatch Mountains were described in the reports of the Fortieth Parallel Survey (King, 1878; Hague and Emmons, 1877; Zirkel, 1876). Crawford (1935) described some quartzites in these rocks east of Bountiful.

Eardley and Hatch (1940) gave a general outline of the rock types and structure of the crystalline rocks in

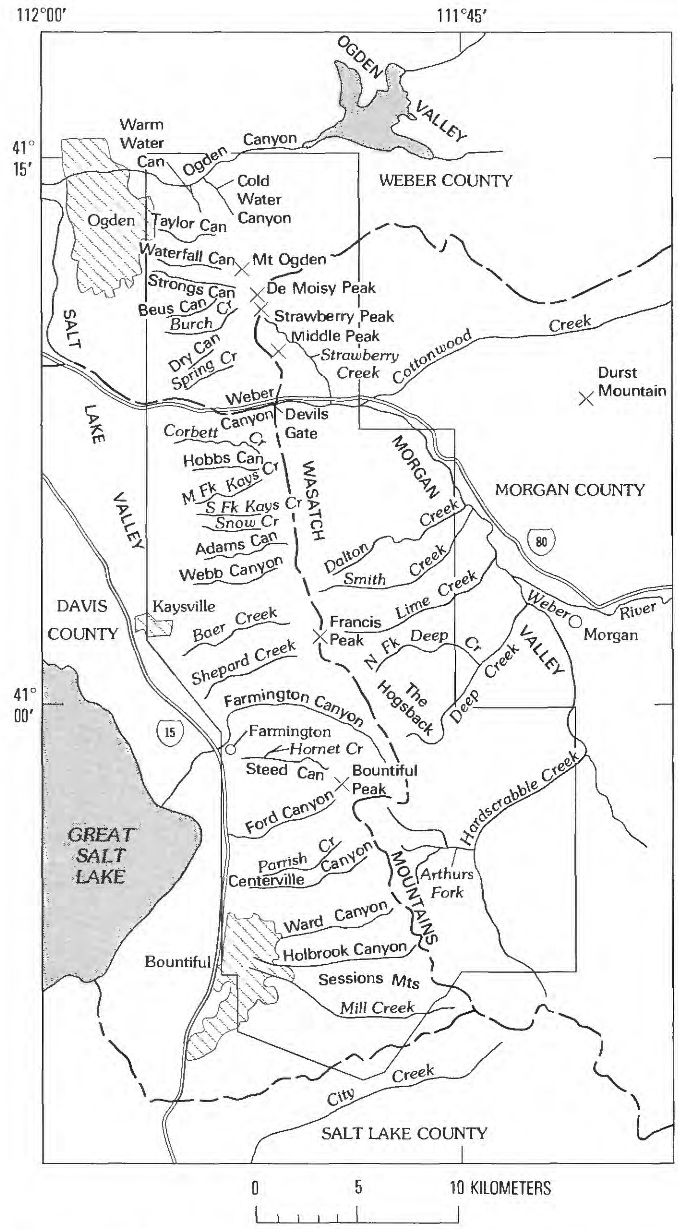


FIGURE 2.—Geography of Wasatch Mountains area underlain by the Farmington Canyon Complex. Area of plate 1 shown by outline.

north-central Utah and named them the Farmington Canyon Complex. They showed two units: (1) a small body of granite and granite gneiss west of Ogden Peak and (2) a mass of gneisses and schists to the south. Generalized foliation trends suggested the complexity of the structure. They concluded that the complex consisted of metamorphosed sedimentary rocks, siliceous igneous rocks, injection gneiss, and mafic rocks of uncertain origin. They ascribed the several degrees of metamorphism apparent in the complex to igneous intrusions during various stages of one prolonged, dominantly low to medium grade dynamic metamorphism.

The principal work on the Farmington Canyon Complex is a study by Bell (1951, 1952) of the central part of the complex between Weber and Farmington Canyons. He subdivided the Farmington Canyon Complex into units characterized by what he called metamorphic facies, which were in reality mineral assemblages. He recognized retrogressive metamorphism along many shear zones but also mapped areas of progressively metamorphosed rocks of the greenschist facies, although he noted that, in places, rocks of that facies were mixed with rocks of higher metamorphic grade in a confusing fashion. "Some zones, only a few feet in length, include nearly all the metamorphic facies and structural characteristics of the entire range" (Bell, 1951, p. 11). He also locally mapped areas of upper(?) Precambrian quartz schist of low metamorphic grade. He reported minerals of high metamorphic grade including sillimanite, cordierite, kyanite, and hypersthene and concluded that most of the metamorphic rocks were of sedimentary origin. He also recognized that a number of high-angle faults cut the Precambrian rocks and were younger than the Wasatch Formation (his Knight Formation).

Temple (1969) mapped the basement rocks from Ogden Canyon south to Burch Creek. Recently, geologists at the University of Utah have conducted more detailed topical studies of the Farmington Canyon Complex in the Wasatch Mountains (Hollett and others, 1978; Bruhn and Beck, 1981).

PRESENT WORK

This report presents the results of an overall reconnaissance study of the Farmington Canyon Complex. In connection with a study of the relations between the thrust belt in southwestern Wyoming and north-central Utah in the Salt Lake and Ogden 1° × 2° quadrangles, I made a reconnaissance geologic map of the Farmington Canyon Complex in the Wasatch Mountains (Bryant, 1984), included in this report as plate 1. I wanted to see if any major thrust faults could be located in the complex by the presence of intercalated slices of younger Precambrian rocks, as suggested by Bell's work (1951, 1952), or by relations between different blocks of rock with different metamorphic histories. Most of the fieldwork was done in the summer of 1977 with the assistance of Peter R. Moon, but included a few additional days in 1978 and 1979 to make a total of about 3 months. Many of the other areas of basement rock in the region were visited briefly. About 250 thin sections were studied in the laboratory. Preliminary results of the work were presented in brief reports by Bryant (1978, 1979, 1980), Bryant and Graf (1980), and

Bryant and others (1980). C. E. Hedge and J. S. Stacey undertook isotopic studies of the Farmington Canyon Complex rocks in conjunction with my work (Hedge and Stacey, 1980; Hedge and others, 1983). A summary of some of the data contained in the present report and further discussion of the Archean and Early Proterozoic regional geology was written subsequent to the completion of this report in 1984 (Bryant, 1988).

TOPOGRAPHY AND VEGETATION

The part of the Wasatch Mountains composed of Farmington Canyon Complex rocks is a narrow block 45 km long and 5–10 km wide and rises some 1.5 km above Salt Lake valley on the west and 1.2 km above Morgan valley on the east (fig. 2). Its northern end is cut by the canyon of the Ogden River, and the Weber River has cut a canyon through it 10 km to the south. Many small perennial streams drain the mountain mass and have cut canyons, which are especially rugged on the west side of the range. Vegetation ranges from open grassland, scrub oak, and mountain mahogany at the base of the range (1,400–2,000 m in altitude) through dense brush of oak, maple, chokecherry, and snowbush (2,000–2,600 m in altitude), to a zone of sagebrush at the crest of the range (2,600–2,900 m in altitude). Some sparse stands of Douglas fir are found on north- or east-facing slopes at the higher altitudes. The cover of dense brush makes field work slow and difficult in many areas in the middle altitudes. A number of small cirques scallop the east flank of the range where crest altitudes exceed 2,700 m.

Accessibility is only fair. Paved roads in Ogden and Weber Canyons, and a paved road to the Snow Basin ski area, serve the northern part of the area. In the southern part of the area, a gravel and paved road up Farmington Canyon loops around the crest of the range and down to Bountiful and has a spur north to Frances Peak. In the southeastern part of the area, a road extends up Hardscrabble Creek from south of Morgan. Several trails penetrate the range from the Salt Lake valley; most of them are not shown on the topographic quadrangles. The land is used for recreation, and, east of the crest of the mountains, for stock grazing in the summer months. Trails used by stockmen extend up many valleys on the east side of the range.

TECTONIC SETTING

The Farmington Canyon Complex is exposed in an elongate, structurally high area framed by overlying Paleozoic rocks containing thrust faults and disrupted by Tertiary extensional structures. The rocks of the complex are parautochthonous in relation to those of the Willard thrust sheet tectonically overlying them to the north (fig. 1; Crittenden, 1972). Below the Willard thrust sheet, Cambrian through Mississippian rocks form another sheet, of which the distance of travel is not known. The Ogden thrust at the base of this lower sheet follows mainly incompetent beds of the Ophir Formation of Middle Cambrian age, and south of Ogden Canyon it cuts downsection into the Farmington Canyon Complex on the west side of the range where it is offset by the Wasatch fault. Mapping (Crittenden and Sorensen, 1985a, 1985b; pl. 1) shows that the Ogden thrust is just one of a complex of thrusts, some branches of which pass above the Farmington Canyon Complex on the east side of the range and disappear beneath the Tertiary rocks of Ogden and Morgan valleys. The Ogden thrust fault zone may connect with the zone of intersliced Precambrian basement and Cambrian rocks mapped by Mullens and Laraway (1973) and R. J. Hite (written commun., 1977) on the west side of the Durst Mountain uplift, which is on trend with faults mapped by me in Hardscrabble Creek involving the Farmington Canyon Complex and rocks of Cambrian through Mississippian age. This zone of structural disturbance trends towards thrust faults mapped south of City Creek, which may be correlative with the Mt. Raymond and Absaroka thrusts (Crittenden, written commun., 1977). These faults were folded when the Farmington Canyon Complex was rumpled into a ramp anticline with a complex, nearly vertical east limb overlain by rocks of Maastrichtian age (Mullens and Laraway, 1964). This folding probably occurred during an early stage of movement on the Absaroka thrust (Royse and others, 1975), which would indicate that the Ogden thrust is older than the Absaroka thrust and that the thrusts outlining the Farmington Canyon Complex do not correlate with the Mt. Raymond or Absaroka thrusts.

Formation of the ramp anticline took place just west of the emergence of a regional decollement from crystalline basement rocks (Royse and others, 1975; Bruhn and others, 1983). Structural relations to the east in southwestern Wyoming indicate that movements on the inferred regional decollement must have continued to Paleocene time (Wiltschko and Dorr, 1983).

Although these thrusts of late Mesozoic age help outline the block of Farmington Canyon Complex in part, high-angle faults of late Tertiary and Quaternary age complete the framework by outlining the Wasatch Mountains. These include the Wasatch fault on the west side of the block and a complex system of faults that dies out southward on the east side.

ROCK UNITS OF THE FARMINGTON CANYON COMPLEX

The rocks of the Farmington Canyon Complex are highly deformed, metamorphosed, and migmatized. The complex has been divided into four principal map units, three of which contain a variety of rock types (pl. 1). Although the rocks, except for one unit of plutonic rock, are predominantly of sedimentary origin, no stratigraphic marker horizons or stratigraphic sequence criteria were found that would allow determination of the large-scale structures of the complex. Even though there are significant differences in the proportions of rock types between the map units, the units are not sharply delineated on the ground because contacts between them are gradational. Within the principal map units smaller units consist of one rock type and have well-defined contacts. Cataclasis and retrogressive metamorphism have overprinted the rocks in parts of the Farmington Canyon Complex. In the following descriptions, modifying mineralogic adjectives are given in reverse order of statistical abundance, the most abundant last and next to the rock name.

PRINCIPAL ROCK UNITS

QUARTZ MONZONITE GNEISS

The northern part of the Farmington Canyon Complex near Mt. Ogden, described by Temple (1969) as the granite of Mt. Ogden, consists of hornblende and biotite-hornblende-quartz monzonite gneiss. This apparently is the unit that makes up the Farmington Canyon along the west front of the Wasatch Mountains from Ogden Canyon north to Willard, which Crittenden and Sorensen (1985a, 1985b) described as gneissic quartz monzonite cut by amphibolite layers and folded quartz veins.

Between Weber and Ogden Canyons, quartz monzonite gneiss was mapped in three bodies separated by areas of migmatite. The southern two bodies (pl. 1) contain more outcrops of layered and somewhat migmatitic rocks than the larger, northern body. Contacts of the northern body are generally fairly sharp and some are faults, but those of the southern bodies are gradational over greater distances.

The quartz monzonite gneiss has a fairly uniform aspect, a grain size of 1-2 mm, and a well-developed foliation formed by aligned hornblende needles and mica flakes. It locally has a streaky and spotty distribution of mafic minerals and grades into rock in which faint layering is due to variation in grain size and distribution of mafic minerals. The streaky and migmatitic rocks contain amphibolite layers and lenses a few centimeters thick, and the hornblende in the quartz monzonite gneiss is in irregularly scattered clots. In the southern two mapped bodies of quartz monzonite gneiss, layering is better developed and is formed by leucocratic pegmatite intercalated with hornblende-quartz monzonite gneiss. In these southern bodies a few of the gneiss layers contain garnet.

Pegmatites are widespread in rocks of this unit, and they form stringers and lenses 1-10 cm thick with sharp to indistinct borders. This type of pegmatite is best developed in the southern bodies of the quartz monzonite gneiss, but less well developed pegmatitic layering is locally found as far north as Ogden Canyon. Crosscutting pegmatites, a few centimeters to a few meters thick, generally have sharp contacts, but, locally, some of the smaller ones have indistinct borders. Some of the larger pegmatites contain hornblende crystals as much as 8 cm long.

INCLUSIONS

Inclusions of amphibolite range from clots a few centimeters long to lenses as much as 50 m long. Two bodies of amphibolite several hundred meters long are shown on plate 1. The body mapped along the mountain front south of Burch Creek contains some pyroxene-amphibole ultramafic rock, but the relations between amphibolite and the ultramafic rock there are unknown. Almost all the inclusions are aligned parallel with the foliation. At 7,890 ft (2,405 m) altitude, in a gully at the head of Taylor Canyon, a dike of metadiabase about 10 m thick has a directionless fabric and cannot be traced many tens of meters. Even this rock has pegmatites cutting it, suggesting either that it was emplaced at least before the end of the metamorphism during which the quartz monzonite recrystallized or that it, too, may be an inclusion. A few amphibole-rich inclusions lack the well-developed nematoblastic texture of the amphibolites and retain a relict igneous texture. Metagabbro is an appropriate name for these rocks. A few pods of gabbro, which are best exposed on the west side of the range crest south of Mount Ogden, lack foliation, even at their margins, and lack pegmatite. Their margins are richer in hornblende and are finer grained than their interiors. These pods could have been emplaced later than the metamorphism of the quartz monzonite gneiss, but no field relations were found to demonstrate intrusion into the quartz monzonite gneiss.

A distinctive rock type within the quartz monzonite gneiss is light-greenish-gray to dark-greenish-gray, and medium-light-gray, medium- to coarse-grained chlorite-sericite-quartz rock. It occurs in lens-shaped bodies several meters to several hundred meters long, which are best seen in the drainages of Waterfall and Strongs Canyons. One long narrow body at the head of Burch Creek is 2 km long and crosses the crest of the range south of Strawberry Peak. Contact relations with the quartz monzonite gneiss range from abrupt to gradational. No bedding or obvious compositional layering is present in the chlorite-sericite-quartz bodies. Some bodies are sheared locally; in one place the rock is a breccia containing vugs partly filled with quartz and hematite. Concentrations of hematite and quartz occur in the most strongly sheared rocks. Within these bodies are zones of shearing in which the rock has been converted to a phyllonite or siliceous blastomylonite, which locally contains lenses and veins of quartz. Although originally mapped as inclusions (Bryant, 1979), further field and petrographic data (see p. 7–8) indicate that these bodies may be zones of altered quartz monzonite. South of Burch Creek near the mountain front, one inclusion 300 m long contrasts with the chlorite-sericite-quartz rocks and is a coarse-grained white quartzite.

PETROGRAPHY

The quartz monzonite gneiss has a granoblastic texture and a grain size that generally ranges from 1 to 2 mm (fig. 3). Locally, quartz and potassic feldspar reach 4–6 mm in diameter. Plagioclase ranges in composition in different samples from An 10 to An 30 and averages about An 20. Plagioclase in samples studied from Ogden Canyon, and from one locality 6 km to the north, is albite. Plagioclase in the quartz monzonite gneiss includes quartz, hornblende, and zircon. Locally, the plagioclase has been saussuritized and sericitized. Myrmekitic intergrowths of quartz occur in small amounts in plagioclase grains adjacent to potassic feldspar.

Potassic feldspar is microcline, which ranges from nonperthitic through locally perthitic to a well-developed string perthite. It includes plagioclase, quartz, hornblende, and apatite.

Hornblende is the principal mafic mineral in the quartz monzonite gneiss and typically makes up about 10 percent of the rock. Amounts range from 2 to 20 percent. It is anhedral to subhedral and generally well aligned. It apparently grew under synkinematic conditions with crystallization outlasting movement in many rocks. Z is generally dark green to dark brownish green and olive green; Y, dark green to brownish green and olive green; and X, light yellowish brown or light brown; $2V$ is small and is estimated to range from 10° to 40° with an average of about 20°. $Z \wedge C$ ranges from 5° to 14°. These characteristics indicate that the hornblende is ferrohastingsite. The hornblende is locally altered to biotite or chlorite.

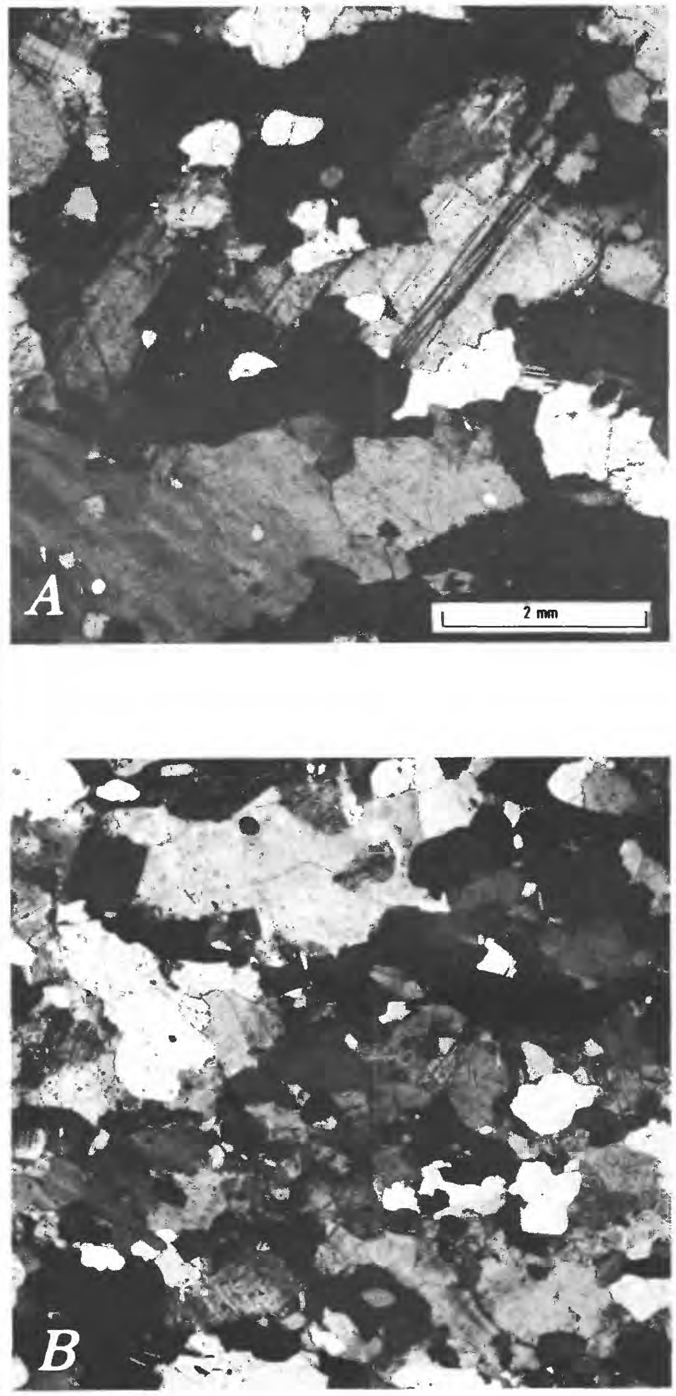


FIGURE 3.—Photomicrographs showing textures of quartz monzonite gneiss. *A*, Gneissic biotite-amphibole granite below Ogden thrust, from spoil at north end of Ogden-Brigham tunnel, sec. 3, T. 6 N., R. 1 W., north of area of fig. 2. Plagioclase is An 8; amphibole is hastingsitic. This texture is coarser than that of most quartz monzonite gneiss above Ogden thrust. Analyzed sample 1, table 1. Dated sample (Hedge and others, 1983). *B*, Biotite-bearing amphibole-quartz monzonite gneiss from 8,520 ft (2,595 m) altitude on ridge east of Strongs Peak. Typical texture of unit above Ogden thrust. Plagioclase is An 29 and somewhat sericitized; amphibole is hastingsitic. Analyzed sample 4, table 1.

Other amphiboles in the quartz monzonite gneiss include a fine-grained colorless one, probably derived by alteration of pyroxene, and light-green actinolite and a blue-green amphibole from hornblende.

Biotite makes up 0–8 percent of the quartz monzonite gneiss. Some of it is synkinematic and contemporaneous with the hornblende, but some formed later, in part through alteration of the hornblende.

Some of the quartz monzonite gneiss has light-brown splotches a millimeter or two in diameter, which, under the microscope, are seen to consist of a brown alteration mineral. In one sample, some grains of light-green monoclinic pyroxene as much as 2 mm in diameter are found.

Common accessory minerals are magnetite, apatite, zircon, and allanite. The magnetite occurs in elongate grains as much as 1 mm long. Apatite ranges from euhedral to round and is as much as 0.15 mm in diameter. Zircon is characteristically round (figs. 4*A* and *C*), but a few grains are euhedral and as much as 0.15 mm long. Many grains are 0.05 mm in diameter. Some have overgrowths and many have a light-purplish-brown color. Reddish-brown allanite forms euhedral crystals (fig. 4*A*) as much as 0.35 mm long. Much of it is between 0.1 and 0.2 mm. Some crystals are twinned (fig. 4*B*).

Sphene is an accessory mineral in samples from Ogden Canyon and 6 km to the north.

Many of the mafic minerals in some rocks have been altered to a mica tentatively identified as stilpnomelane.

Most of the amphibolite inclusions resemble amphibolite found throughout the Farmington Canyon Complex, and all the mafic rocks are described in the section on amphibolite.

A sample of ultramafic rock, from the area mapped as amphibolite along the mountain front south of Burch Creek, consists of light-green amphibole (58 percent) in grains as much as 2 mm long, part of which appears to be contemporaneous with pyroxene and part younger. Olivine (6 percent), as much as 2 mm in diameter, is fractured and sheared. Much of it is altered to serpentine, which constitutes about 22 percent of the rock. The olivine appears to have formed before the amphibole and pyroxene. Hypersthene (10 percent) forms grains as large as 4 mm, and it has numerous inclusions of amphibole. Magnetite makes up about 4 percent of the rock. A dark-gray-green, high-relief, isotropic, accessory mineral is tentatively identified as hercynite.

The quartzitic inclusions consist of granoblastic-textured quartz with a maximum grain size ranging from 2 to 4 mm. Some rocks show the effects of cataclasis, and the quartz is strongly strained and locally broken. Sericite and chlorite, in various proportions, make up as much as 50 percent of some rocks (fig. 5*B*).

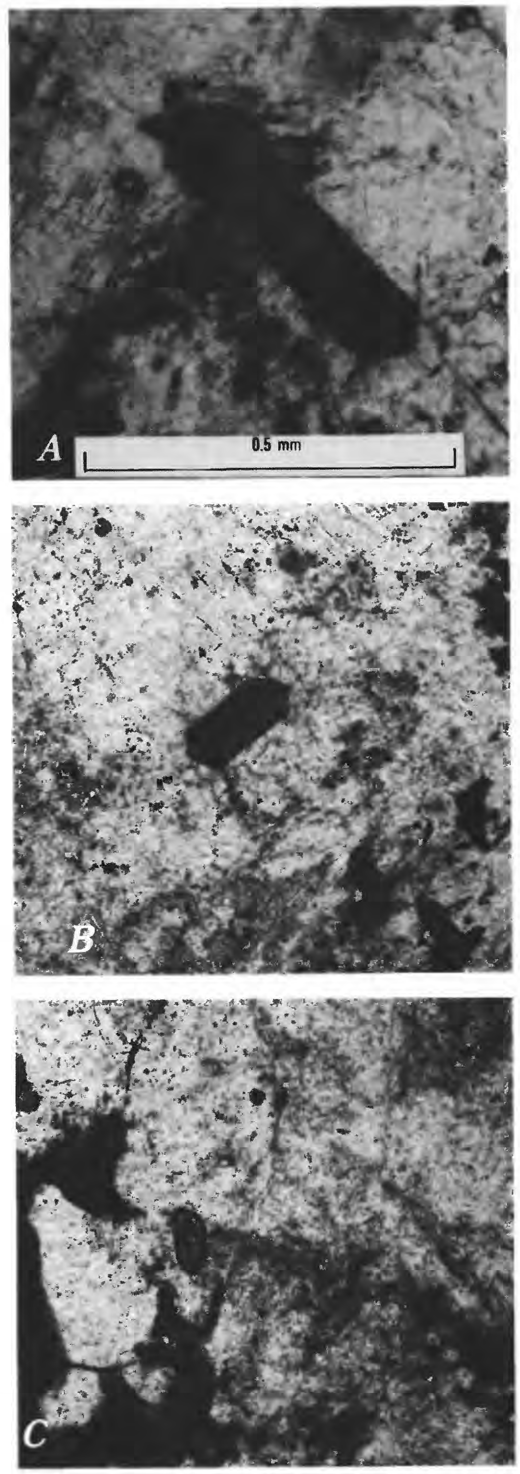


FIGURE 4.—Photomicrographs of allanite and zircon in quartz monzonite gneiss. *A*, euhedral allanite and round zircon; *B*, euhedral twinned allanite; *C*, zircon with rounded terminations. All from pyroxene-biotite-hornblende-quartz monzonite gneiss from base of slope above Lake Bonneville shoreline north of Beus Canyon. Analyzed sample 2, table 1.

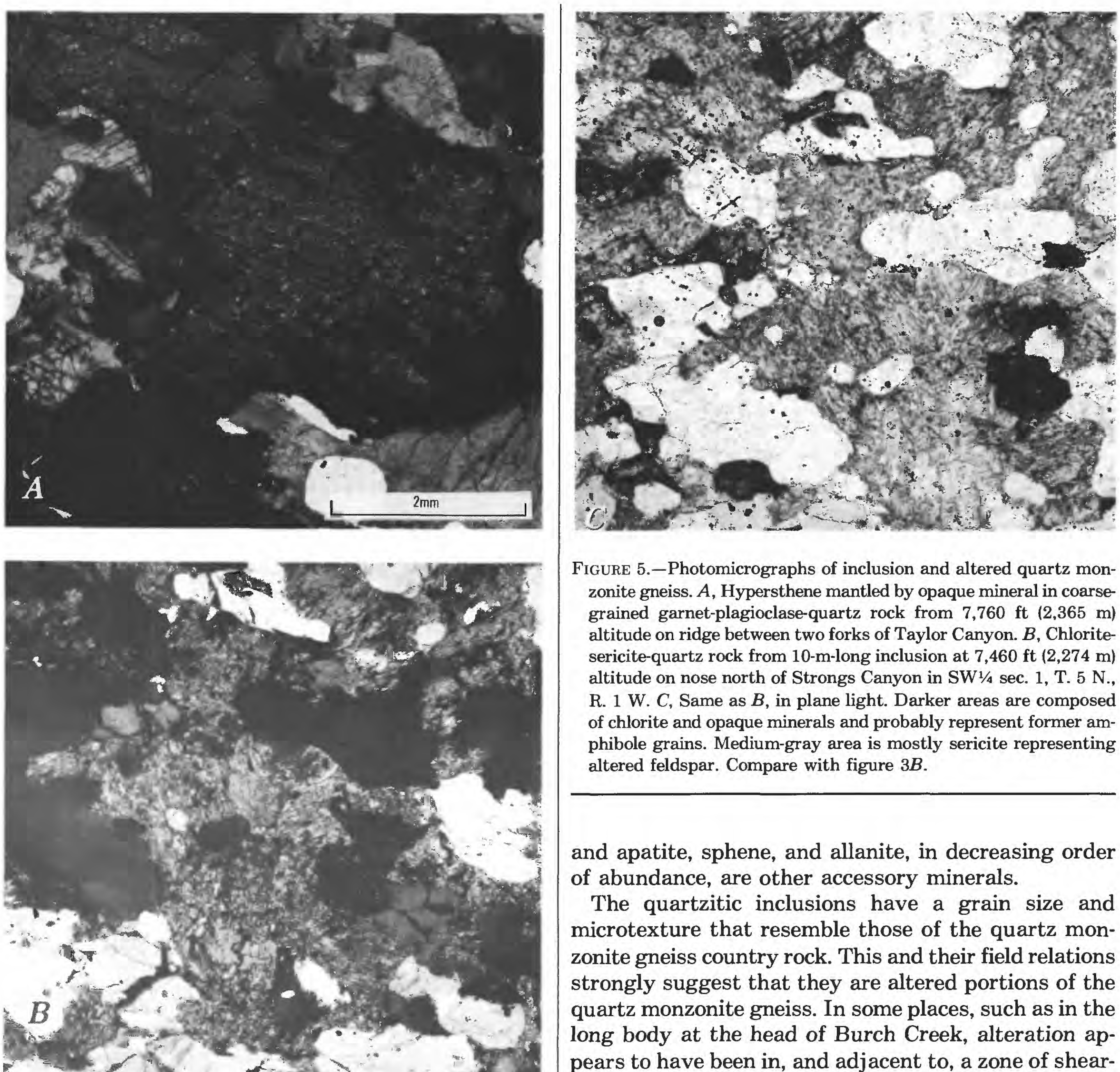


FIGURE 5.—Photomicrographs of inclusion and altered quartz monzonite gneiss. *A*, Hypersthene mantled by opaque mineral in coarse-grained garnet-plagioclase-quartz rock from 7,760 ft (2,365 m) altitude on ridge between two forks of Taylor Canyon. *B*, Chlorite-sericite-quartz rock from 10-m-long inclusion at 7,460 ft (2,274 m) altitude on nose north of Strongs Canyon in SW¼ sec. 1, T. 5 N., R. 1 W. *C*, Same as *B*, in plane light. Darker areas are composed of chlorite and opaque minerals and probably represent former amphibole grains. Medium-gray area is mostly sericite representing altered feldspar. Compare with figure 3*B*.

Instead of sericite and chlorite, some inclusions have a substantial (as much as 35 percent) proportion of epidote 0.1–0.3 mm in diameter either concentrated between the quartz grains or scattered throughout them. In the chlorite-bearing rocks, chlorite and titanium oxide occur together in a pattern that suggests that they are pseudomorphous after some earlier mafic mineral. Opaque minerals make up as much as a few percent of the rock and are hematite, magnetite, and ilmenite. Anhedral round to euhedral light-brown to light-reddish-brown zircon occurs in all the quartzites; and apatite, sphene, and allanite, in decreasing order of abundance, are other accessory minerals.

The quartzitic inclusions have a grain size and microtexture that resemble those of the quartz monzonite gneiss country rock. This and their field relations strongly suggest that they are altered portions of the quartz monzonite gneiss. In some places, such as in the long body at the head of Burch Creek, alteration appears to have been in, and adjacent to, a zone of shearing. The smaller bodies with sharp contacts seem more difficult to explain. Even the Burch Creek body seems to end abruptly to the northwest. If my conclusion is correct that these bodies are derived from the quartz-monzonite gneiss, a number of chemical changes must have occurred to explain the variation in content of Ca-, Na-, K-, and Mg-bearing minerals in the rocks.

The quartzite inclusion mapped on the mountain front, south of the mouth of Burch Creek, differs from the others in that it is almost 100 percent quartz with accessory muscovite, epidote, zircon, and apatite, and it resembles quartzite in the Farmington Canyon Complex east of Bountiful.

At 7,760 ft (2,365 m) altitude on the ridge north of Taylor Canyon, an inclusion of garnet-plagioclase-quartz rock contains anhedral quartz to 9 mm in diameter, anhedral to subhedral calcic oligoclase to 4 mm in diameter, euhedral to anhedral very light brown garnet to 2 mm in diameter, magnetite and accessory magnetite, apatite, zircon, and allanite. One anhedral grain of very weakly pleochroic hypersthene, 5 mm in diameter, is rimmed by magnetite (fig. 5*A*).

Another inclusion 2 m long and 0.5 m thick, from 7,460 ft (2,274 m) altitude on a cliffy nose on the north side of Strongs Canyon, is a quartz amphibole rock containing ferrohastingsite like that of the quartz monzonite gneiss. Quartz grains to 5 mm in diameter fill in around the amphibole. Magnetite, apatite, and zircon are accessory minerals.

COMPOSITION

Modes of petrographically studied samples of quartz monzonite gneiss fall in the center of a quartz-potassic-feldspar-plagioclase diagram (fig. 6). Some samples from Ogden Canyon and from a locality 6 km to the north are granite; both localities are from below the Ogden thrust. However, areas of Farmington Canyon Complex mapped in the North Ogden and Willard quadrangles have been described as gneissic quartz monzonite by Crittenden and Sorensen (1985a, 1985b). Color index is generally in the range of 10–20, and hornblende is the principal mafic mineral.

CHEMISTRY

Chemical analyses of four samples of quartz monzonite gneiss (table 1) show that in relation to averages of adamellite (Nockolds, 1954) the rocks have less Al_2O_3 and MgO and more Fe_2O_3 and FeO. Alkali and CaO contents are in the range of averages for adamellites. These chemical differences are reflected in the modal composition by the scarcity of mica and abundance of ferrohastingsite. Sample 1 from north of the Ogden River is a granite as indicated by the K_2O and total alkali content, but it also has a low Al_2O_3 and high Fe_2O_3 content in relation to the averages for granites.

Normative compositions plotted on a projection of isotherms on cotectic surfaces in the Qz-Ab-Or-An tetrahedron (fig. 7) show that the quartz monzonite gneiss plots near the cotectic lines in both diagrams and, thus, near the intersection of the cotectic surfaces, indicating that the rocks were derived from a relatively low temperature melt.

AGE AND ORIGIN

Rb-Sr isochron and Pb-U isotopic ratios from zircon

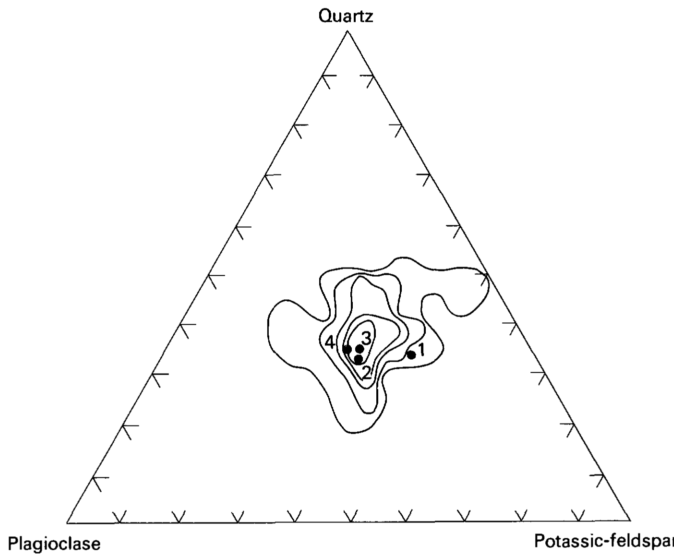


FIGURE 6.—Proportions of quartz, plagioclase, and potassium feldspar in samples, from the quartz monzonite gneiss that do not show severe effects of retrogressive metamorphism. Contours at 4, 9, 13, 17, and 22 percent. n = 23. Numbers of analyzed samples refer to table 1.

show that the quartz monzonite gneiss crystallized about 1,790 m.y. ago (Hedge and others, 1983). The very high initial $^{86}Sr:^{87}Sr$ ratio of 0.769 indicates that the quartz-monzonite gneiss originated in the upper crust, probably during a metamorphic event about 1,790 m.y. ago. Pb-U studies of feldspar from the gneiss show that much older crustal material participated in its genesis. The chemical composition of the rock is that of an early melting fraction (fig. 7), and much of the rock could have been derived by partial melting of migmatites like those exposed to the south of it. The numerous inclusions of amphibolite, quartzite, and other rock types may be interpreted as undigested relics of crustal material from the zone of melting and above. The well-developed foliation and granoblastic texture of the gneiss shows that it was emplaced in an early or middle stage of the metamorphic event.

Conditions of melting could reasonably have been 5 kb (fig. 7) in the Early Proterozoic, but a difference of a kilobar or two in pressure would not markedly change the relations shown. If $P_{H_2O} < P_{total}$, as it may have been, the temperature of melting would be higher than the 670° to 685°C shown in the diagram.

The origin of the quartz monzonite gneiss suggested here would put it in the class of the S-type granitoid of Chappell and White (1974). However, many features of it differ from those of S-type granitoids as compiled by Ferguson and others, 1980. Chemical criteria similar to those of S-type granitoids are $K_2O:Na_2O$ ratios, SiO_2 contents, and initial $^{87}Sr:^{86}Sr$ ratio. Chemical features

TABLE 1.—*Chemical analyses, modes, and norms of quartz-monzonite gneiss*

[Major oxides determined by X-ray spectroscopy by J. S. Wahlberg, J. T. Taggart, and James Baker, U.S. Geological Survey 1980 and 1981. FeO determined by chemical means by Edythe Engleman, U.S. Geological Survey 1981 and 1982, and Fe_2O_3 calculated by difference with Fe_2TO_3. Ten minor elements determined by quantitative emission spectroscopy by Carol Gent, U.S. Geological Survey, 1981, and Paul Briggs, U.S. Geological Survey, 1983. Other minor elements determined by ICP direct reading semiquantitative emission spectroscopy by George Riddle, U.S. Geological Survey, 1981. Looked for but not found Cd, U, Tb. Mode by count of 1,000 points. P, present but not intersected in counting; leaders (---), not present; N, not determined; <, less than. Major oxides and CIPW norm in weight percent; minor elements in parts per million; mode in volume percent. CIPW norms calculated on basis of analysis recalculated to 100 percent after deduction of volatiles]

Sample No.----	1	2	3	4
Lab No.-------	D-227762	D-227763	D-234324	D-234325
Field No.----	NO-1-B	O-41-C	O-22-A	O-38-A
Major oxides (percent)				
SiO_2	71.4	69.6	69.0	68.7
Al_2O_3	11.9	11.8	12.2	12.1
Fe_2O_3	2.14	2.33	3.28	2.52
FeO	3.20	4.63	3.53	3.90
MgO	0.3	0.4	0.43	0.45
CaO	1.34	2.27	2.21	2.08
Na_2O	2.4	2.6	2.80	3.05
K_2O	5.17	4.00	4.24	4.04
TiO_2	0.41	0.58	0.59	0.57
P_2O_5	<0.1	0.2	0.16	0.16
MnO	0.14	0.13	0.12	0.14
LOI	0.45	0.65	0.37	0.30
Quantitative minor elements (parts per million)				
Ba	697	1,180	1,330	1,180
Cr	16	45	<10	<10
Nb	31	74	40	50
Ni	4	14	<10	<10
Sr	42	67	70	70
Y	160	121	150	160
Zr	816	824	788	759
Ce	213	396	210	210
Nd	99	99	90	100
Rb	175	115	150	140
Semiquantitative minor elements (parts per million)				
Be	5.5	3	3	5
Bi	<30	54	<30	<30
Co	<1	10	2	4
Cu	15	30	8	9
Ga	24	29	<10	<10
La	120	3	74	130
Li	9	N	<5	6
Mo	6	3	<5	<5
Pb	88	18	100	60
Sc	8	13	13	18
Sn	<5	140	<10	<10
Th	36	5	15	30
U	<50	79	<50	<50
V	<3	10	6	9
Zn	160	8	130	180
Pr	30	16	<20	<20
Sm	20	<10	<20	20
Eu	<3	3	<5	<5
Gd	40	10	N	N
Dy	30	41	N	N
Ho	7	<4	<10	<10
Er	20	11	N	N
Yb	10	15	10	17
Ta	N	71	<10	10
Modes (percent)				
Quartz	29.9	28.3	29.2	30.0
Plagioclase	18.0	26.6	25.4	27.3
Potassic feldspar	38.7	30.4	29.7	27.7
Hornblende	9.1	8.0	6.0	12.2
Biotite	2.5	2.3	---	0.5
Pyroxene	---	1.7	0.5	---
Opaque	0.8	1.5	1.7	1.8
Actinolite	---	---	1.7	---
Epidote	---	---	0.1	P
Chlorite	---	---	---	P
Stilpnomelane	1.0	0.5	1.1	P
Brown alteration product	---	---	3.9	---
Apatite	P	0.2	0.1	0.2
Allanite	0.1	0.4	0.3	0.3
Sphene	0.3	---	---	---
Zircon	0.2	0.1	0.2	P

TABLE 1.—*Chemical analyses, modes, and norms of quartz monzonite gneiss*—Continued

Sample No.——	1	2	3	4
Lab No.——	D-227762	D-227763	D-234324	D-234325
Field No.——	NO-1-B	O-41-C	O-22-A	O-38-A
	CIPW Norms (percent)			
q	33.23	32.34	31.00	29.1
c	—	—	—	—
or	31.85	23.98	25.42	24.43
ab	20.64	22.32	24.04	26.41
an	6.53	8.84	8.32	7.57
di-wo	0.09	0.53	0.73	0.80
di-en	0.02	0.08	0.20	0.17
di-fs	0.08	0.49	0.56	0.68
hy-en	0.74	0.93	0.89	0.97
hy-fs	3.67	5.46	2.50	3.82
ol-fo	—	—	—	—
ol-fa	—	—	—	—
mt	3.15	3.43	4.82	3.74
hm	—	—	—	—
il	0.79	1.12	1.14	1.11
ap	—	0.48	0.38	0.39

LOCATION AND DESCRIPTION OF SAMPLES

1. Gneissic biotite hornblende granite from dumps at north end of tunnel on the Ogden-Brigham canal; sec. 3, T. 6 N., R. 1 W., North Ogden 7.5-minute quadrangle. Anhedral quartz to 2 mm in diameter, anhedral somewhat perthitic microcline as much as 4 mm long and anhedral plagioclase (An 8) as much as 2 mm in diameter. Anhedral ferrohastingsite as much as 4 mm long with a $2V$ of 10°-20° and dark-green absorption for Z. Biotite as much as 2 mm long. A small proportion of it replaces hornblende. A thin cataclastic zone cuts the rock. Zircon ranges from round to prismatic. Dated sample (Hedge and others, 1983).
2. Pyroxene biotite hornblende quartz monzonite gneiss from base of slope above the Bonneville shoreline north of Beus Canyon 280 m N. 17°E. from SE corner sec. 11, T. 5 N., R. 1 W., Ogden 7.5-minute quadrangle. Quartz is anhedral and 2 mm in diameter, but a few grains form masses as much as 6 mm long interstitial to the other minerals. Plagioclase in grains as much as 2 mm in diameter is locally sericitized and has a composition of about An 20. Pyroxene is light green, anhedral, as much as 2 mm in diameter and altered to an Fe oxide. Ferrohastingsite is anhedral as much as 3 mm long, has $Z \wedge C=13°$, $2V=20°$, and a dark-greenish-brown absorption for Z. Biotite is 2 mm in diameter. Zircon ranges from round to anhedral and has overgrowths.
3. Hornblende quartz monzonite gneiss from 7,460 ft (2,275 m) altitude on cliffy nose north of Strongs Canyon 360 m S. 44° W. of center sec. 1, T. 5 N., R. 1 W., Ogden 7.5-minute quadrangle. Quartz, plagioclase (about An 20), and finely perthitic microcline are anhedral and reach 2 mm in diameter. Synkinematic anhedral ferrohastingsite is as much as 1.5 m long, and has $Z \wedge C=14°$, $2V=10°-20°$, and Z dark olive green. Zircon is rounded to subhedral prismatic.
4. Biotite-bearing hornblende quartz monzonite gneiss from 8,250 ft (2,515 m) altitude on the ridge east of Strongs Peak, Ogden 7.5-minute quadrangle. Quartz as much as 1.8 m in diameter, somewhat sericitized plagioclase (An 29) as much as 1 mm in diameter and locally perthitic microcline as much as 1.5 mm in diameter are all anhedral. Synkinematic dark-green ferrohastingsite as much as 1.5 mm has $Z \wedge C=12°$, $2V \approx 10°$. Zircon forms rounded grains.

differing from S-type granitoids are mol $Al_2O_3/(Na_2O + K_2O + CaO)$ and $Fe^3/(Fe^2 + Fe^3)$ ratios and the absence of normative corundum. All the mineralogical criteria, except the local presence of garnet, disagree with those given for S-type granitoids. The reason for these discrepancies may be the nature of the source material for the quartz monzonite. It was derived from older continental crust that may have been desiccated by a granulite facies metamorphism rather than from first cycle eugeoclinal rocks.

MIGMATITE

South of the quartz monzonite gneiss as well as interlayered with it are heterogeneous gneisses and schists all mapped as one unit, designated migmatite. Bell's study (1951, 1952) covered a large part of the unit described as migmatite here. These rocks are well exposed in many places. The most accessible and spectacular exposures are in Weber Canyon, which cuts a 1-km-deep furrow across the range southeast of Ogden. Other large exposures are in glacial cirques on the east side of the range north of Francis Peak.

Contacts with the quartz monzonite gneiss range from sharp to gradational. Contacts with the schist and gneiss unit to the south are arbitrary as the number of layers of gneiss with granitic aspect gradually diminishes in that direction. I estimate that it is not possible to objectively determine the position of that contact closer than within about 1 km.

The migmatite unit contains a variety of rock types interlayered and intergrading with each other. Most exposures display compositional layering (fig. 8), but a few are composed of uniform nonlayered gneiss that generally has indistinct contacts with layered gneiss. In some rocks a lensy or patchy distribution of quartz and feldspar-rich garnet, biotite, and (or) hornblende-rich portions is well developed. In some nonlayered rocks the mafic minerals form irregular clots. Most of the rock in the unit is gneiss, and schist and amphibolite form a minor proportion of the unit. Some of the gneiss, such as exposed at Devils Gate on the Weber River, is hornblende, biotite-hornblende, and biotite-quartz monzonite gneiss or, rarely, granite gneiss that looks similar to rock of the quartz monzonite gneiss unit. Other gneisses are

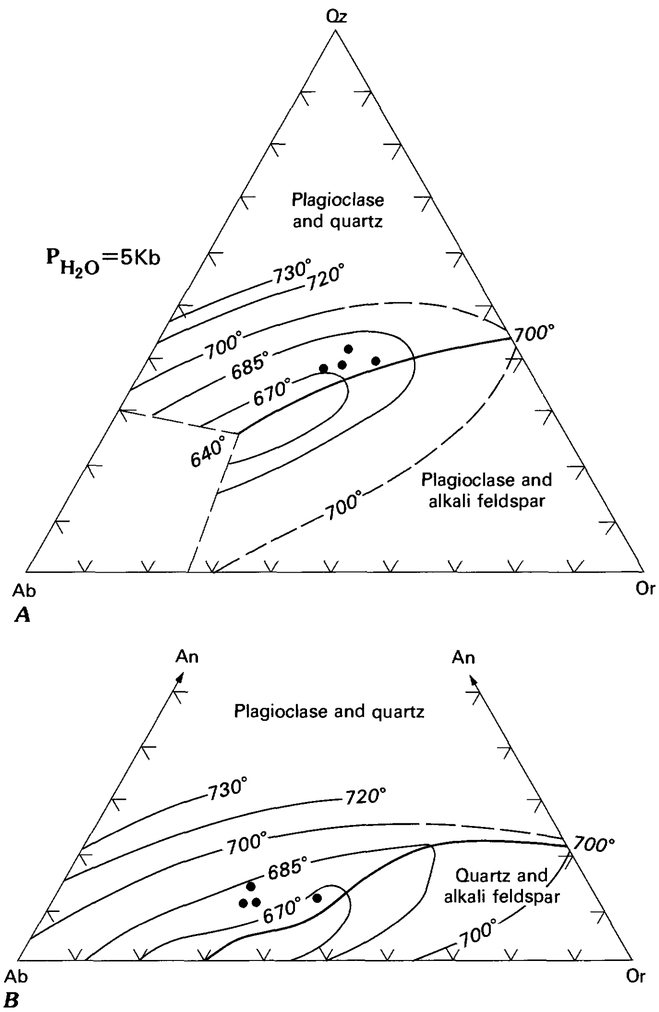


FIGURE 7.—Normative composition of quartz monzonite gneiss (dots) compared to experimental data for the system quartz (Qz)-albite (Ab)-orthoclase (Or)-anorthite (An) at 5 kb and $P_{H_2O} \approx P_{total}$ (after Winkler, 1979). Isotherms show configuration of liquidus surface: solid line shows cotectic composition between 700° and 640°C. *A*, Data projected on Qz-Ab-Or surface of the Qz-Ab-Or-An tetrahedron; *B*, Data projected on the Ab-Or-An surface.

garnet-hornblende-feldspar-quartz gneiss,
garnet-biotite-hornblende-feldspar-quartz gneiss,
garnet-feldspar-quartz gneiss,
garnet-biotite-feldspar-quartz gneiss, and
garnet-biotite-quartz-feldspar gneiss.

A distinctive rock type is white garnet-feldspar-quartz gneiss in which garnets range from 0.5 mm to 2 cm, but average 1–5 mm in diameter, and constitute as much as 30 percent of the rock. Fresh exposures of garnet-feldspar-quartz gneiss occur in cuts along the pipeline on the south side of Weber Canyon above the powerhouse.

Layers of schist, which are much less numerous than those of gneiss, consist of biotite schist, garnet-biotite schist, and sillimanite-garnet-biotite schist. Sillimanite-schist layers are most numerous in the migmatitic gneiss on the ridge south of Beus Canyon and in the Strawberry Creek drainage. Quartzite layers are much less common than the schist layers, and layers of calc-silicate rock are even less abundant. Amphibolite layers, lenses, and pods are scattered throughout the migmatitic gneiss. They range from a few centimeters to 8 m in thickness. Some larger pods are well exposed at Devils Gate on the Weber River.

Irregular variations in grain size and distribution and amount of felsic and mafic minerals give the rocks of the unit their migmatitic aspect. In addition, numerous pegmatites contribute to that aspect (fig. 8).

Pegmatite occurs in thin stringers and pods 1–10 cm thick, which commonly have indistinct contacts (fig. 8). Irregular crosscutting pegmatites are usually younger than the concordant ones and tend to have sharp contacts (figs. 8*A* and *B*). Many of the pegmatites are 1–3 m thick, but some are 80 m or more thick and 200–300 m long. Some pegmatites have a foliation defined by aggregates of biotite, probably derived from former books of biotite. In several outcrops the concordant pegmatites were folded before the emplacement of the discordant ones.

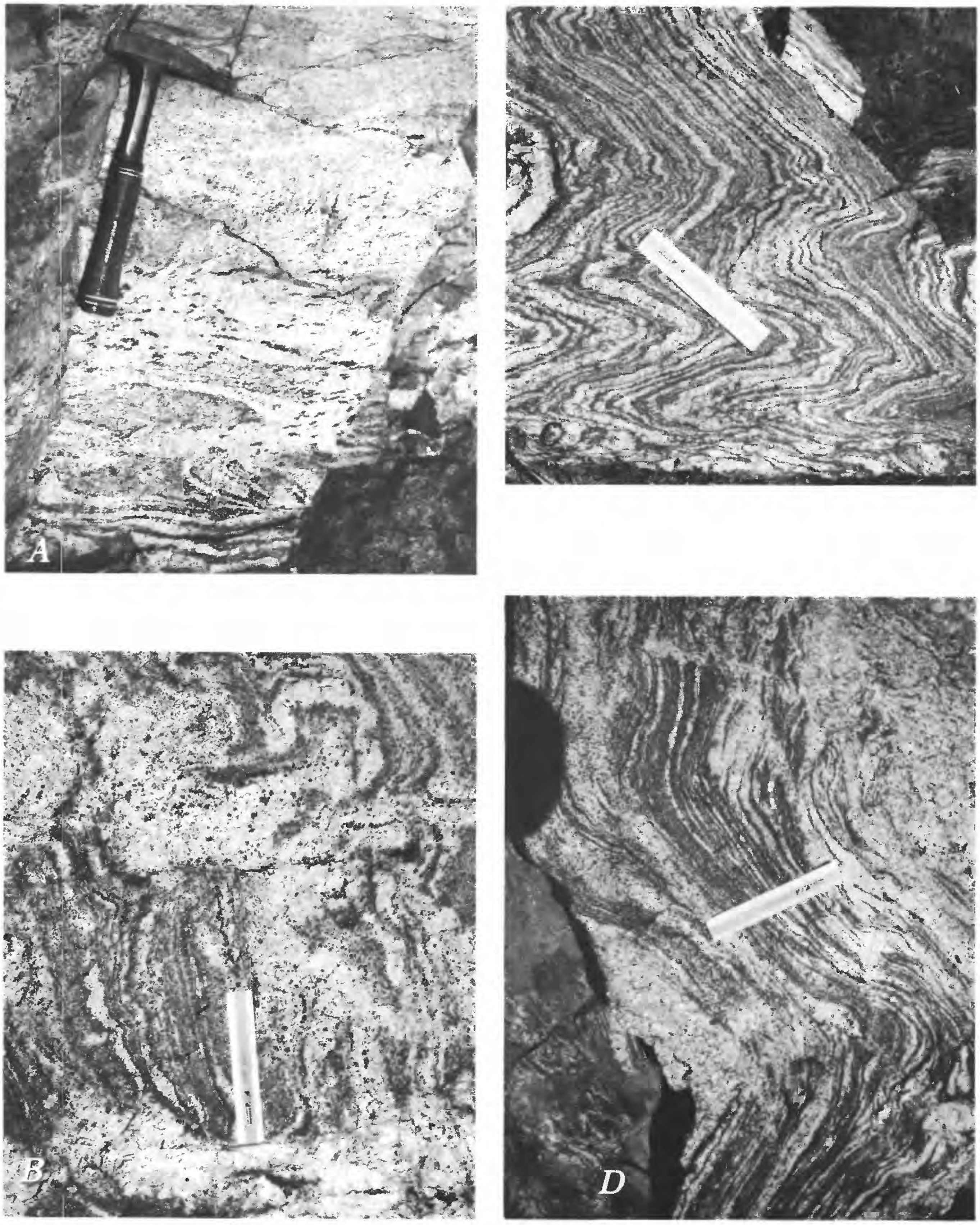
Severe cataclasis and retrogressive metamorphism have affected about 20 percent of the exposed area of the migmatite, gneiss, and schist unit, and will be discussed further in the section on cataclastic rocks.

PETROGRAPHY

The rocks of the migmatite unit show a wide range in texture, mineralogy, and modal composition compared to the quartz monzonite gneiss. Some layers, however, appear identical to the quartz monzonite gneiss to the north.

Almost all the gneisses contain anhedral quartz grains ranging from 1 to 4 mm in diameter. In some

FIGURE 8 (facing page).—Outcrops of migmatite. Scale is 18 cm long; hammer is 31 cm long. *A*, Fold in migmatitic biotite-hornblende-quartz monzonite gneiss on west end of outcrop on old U.S. 30, west of bridge across Weber River on I-80, just east of Devils Gate. Note patchy and irregular development of pegmatite. *B*, Migmatitic biotite-garnet-quartz-feldspar gneiss north of saddle, south of 9,100 ft (2,774 m) knob on crest of range, Snow Basin quadrangle. Concordant and crosscutting pegmatites grade to gneiss; larger pegmatite at bottom has a partly sharp contact. *C*, Folded migmatitic biotite and garnet-biotite-quartz-feldspar gneiss just south of VABM 9707 on crest of the range in the Peterson quadrangle. Intercalated thin concordant pegmatites. Folds of this type plunge southwest in this area. *D*, Migmatitic garnet-biotite-quartz-feldspar gneiss north of VABM 9707, Peterson quadrangle. Concordant pegmatites with indistinct margins. Nearby concordant and discordant pegmatites are as much as 3 m thick.


A
B
D

rocks the larger quartz grains engulf the other minerals and are apparently porphyroblasts formed late in the crystallization history of the rocks. Plagioclase occurs in anhedral grains 1-3 mm in diameter. Composition ranges from An 20 to An 35, and in many rocks the plagioclase is about An 25. Sparse myrmekitic intergrowths occur in plagioclase adjacent to microcline grains in many of the samples studied. Microcline is generally anhedral and ranges from 1 to 3 mm in grain size and from a fine-textured string perthite to entirely nonperthitic in character. The relative amounts of quartz, plagioclase, and microcline differ widely among the various layers in the migmatite unit.

Garnet is characteristic of many of the migmatitic gneisses and, locally, makes up as much as 30 percent of the rock. About half of the gneisses studied contain some garnet. In some rocks it is the sole mafic mineral aside from the accessory minerals; in others it is associated with various quantities of biotite and (or) hornblende. The garnet ranges in grain size from 0.5 to 5 mm and is usually anhedral to subhedral, although in a few rocks it is euhedral. Commonly garnet has a light-grayish-orange tinge and is poikiloblastic with included quartz, plagioclase, biotite, and opaque minerals. In some rocks it is altered along fractures to chlorite and epidote, or biotite and sericite.

Hornblende forms anhedral to subhedral grains 1-3 mm long. Overall, it formed both synkinematically and postkinematically with a slight dominance of crystallization during movement. $Z \wedge C$ of the hornblende in many rocks is about 10°, and $2V$ is 10°-30°, suggesting that it is ferrohastingsite like that in the quartz monzonite gneiss unit. However, the hornblende in some rocks has high values for $Z \wedge C$ and $2V$ and is probably a more Mg-rich variety.

Biotite ranges from 0.3 to 5 mm long, but is generally 1-2 mm long. Like the hornblende, it formed both synkinematically and postkinematically, but more of it appears to be strictly synkinematic than the hornblende.

A sample of one layer from an outcrop on the north side of Burch Creek contains some relict hypersthene in the cores of aggregates of fine-grained, light-yellowish-brown amphibole.

Magnetite is a common accessory mineral and makes up as much as 4 percent of some rocks. In places it occurs in elongate grains as much as 5 mm long.

Apatite, zircon, and allanite occur in most of the gneisses. Apatite forms anhedral to euhedral grains averaging 0.1-0.2 mm long. Zircon generally occurs as round grains 0.05-0.20 mm in diameter, but a few grains are subhedral to euhedral. Some crystals have overgrowths and may have faint light-reddish, purplish, purplish-brown, or very light brown hues. Reddish-brown, locally twinned allanite forms anhedral to euhedral crystals 0.1-0.2 mm long, and few are as much as 0.7 mm long.

Secondary minerals in the gneisses are chlorite from garnet, hornblende, and biotite; epidote from those minerals and plagioclase; sericite from plagioclase; stilpnomelane from mafic minerals; and actinolite from hornblende.

Schist layers contain larger quantities of biotite and less quartz and feldspar than the gneiss layers. Plagioclase ranges from An 18 to 27, and microcline in some rocks is perthitic, just as in the gneisses. The schists commonly contain anhedral to subhedral garnet as much as 5 mm in diameter with many inclusions of quartz, plagioclase, and opaque minerals. The most distinctive mineral in the schists is sillimanite, which occurs as well-aligned needles as much as 4 mm long. In rocks containing both sillimanite and microcline, these two minerals appear to have formed contemporaneously. Sericite and muscovite formed later from the sillimanite. One sample from northwest of Middle Peak contains about 10 percent cordierite in grains as much as 6 mm long, many of which have been altered to sericite and chlorite. A few needles of sillimanite are enclosed by the cordierite. Accessory minerals in the schists are apatite and zircon.

Bell (1951, p. 28) reported kyanite in schist from somewhere in the migmatite unit. Apparently it was associated with muscovite, but he never mentioned the age relations between the muscovite and kyanite in that rock. Sillimanite is much more widespread in this terrane than either kyanite or cordierite.

No layers of calc-silicate rock or marble were identified in the field in the migmatite and gneiss unit, but two thin sections were found to be of tremolite-rich rock. One consists of tremolite (80 percent), sericitized feldspar (16 percent), and magnetite (4 percent). The other is a biotite-garnet-epidote-tremolite schist containing a small amount of calcic andesine to labradorite.

COMPOSITION AND ORIGIN

Figure 9 summarizes modal compositions of the felsic component in samples of quartz-feldspar gneisses in the migmatite unit. Not included here are schists, amphibolites, or rocks in which significant quantities of feldspar have been converted to mica during retrogressive metamorphism. Color index of these rocks ranges from 4 to 40, and 80 percent of the rocks have a color index of less than 20. Garnet ranges from 0 to 18 percent, hornblende from 0 to 34 percent, and biotite from 0 to 34 percent. In some rocks, all these mafic minerals occur together; in others, just one or two occur. Modal compositions of the more biotite-rich gneisses overlap those of schist layers. The difference

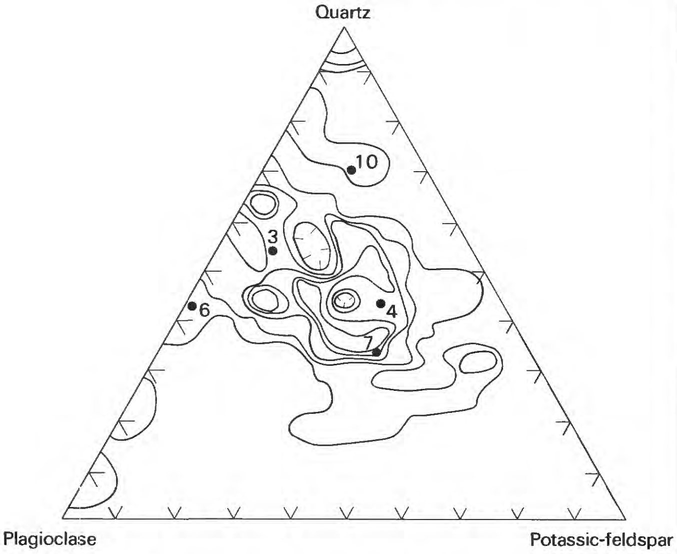


FIGURE 9.—Proportions of quartz, plagioclase, and potassium feldspar in samples from migmatite that do not show severe effects of retrogressive metamorphism. Schist samples not included. Contours at 1.5, 3, 4.5, 6, and 7.5 percent. n = 66. Numbers of analyzed samples refer to table 2.

is in texture; quartz and feldspar have a larger grain size in the mica-rich gneisses than in the schists of similar composition.

Many layers in the migmatite are quartz monzonite gneiss like that in the unit to the north, as shown by the larger area bounded by the 7.5-percent contour in figure 9. However, unlike the rocks of the quartz monzonite unit, some of these layers contain significant percentages of garnet. Other quartz monzonite gneiss layers are identical in mafic content to rocks of the quartz monzonite gneiss unit. More than half of the felsic minerals in many other layers are quartz. These layers could be derived from tuffaceous sediments, graywackes, or arkoses. The sillimanite schist, garnet-biotite schist, biotite schist, and some of the gneisses with high biotite content were derived from rocks rich in alumina and represent beds of shale in the depositional sequence.

SCHIST AND GNEISS

Rocks of the Farmington Canyon Complex gradually lose their migmatitic aspect southward. A contact between migmatitic rocks to the north and nonmigmatitic rocks to the south is arbitrarily drawn north of Farmington Canyon. The schist and gneiss unit extends south beyond Mill Creek to where it is covered by Paleozoic sedimentary rocks (pl. 1). The type locality of the Farmington Canyon Complex (Eardley and Hatch, 1940) is in this unit in Farmington Canyon and on Bountiful Peak, which have the best outcrops of the schist and gneiss unit.

This unit typically contains thinly layered biotite-feldspar-quartz gneiss and biotite schist (fig. 10*A*). Some outcrops have layers of schist as thin as 0.5 cm between layers of gneiss 5 cm thick. Some of the gneiss, especially in the northern part of the unit, has incipient pegmatitic segregation layering and looks migmatitic. Layers of feldspathic quartzite and quartzite are numerous, especially near contacts with the schist, gneiss, and quartzite units. Some of the quartzite layers are lens-shaped with a thickness of a meter or two and a length of 10 or 20 m. The thickest quartzite layer observed was about 6 m. Garnet is common in schist and gneiss, and rare in quartzite.

Many of the rocks in the southern part of the area, especially in the Mill Creek and Hardscrabble Creek drainages, are colored greenish gray by chlorite scattered throughout the rock. The chlorite was previously misidentified in the field as a calc-silicate mineral (Bryant, 1978). Petrographic study shows that the mafic minerals, mainly biotite, and lesser amounts of hornblende and garnet, are altered to chlorite. Unlike most rocks in other areas where the mafic minerals have been extensively chloritized, much of the rock was not cataclastically deformed during the alteration.

Layers of sillimanite-biotite schist and sillimanite-muscovite-biotite schist are much more numerous in the schist and gneiss unit than in the migmatitic gneiss unit to the north.

The only calc-silicate rocks found in the unit were from a dump of a prospect in the Middle Fork of Arthurs Fork at about 6,900 ft (2,100 m) altitude, and on Hardscrabble Creek. Green calc-silicate gneiss, granofels, and white marble occur at Arthurs Fork and calc-silicate quartzite and granofels on Hardscrabble Creek.

Pegmatites are ubiquitous in the rocks of this unit, and they range from stringers a centimeter or two thick to lenses 10–20 m thick and a few hundred meters long (figs. 10*B* and *C*; see also pl. 1). The larger pegmatites tend to form prominent outcrops relative to the more easily weathered schist and gneiss country rock (fig. 10*C*). Concordant stringers and lenses are cut by generally thicker discordant dikes. Many of the pegmatites have sharp contacts (fig. 10*B*), but some of the smaller concordant ones have indistinct margins. Grain size of quartz and feldspar is as much as 15 cm, and some pegmatites have deformed biotite books several centimeters thick. Some pegmatites are foliated. In the southern part of the area, some pegmatites contain muscovite books as much as 2 cm in diameter.

As in the north, amphibolite bodies range from small

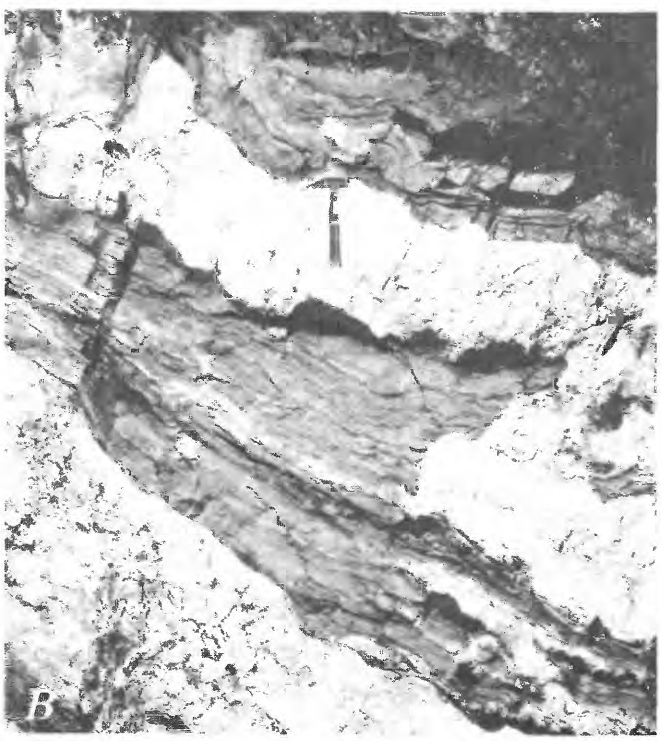


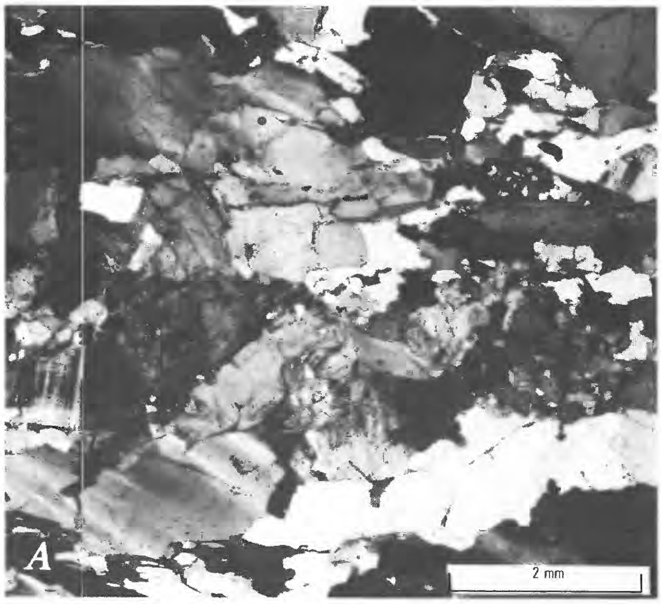
FIGURE 10.—Outcrops of schist and gneiss. Hammer is 31 cm long. *A*, Contorted, thinly layered biotite-quartz-feldspar gneiss and feldspar-quartz gneiss with thin partings of biotite schist. Typical west-plunging minor folds; plane of outcrop makes an angle of about 45° to their axes. At 9,000 ft (2,743 m) altitude northeast of Bountiful Peak. *B*, Sharply bounded, crosscutting pegmatites in biotite-quartz-feldspar gneiss. At 8,800 ft (2,682 m) altitude at base of cliff on northeast side of Bountiful Peak. *C*, View down Fords Canyon from 6,280 ft (1,915 m) altitude on ridge to south. Numerous pegmatite stringers, lenses, and pods as much as 10 m thick in gneiss.

pods less than a meter thick to lenses 15 m thick and 500 m long. Most of the amphibolites are typically black hornblende-rich rocks, but a few have aggregates of plagioclase that suggest the former presence of plagioclase phenocrysts.

PETROGRAPHY

Many of the petrographic characteristics of the schist and gneiss unit are similar to those of the migmatite unit. The rocks have a grain size of 1–2 m on the average and a well-developed granoblastic texture (fig. 11).

In the gneisses, quartz is 0.3–6 mm in diameter (fig. 11*A*). Some of the larger grains engulf the other minerals and are porphyroblasts. Plagioclase is 0.2–2 mm in diameter and ranges from calcic oligoclase to andesine, but is generally calcic oligoclase or sodic andesine. Myrmekite is sparingly present in a few rocks. Potassic feldspar is microcline, and it ranges from 0.5 to 3 mm in diameter. It is locally perthitic. Biotite occurs in flakes 0.5–2 mm long (fig. 11). Most of it formed synkinematically, but in some rocks it crystallized postkinematically. Light-pinkish-gray garnet forms anhedral to subhedral crystals 0.3–4 mm in diameter (fig. 11*C*). Many grains have sieve texture with biotite, quartz, and plagioclase. Locally, the garnet is altered to sericite and chlorite along fractures and in a few



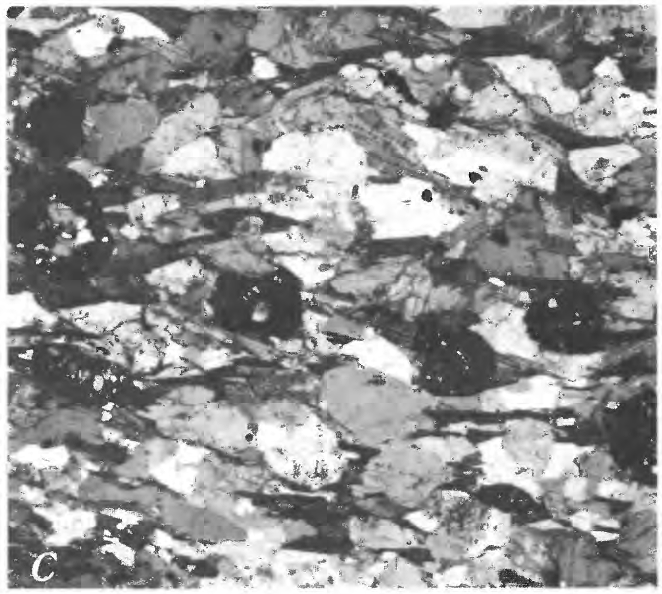


FIGURE 11.—Photomicrographs showing textures of gneiss. *A*, Biotite-microcline-plagioclase gneiss from 8,920 ft (2,719 m) altitude southeast of Bountiful Peak. Strained quartz and synkinematic biotite. *B*, Garnet-bearing quartz-biotite-plagioclase gneiss from roadcut west of BM 7946 in sec. 11, T. 2 N., R. 1 E. Plagioclase An 30, synkinematic biotite. Analyzed sample 1, table 2. Dated sample (Hedge and others, 1983). *C*, Sillimanite-bearing garnet-quartz-biotite-plagioclase gneiss from Mill Creek northwest of Big Rock. Synkinematic biotite, garnet with sieve texture. Plagioclase An 32 locally saussuritized. Analyzed sample 14, table 2.

places to biotite. Anhedral to subhedral hornblende as much as 3 mm long occurs in a few of the gneisses. Its composition covers a range, as suggested by variations in $2V$, $Z \wedge C$, and absorption colors. Muscovite occurs in small amounts in a few of the gneisses. It is intergrown with biotite in crystals 1 mm or less in length.

Zircon, the most common accessory mineral, forms in mostly round but, locally, subhedral to euhedral grains 0.03–0.25 mm long. Some grains are purplish gray or light reddish brown, and a few have overgrowths. An opaque mineral, identified as magnetite in many samples, occurs in grains 0.3–0.7 mm in diameter. Apatite is in anhedral to euhedral grains as much as 0.25 mm long. Allanite, which occurs in anhedral to euhedral grains as much as 0.25 mm long, is less widespread than in the migmatitic gneiss unit. A few rocks contain accessory sphene.

Secondary minerals are sericite, chlorite, and epidote derived from plagioclase and mafic minerals. A small amount of late-formed stilpnomelane occurs in a few rocks.

The schists contain more biotite and less microcline than the gneisses. They have plagioclase of similar composition.

In the schists, biotite as much as 4 mm long is generally synkinematic, although minor amounts of biotite formed postkinematically. Muscovite makes up as much as 50 percent of some of the schists, but it is usually present as only a few percent of the rock where it is intergrown with biotite.

The widespread and distinctive mineral of the schists is sillimanite (fig. 12), which makes up as much as 20 percent of the rock. It occurs in needles as much as 3 mm long, which may form aggregates as much as 1 cm long; it also occurs in fibrolitic aggregates. The sillimanite formed synkinematically and is commonly

FIGURE 12.—Photomicrograph showing intergrowth of sillimanite and microcline. Plagioclase-microcline-sillimanite-biotite schist from 5,620 ft (1,713 m) altitude on west side of Hornet Creek. A trace of muscovite (not shown) is intergrown with biotite in this rock.

intergrown with biotite. In many of the schists in the gneiss and schist unit, sillimanite appears to have formed contemporaneously with muscovite and microcline, but south of Holbrook Canyon, microcline was not found in the sillimanite mica schist.

Relicts of staurolite grains occur in sericitic aggregate in a retrogressively metamorphosed schist from the southeasternmost exposure of Farmington Canyon Complex on Hardscrabble Creek.

Zircon, apatite, and opaque minerals are common accessory minerals in the schists. Allanite and sphene occur locally in them.

The calc-silicate rocks, found at the prospect on the Middle Fork of Arthurs Fork, consist of chlorite-bearing marble, chlorite-quartz-epidote granofels, and calc-silicate granofels. The calc-silicate granofels is composed primarily of diopside with accessory epidote. Tremolite found in the rock appears to have formed later than the diopside. The only other calc-silicate rock found is a sheared epidote-quartz granofels from Hardscrabble Creek area.

Quartzite layers in the schist and gneiss unit consist predominantly of a mass of granoblastic-textured quartz with a grain size that reaches 2 cm. Accessory minerals are synkinematic biotite and muscovite as much as 2 mm long; anhedral to subhedral garnet to 1 mm long; and lesser amounts of zircon, apatite, rutile (?) or sphene, and opaque minerals. One sample of garnet-bearing quartzite, from southwest of Bountiful Peak, contains relict cordierite in patches of sericite and chlorite.

COMPOSITION AND ORIGIN

The gneisses of the schist and gneiss unit that lack severe effects of retrogressive metamorphism have a different distribution pattern on a quartz-potassic-feldspar-plagioclase diagram (fig. 13) than similar gneisses from the migmatite unit (fig. 9). Many of the gneisses from the schist and gneiss unit lack, or are very poor in, microcline. This is especially true of the gneisses studied from the Mill Creek and Hardscrabble Creek drainages in the southern part of the area.

Color index of the gneisses ranges from 0 to 50 percent with average values of 10–20. Biotite is the principal mafic mineral and makes up as much as 35 percent of some rocks. Garnet occurs in about half the rocks studied, but constitutes only a few percent of most of the rocks. Hornblende occurs in about 20 percent of the gneisses examined and in quantities of 20 percent or more in about 10 percent of the gneiss. In the hornblende gneiss, microcline is generally lacking or occurs in very small amounts, and quartz makes up from 50 to 100 percent of the felsic minerals.

The rocks containing more than 50 percent quartz could have been derived from arkosic sandstones or

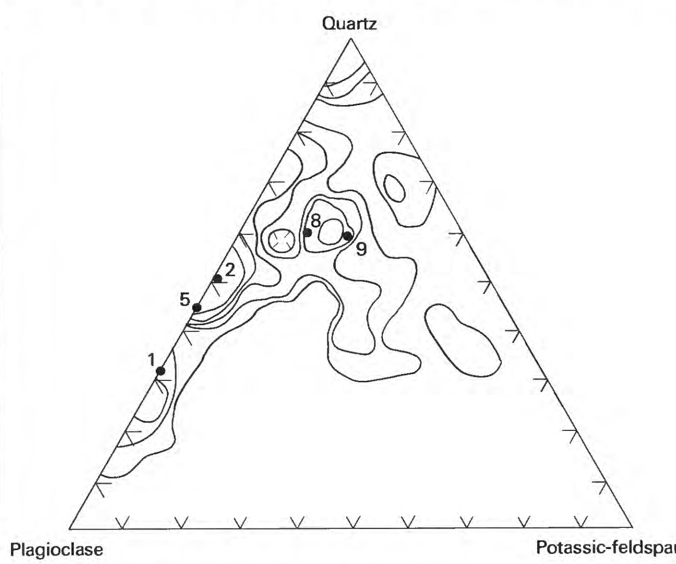


FIGURE 13.—Proportion of quartz, plagioclase, and potassic feldspar in samples of gneiss from the schist and gneiss that do not show effects of severe retrogressive metamorphism. Contours at 2, 4,5, 9, and 18 percent. n = 55. Numbers of analyzed samples refer to table 2.

graywackes; those with more than 50 percent feldspar might have some volcanic component in them. The rather small scale compositional layering in most outcrops argues against any lava flows as protoliths of these gneisses.

In summary, it appears that arkose, graywacke, and possibly tuffaceous sandstone, siltstone, or tuff may have been protoliths of the gneisses. The gneisses in Mill Creek must have been derived from sediments lacking clastic potassic feldspar in many beds, but containing abundant clastic plagioclase.

In the schists, biotite content ranges up to 40 percent. Garnet occurs in about half the samples studied in quantities ranging up to 20 percent. Muscovite occurs in about half the schists, but none occurs in samples from the northern part of the schist and gneiss unit. The proportion of quartz and plagioclase ranges from all quartz to all plagioclase. Microcline occurs in about a quarter of the schists.

Sillimanite is found in about half the schist samples studied in amounts as much as 20 percent of the rock. Its presence is not related to quartz:feldspar ratios.

The schists represent shales and geochemically immature shales with varying proportions of alkalis. Many have substantial contents of Na_2O as indicated by their plagioclase content.

QUARTZITE, GNEISS, AND SCHIST

Quartzite layers are thicker and much more numerous from the upper part of Ward Canyon to south of the crest of the Sessions Mountains where they form a moderately well defined map unit (pl. 1). This map unit contains the same rock types as the schist and gneiss unit, except that the quartzite layers are thicker and more numerous. Steeply dipping layers of quartzite several meters thick in outcrops at the head of Ward and Holbrook Canyons are obvious from a distance (fig. 14*A*). The most accessible and best exposures of this unit are in pipeline cuts on the upper part of the ridge separating Ward and Holbrook Canyons. There, amphibolite layers 0.3 to 10 m thick are intercalated with the quartzite. Pegmatites range from concordant bodies as much as several meters or more thick to irregular discordant ones as much as 0.4 m thick in the amphibolite layers. Two pegmatites 100 m long are shown in the quartzite unit on the reconnaissance geologic map (pl. 1).

The main mapped area of this unit has quite sharp contacts on its east and south borders, but its west contact is more gradational, and it contains more mica schist and biotite-feldspar-quartz gneiss in its western part. The contacts of the unit were mapped where quartzite is a consistent and obvious component in

FIGURE 14.—Outcrops of quartzite. *A*, View north across upper part of Holbrook Canyon showing steeply dipping quartzite beds. At upper right are some gently dipping beds. Pipeline along crest of the ridge. *B*, Fold in quartzite outlined by slightly more micaceous and feldspathic bed. Perpendicular to west-plunging fold axis. Hammer is 31 cm long.

outcrops or colluvium. Foliation, layering, and the plunges of the minor folds suggest that this area of quartzite occurs in a doubly plunging antiform.

A smaller area of this unit was mapped on the ridge between Centerville Canyon and Parrish Creek. The rocks there have much more schist and gneiss interlayered with them than in the Ward Canyon-Holbrook Canyon area. At Centerville Canyon, the unit appears to cap the ridge and dip west down the ridge toward the Salt Lake valley, although, in detail, the structure is more complex.

The schist and gneiss in the area from Ford Canyon south to Ward Canyon contain more layers of quartzite than elsewhere in the Farmington Canyon Complex outside the mapped areas of quartzite. Many of the layers are of feldspathic quartzite and, thus, are gradational between the quartzite and the typical biotite-feldspar-quartz gneiss of the schist and gneiss unit.

The quartzite contains no relict sedimentary textures. The only sedimentary structures observed are compositional layering (fig. 13*B*), which probably represents bedding, and a few heavy mineral partings. No facing criteria were seen.

The quartzite is white to light greenish gray, has a grain size of 4–5 mm, and contains light-green muscovite in addition to quartz. Amphibolite layers and lenses are more numerous in the unit than in the adjacent schist and gneiss. Other rock types in the unit are biotite-feldspar-quartz gneiss and schist, sillimanite-biotite schist, and sillimanite-garnet-biotite schist and gneiss.

PETROGRAPHY

The quartzites in this unit are composed primarily of a granoblastic aggregate of interlocking grains of quartz as much as 12 mm in diameter (fig. 15). Muscovite occurs in synkinematic to postkinematic crystals 1–4 mm long. Some rocks contain scattered grains of plagioclase, and a few have a small amount of biotite. Light-brown zircon, a common accessory mineral, forms anhedral to subhedral grains as much as 0.15 mm long. Less common accessory minerals are apatite, sphene and opaque minerals, and rutile(?). One heavy mineral parting seen in thin section consists of zircon and rutile(?). Secondary minerals are sericite, chlorite, and epidote. Other rock types in the unit resemble those described in the section on the schist and gneiss unit.

COMPOSITION AND ORIGIN

The quartzites are composed mostly of quartz with small admixtures of plagioclase and muscovite, and they represent clean quartz sandstones with only small amounts of clastic plagioclase and muscovite or clay minerals.

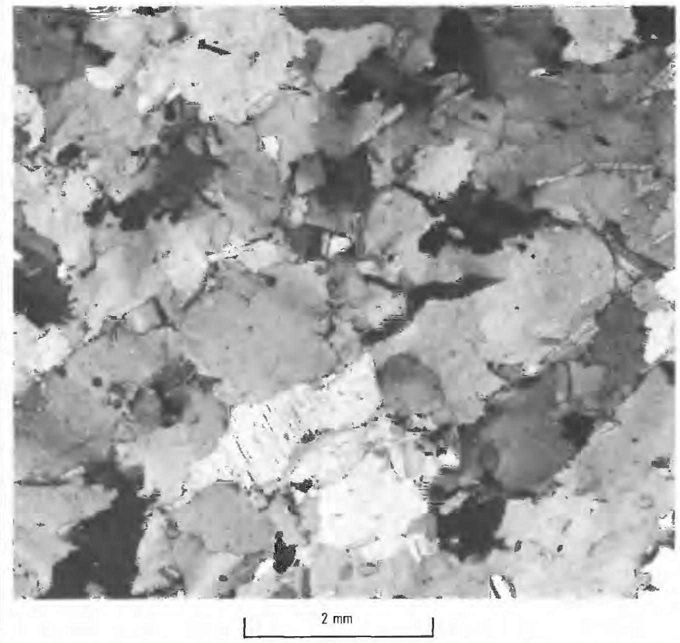


FIGURE 15.—Photomicrograph of typical, white, coarse-grained quartzite. From 8,680 ft (2,645 m) altitude on crest of Sessions Mountains SW¼ sec. 25, T. 2 N., R. 1 E. Granoblastic intergrowth of quartz with scattered flakes of muscovite.

CHEMISTRY OF THE LAYERED METAMORPHIC ROCKS

Chemical analyses of representative samples of gneiss and schist from the three units of layered metamorphic rocks (table 2) show that those rocks were probably derived from sandstone, siltstone, and shale. The rocks cover the range from quartzite through arkose and graywacke to shale, and many samples are close to graywacke in composition (fig. 16*A*). The light-colored gneiss from the migmatite unit (samples 3, 6, and 7) differs from the quartz monzonite gneiss in that it has a lower content of alkalis (fig. 16*B*), even though the quartz monzonite gneiss has a slightly lower SiO_2 content. Na_2O:K_2O ratios in the gneisses from the migmatite unit, and the schist and gneiss unit to the south are variable, in contrast to more uniform ratios in the quartz monzonite gneiss. Sample 5 is from the area of schist and gneiss lacking migmatite, but chemically it resembles the gneisses from the migmatite unit. Sample 7 plots with the layered gneisses, although it resembles the quartz monzonite gneiss in the field and under the microscope. Samples from the migmatite have a similar scatter on the Na_2O-K_2O diagram as those from the schist and gneiss unit. However, on the Al_2O_3/CaO-Na_2O-K_2O diagram, those samples are concentrated near the mean composition of graywacke rather than being spread along the trend that connects quartzite near the bottom of the diagram with shale,

TABLE 2.—*Chemical analyses and modes of gneiss, schist, and quartzite*

[Major oxides determined by X-ray spectroscopy by J. S. Wahlberg, J. T. Taggart, and James Baker, U.S. Geological Survey 1980 and 1981. FeO determined by chemical means by Edythe Engleman, U.S. Geological Survey 1981 and 1982, and Fe_2O_3 calculated by difference with Fe_2TO_3. Ten minor elements determined by quantitative emission spectroscopy by Carol Gent, U.S. Geological Survey, 1981, and Paul Briggs, U.S. Geological Survey, 1983. Other minor elements determined by ICP direct-reading semiquantitative emission spectroscopy by George Riddle, U.S. Geological Survey, 1981. Looked for but not found: Bi, Cd, U, Tb. Mode by count of 1,000 points. P, present but not intersected in counting; leaders (---), not present; N, not determined; <, less than. Major oxides in weight percent; minor elements in parts per million; mode in volume percent]

Sample No.-	1	2	3	4	5	6	7	8	9	10	11	12	13	14
Lab No.——	D-227768	D-227769	D-234328	D-234329	D-234338	D-227766	D-227765	D-234333	D-234332	D-234330	D-234335	D-234336	D-234339	D-234337
Field No.—	B-118-Z	B-118SSR	O-65	F-12-A	F-24-A	O-64	O-57-1-B	B-37-B	B-35	F-44-A	B-121	MB-24	F-27-B	F-12-A
	Gneiss										Quartzite	Schist		
Major oxides (percent)														
SiO_2	60.3	62.1	71.6	72.1	72.3	72.7	73.1	81.2	82.0	86.5	94.4	44.8	50.0	57.5
Al_2O_3	17.4	16.4	12.1	12.4	12.9	11.9	11.9	9.44	9.30	6.80	3.36	29.9	20.9	18.6
Fe_2O_3	0.69	0.65	2.30	0.66	1.45	1.74	1.47	0.19	0.28	0.18	0.02	2.33	0.83	0.99
FeO	6.46	6.98	4.68	4.92	3.28	4.81	3.86	0.52	0.40	0.39	0.04	6.55	9.20	7.24
MgO	3.3	3.1	0.37	0.30	0.33	0.3	0.3	0.33	0.33	0.36	<0.10	2.59	5.68	2.66
CaO	2.96	4.08	2.61	1.43	2.80	2.48	2.02	1.17	0.64	0.25	0.03	0.58	0.97	3.05
Na_2O	2.3	1.7	2.60	1.98	3.40	2.2	2.2	1.71	1.93	1.12	0.16	2.08	0.82	3.74
K_2O	2.68	1.69	1.92	3.86	1.40	2.34	3.32	3.17	3.61	2.75	0.61	6.69	3.15	2.22
TiO_2O	0.83	0.85	0.54	0.40	0.30	0.48	0.32	0.11	0.14	0.08	<0.02	1.79	0.91	1.18
P_2O_5	0.2	0.1	0.14	0.12	0.11	0.1	<0.1	<0.05	<0.05	<0.05	<0.05	0.10	0.19	0.20
MnO	0.1	0.09	0.16	0.24	0.11	0.13	0.15	<0.02	<0.02	<0.02	<0.02	0.06	0.08	0.19
LOI	1.65	1.16	0.16	0.34	0.78	0.12	0.35	0.50	0.56	0.56	0.39	1.88	5.07	1.05
Quantitative minor elements (parts per million)														
Ba	725	210	1,100	1,460	654	1,330	1,340	841	671	867	50	1,330	365	723
Cr	122	123	<10	<10	<10	13	10	<10	10	<10	<10	160	500	160
Nb	<2	<2	30	30	40	11	15	<20	<20	<20	<20	<20	<20	<20
Ni	78	82	<10	<10	<10	4	9	<10	<10	<10	<10	110	240	70
Sr	183	142	80	70	150	85	82	180	170	100	<10	60	30	130
Y	25	26	140	170	120	99	126	<10	<10	<10	<10	30	20	40
Zr	227	222	743	910	591	623	832	65	54	55	28	488	110	300
Ce	73	69	190	250	170	174	216	<30	30	<30	<30	160	<30	80
Nd	33	26	90	120	80	80	97	<40	<40	<40	<40	90	<40	50
Rb	100	50	65	140	80	75	85	110	96	64	<50	420	180	130
Semiquantitative minor elements (parts per million)														
Be	2.5	2.3	3	2	8	2.5	3	2	<1	<1	<1	2	2	1
Co	39	27	4	3	3	1	1	3	4	<2	<2	47	45	31
Cu	96	150	15	8	8	14	36	6	36	16	8	310	6	61
Ga	17	15	10	20	20	18	20	<10	<10	<10	<10	20	10	20
Ge	N	N	<10	<10	<10	N	N	N	<10	<10	<10	20	60	30
La	39	34	130	160	120	93	110	21	26	19	9	140	38	81
Li	40	32	7	7	19	4	5	7	6	<5	<5	68	230	95
Mo	5	6	<5	<5	<5	7	5	<5	<5	<5	<5	<5	<5	<5

TABLE 2.—*Chemical analyses and modes of gneiss, schist, and quartzite*—Continued

Sample No.—	1	2	3	4	5	6	7	8	9	10	11	12	13	14
Lab No.—	D-227768	D-227769	D-234328	D-234329	D-234338	D-227766	D-227765	D-234333	D-234332	D-234330	D-234335	D-234336	D-234339	D-234337
Field No.—	B-118-Z	B-118RSR	O-65	P-12-A	P-24-A	O-64	O-57-1-B	B-37-B	B-35	P-44-A	B-121	MB-24	P-27-B	P-12-A
	Gneiss										Quartzite	Schist		
Semiquantitative minor elements (parts per million)														
Pb	64	47	20	20	<10	49	57	20	30	20	<10	20	<10	10
Sc	24	23	15	19	16	10	10	6	8	<5	<5	62	28	43
Sn	25	<5	<10	<10	<10	<5	<5	<10	<10	<10	<10	<10	<10	<10
Th	15	15	24	36	26	27	36	6	10	<5	<5	42	14	19
V	130	130	15	<5	18	<3	<3	13	20	11	10	300	210	180
W	N	N	<10	<10	<10	N	N	<10	<10	<10	<10	10	<10	<10
Zn	110	100	150	93	180	120	110	270	43	36	14	440	98	170
Pr	<10	<10	<20	<20	<20	30	30	<20	<20	<20	<20	<20	<20	<20
Sm	<10	<10	20	20	<20	20	20	<20	<20	<20	<20	<20	<20	<20
Eu	<3	<3	<5	<5	<5	3	3	<5	<5	<5	<5	<5	<5	<5
Gd	20	20	N	N	N	30	30	N	N	N	N	N	N	N
Dy	<10	<10	N	N	N	20	30	N	N	N	N	N	N	N
Ho	<3	<3	<10	<10	<10	4	5	<10	<10	<10	<10	<10	<10	<10
Er	<5	<5	N	N	N	11	15	N	N	N	N	N	N	N
Yb	3	3	15	21	N	11	15	<2	<2	<2	<2	<2	2	5
Ta	N	N	20	10	10	N	N	<10	<10	<10	<10	20	30	20
Modes (percent)														
Quartz	19.2	38.9	51.6	40.2	38.7	40.1	32.3	65.3	57.3	68.1	92.4	6.4	26.6	18.3
Plagioclase	41.6	36.7	32.4	19.6	47.9	50.4	27.3	18.0	19.5	12.4	—	7.0	3.8	47.7
Potassic feldspar	P	0.6	8.8	32.0	—	0.6	28.7	13.4	20.5	15.8	—	15.3	—	—
Myrmekite	—	—	—	P	—	—	P	—	—	—	—	—	—	—
Biotite	30.2	18.4	3.8	0.9	1.5	—	—	1.2	1.9	0.7	—	33.6	21.8	26.4
Muscovite	—	—	—	—	—	—	—	—	P	1.2	—	1.8	7.5	—
Amphibole	—	—	—	—	6.5	5.5	9.3	—	—	—	—	—	—	—
Sillimanite	—	—	—	—	—	—	—	—	—	—	—	26.5	—	0.1
Garnet	3.9	2.8	1.7	5.9	0.3	2.0	—	P	—	—	—	—	—	5.1
Opaque minerals	0.1	0.2	1.1	0.4	—	1.2	2.0	—	P	0.2	—	1.5	—	0.5
Apatite	0.1	P	P	0.2	0.2	P	P	—	—	P	—	—	—	0.2
Zircon	P	P	0.1	P	P	P	P	0.2	P	P	P	0.1	P	P
Allanite	—	—	0.4	0.6	0.1	0.2	0.4	—	—	—	—	—	—	—
Sericite	3.8	1.1	—	—	—	—	—	0.6	0.2	1.4	—	7.8	21.0	1.0
Chlorite	0.8	0.9	0.1	0.2	3.2	—	—	1.3	0.6	0.2	—	P	19.2	0.6
Epidote	0.3	0.4	—	P	1.6	—	—	—	P	—	—	—	P	0.1
Sphene	—	—	—	—	P	—	—	—	—	—	—	—	0.5	—
Muscovite and sericite	—	—	—	—	—	—	—	—	—	—	7.6	—	—	—

TABLE 2.—*Chemical analyses and modes of gneiss, schist, and quartzite*—Continued

LOCATION AND DESCRIPTION OF SAMPLES

1. Garnet-quartz-plagioclase biotite gneiss from roadcut, 180 m west of BM-7946 on Skyline Drive, west side of crest of Wasatch Mountains east of Centerville, Bountiful Peak 7.5-minute quadrangle. Synkinematic biotite as much as 3.5 mm long, anhedral plagioclase (An 30) as much as 2.5 mm in diameter locally saussuritized, anhedral strongly strained quartz as much as 2 mm in diameter, and subhedral to euhedral garnet 1-2 mm in diameter. Dated sample (Hedge and others, 1983).
2. Garnet-biotite-plagioclase-quartz gneiss from roadcut 185 m west of BM-7946 on Skyline Drive west side crest of Wasatch Mountains. Bountiful Peak 7.5-minute quadrangle. Anhedral strained quartz as much as 1.5 mm in diameter, anhedral locally saussuritized plagioclase (An 40-48) as much as 1 mm in diameter, and synkinematic to postkinematic locally bent biotite as much as 2 mm long. Dated sample (Hedge and others, 1983).
3. Garnet-bearing, biotite-plagioclase-quartz gneiss, from pipeline cut at 4,760 ft (1,450 m) altitude on the south side of Weber Canyon, S. 38° W. from powerhouse, Ogden 7.5-minute quadrangle. Migmatitic gneiss with lenses of amphibolite and dikes of pegmatite. Anhedral quartz as much as 3 mm long, anhedral plagioclase (An 28) as much as 1 mm in diameter, anhedral potassic feldspar, synkinematic biotite as much as 1 mm long, and anhedral to subhedral light-brown garnet as much as 2 mm in diameter with some sieve texture with quartz. Dated sample (Hedge and others, 1983).
4. Biotite-garnet-plagioclase-microcline quartz gneiss from migmatite unit at 9,640 ft (2,938 m) altitude on ridge 120 m, N. 20° W. from VABM 9797 on crest of Wasatch Mountains, Peterson 7.5-minute-qusdrangle. Anhedral quartz as much as 3 mm in diameter with large grains replacing feldspar, anhedral partly saussuritized plagioclase (An 25) as much as 1 mm in diameter, anhedral microcline as much as 1.3 mm in diameter, anhedral to subhedral garnet as much as 1.5 mm in diameter with inclusions of allanite, quartz, plagioclase, and apatite and fractures filled with biotite, epidote, and chlorite, and synkinematic and postkinematic biotite as much as 1.7 mm long.
5. Garnet-bearing biotite-hornblende-quartz plagioclase gneiss from 6,460 ft (1,970 m) altitude on the south side of Mill Creek valley 660 m S. 80° W. from NE corner sec. 2, T. 1 N. R. 1 E, Fort Douglas 7.5-minute quadrangle. Layered gneiss with semiconcordant pegmatites. Anhedral, partly saussuritized plagioclase (An 26) as much as 2 mm in diameter; anhedral quartz as much as 2 mm in diameter, concentrated in a layer possibly in part due to metamorphic segregation; dark-green anhedral hastingsitic hornblende as much as 1.7 mm long, concentrated in a layer; synkinematic to postkinematic biotite as much as 0.5 mm long, concentrated in a layer; and anhedral garnet as much as 0.7 mm in diameter concentrated in biotitic layer.
6. Garnet-hornblende-quartz-plagioclase gneiss from pipeline cut at 4,800 ft (1,463 m) altitude, S. 13° E. from powerhouse in Weber Canyon, Ogden 7.5-minute quadrangle. Outcrop of typical migmatitic gneiss containing lenses of amphibolite and pegmatites. Anhedral locally saussuritized plagioclase (An 27) as much as 1.5 mm in diameter, anhedral strongly strained quartz as much as 4 mm long, synkinematic anhedral brownish-green hastingsitic hornblende as much as 2 mm long, and anhedral to subhedral light-brown garnet as much as 2 mm long. Dated sample (Hedge and others, 1983).
7. Hornblende quartz monzonite gneiss from outcrop of swirly gneiss with amphibolite lenses and pegmatite layers just south of U.S. Interstate Highway I-80 in Weber Canyon 70 m west of east edge of Ogden 7.5-minute quadrangle. Anhedral microcline to 1.5 mm long, quartz to 3 mm long, plagioclase (An 20-25) to 1 mm in diameter, and synkinematic dark-green ferrohastingsite as much as 2 mm long, has $2V=20°$ and $Z\Lambda C=13°$. Dated sample (Hedge and others, 1983).
8. Garnet-bearing biotite-microcline-plagioclase-quartz gneiss from 8,700 ft (2,650 m) altitude, 440 m S. 54° E. from top of Bountiful Peak, Bountiful Peak 7.5-minute quadrangle. More quartzitic layer from outcrop of thinly laminated biotite schist and gneiss. Anhedral, strongly strained quartz as much as 3 mm long; anhedral locally saussuritized plagioclase (An 30) as much as 0.6 mm in diameter; anhedral partly perthitic microcline as much as 0.6 mm in diameter, postkinematic biotite as much as 0.5 mm long; zircon and secondary calcite and epidote.
9. Biotite-plagioclase-microcline-quartz gneiss from 8,920 ft (2,719 m) altitude, 160 m N. 20° E. from Bountiful Peak, Bountiful Peak 7.5-minute quadrangle. Anhedral porphyroblasts of quartz as much as 4 mm in diameter, strongly strained and partly broken; anhedral saussuritized plagioclase as much as 1 mm in diameter; anhedral somewhat perthitic microcline as much as 1 mm in diameter; and slightly bent biotite as much as 0.6 mm long.
10. Biotite-bearing plagioclase-microcline-quartz gneiss from 8,320 ft (2,535 m) altitude on ridge south of Smith Creek Lakes. Peterson 7.5 minute quadrangle. Gneiss layer in migmatite unit. Anhedral strongly strained quartz as much as 4 mm in diameter, anhedral sericitized and saussuritized plagioclase as much as 1 mm in diameter, synkinematic biotite as much as 0.4 mm long, and postkinematic muscovite as much as 0.6 mm long.
11. Muscovite quartzite from pipeline cut, 100 m S. 32° E. of spot elevation 8,705 ft (2,653 m) between heads of Holbrook and Ward Canyons, Bountiful Peak 7.5-minute quadrangle. Quartzite unit. Granoblastic textured aggregates of interlocking quartz grains as much as 8 mm in diameter, postkinematic muscovite as much as 3 mm long and in graphic intergrowths with quartz, and sericite from muscovite.
12. Plagioclase-microcline-sillimanite schist from 5,620 ft (1,713 m) altitude on west side of Hornet Creek, 200 m S. 20° W. from center of sec. 20, T. 3 N., R. 1 E., Bountiful Peak 7.5-minute quadrangle. Synkinematic to postkinematic biotite as much as 4 mm long; sillimanite as much as 3 mm long but mostly 1 mm occurs in polygonal arcs, is bent, and is parallel with axial plane of folds indicating crystallization before and after folding; anhedral locally perthitic microcline as much as 2 mm in diameter; anhedral plagioclase (An 22) as much as 2 mm in diameter; and muscovite intergrown with biotite.
13. Somewhat sheared and retrogressively metamorphosed muscovite-biotite schist from 6,760 ft (2,106 m) altitude on south slope of Mill Creek valley, 170 m N. 70° W. from SE corner sec. 35, T. 1 N., R. 1 E., Fort Douglas 7.5-minute quadrangle. Synkinematic and postkinematic biotite to 2 mm diameter; muscovite to 2 mm long intergrown with biotite; and anhedral plagioclase (An 28) to 2 mm long; anhedral quartz in thin segregation stringer.
14. Garnet-biotite-plagioclase schist 180 m N. 32° W. from top of Big Rock in Mill Creek Valley, Fort Douglas 7.5-minute quadrangle. Anhedral locally saussuritized plagioclase (An 30-32) as much as 1 mm in diameter, synkinematic to postkinematic biotite as much as 2 mm long, anhedral quartz as much as 2 mm in diameter but mostly less than 1 mm, and subhedral garnet as much as 1 mm in diameter with sieve texture formed by inclusions of quartz, plagioclase, biotite, and opaque minerals. A needle of sillimanite is included in plagioclase.

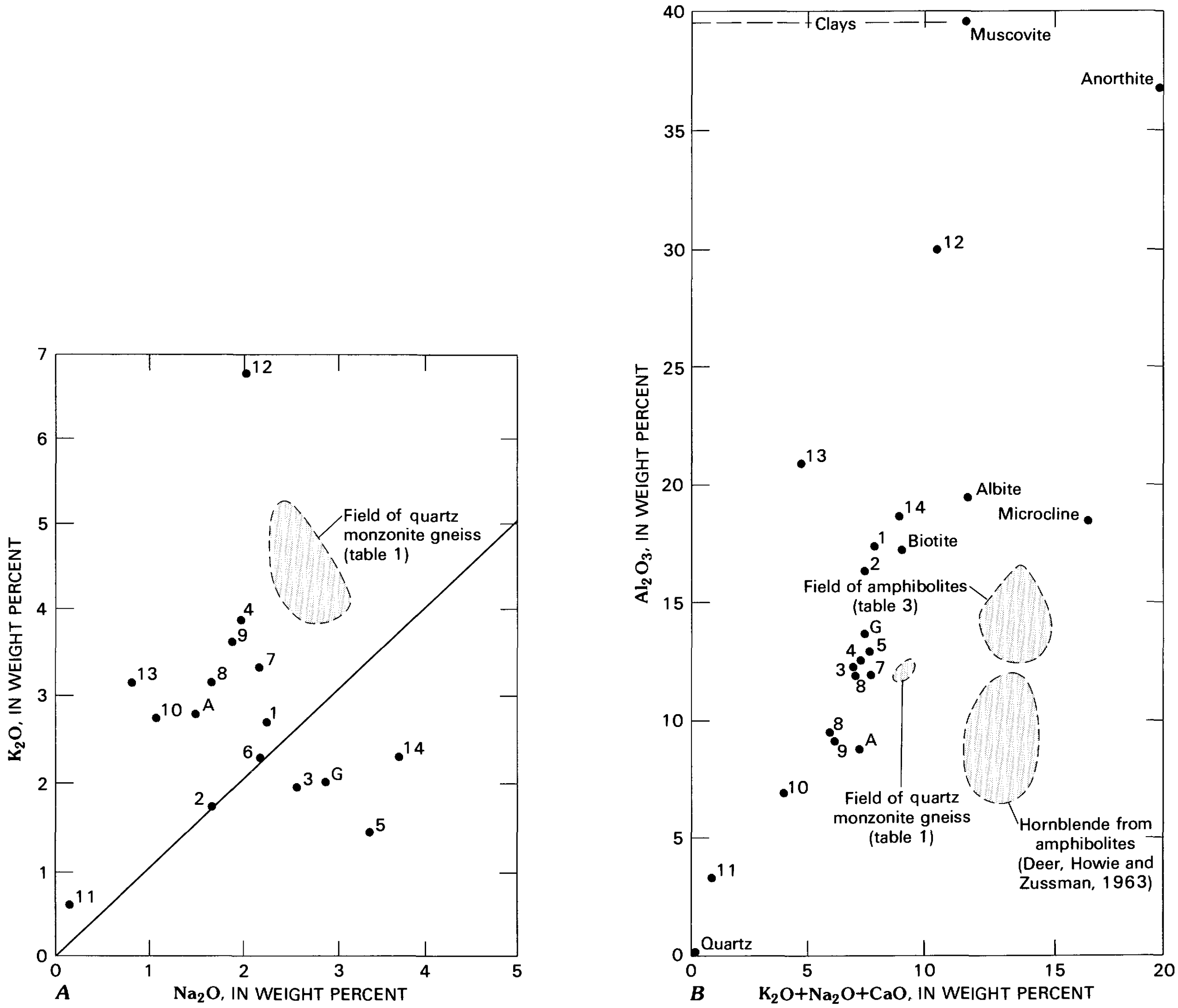


FIGURE 16.—Diagrams showing chemical ratios of analyzed samples in the Farmington Canyon Complex. Numbers refer to analyzed samples in table 2. A, Mean composition of arkose; G, mean composition of graywacke (Pettijohn, 1963). *A*, K_2O/Na_2O diagram. Line represents ratio of 1.0. *B*, Al_2O_3/K_2O+Na_2O+CaO diagram showing position of selected minerals and fields of quartz monzonite gneiss and amphibolite of the Farmington Canyon Complex.

which would plot in the upper part. Minor elements (in samples 3–7) of gneiss are quite similar to those in the quartz monzonite gneiss (table 1). That is permissive evidence for derivation of the quartz monzonite gneiss by melting of metamorphic rocks like those exposed in the area.

MINOR ROCK UNITS

AMPHIBOLITES

Amphibolites are widely distributed throughout the Farmington Canyon Complex, but only a few amphibolite bodies are shown on plate 1, because most of the amphibolites are too small to show at the scale of the map.

The amphibolites typically occur in lenses only 1–3 m thick and a few meters to 10 m long. The largest amphibolites are as much as 10 m thick and 500 m long. One body that size is shown on the map (pl. 1) on the west side of Bountiful Peak. At Devils Gate, some pod-shaped amphibolites as much as 7 m thick are exposed. The amphibolites generally have sharp contacts with the country rock, are nonlayered, and are cut by pegmatites. On the whole, pegmatites are less numerous in the amphibolites than in their country rock. Rarely do the amphibolites appear to have been weakly migmatized.

A few amphibolites have layering. One garnet amphibolite contains partings of garnet-rich biotite schist. Another amphibolite grades to garnet-hornblende gneiss and garnet-biotite-quartz gneiss.

The amphibolites usually have a medium-grained granoblastic to nemanoblastic texture (fig. 17*B*). Granoblastic texture is best developed in the interiors of the pods exposed at Devils Gate. Some amphibolites have lenses of feldspar and quartz that suggest a relict gabbroic texture. In highly lineated rocks, these lenses have been stretched out into rods. One short dikelike body in the quartz monzonite gneiss has a relict diabasic texture in which some plagioclase retains a lathlike shape, and recrystallized pyroxene and hornblende fill in around the plagioclase grains (fig. 17*A*). The least metamorphosed mafic igneous rocks are gabbro pods that occur in the quartz monzonite gneiss south of Ogden Peak. The gabbro pods have an ophitic texture in which laths of labradorite are enclosed in monoclinic pyroxene and later (metamorphic?) brown hornblende from pyroxene occurs in mosaic-textured aggregate.

Plagioclase makes up 30–50 percent of the amphibolite and is composed of anhedral to subhedral calcic oligoclase to sodic labradorite. Plagioclase is saussuritized in many rocks, especially in the central part of the complex where the effects of cataclasis and retrogressive metamorphism are more intense.

Hornblende (45–70 percent) occurs in anhedral to subhedral grains averaging 2 mm long, but reaching 7 mm long. It is olive green, brownish green, yellowish green, and moderately light green in the Z direction. It has a high $2V$ and $Z \wedge C$ of 16°–19°. Locally, it is altered to very light green actinolite with the same extinction angle.

Anhedral to subhedral grains, about 1 mm in diameter, of light-green to colorless monoclinic pyroxene with $Z \wedge C$ 42°–44°, make up as much as 20 percent of some amphibolites. Amphibolite containing monoclinic pyroxene generally occurs north of Farmington Canyon in and near the area of migmatite and in the quartz monzonite gneiss. The monoclinic pyroxene is contemporaneous with hornblende in the amphibolite and is thus a part of the main mineral assemblage. The area in which monoclinic pyroxene constitutes several percent or more of some amphibolites is coextensive with the area in which sillimanite and microcline are present, and muscovite absent, in aluminum-rich schists.

Quartz occurs in grains 0.2–2 mm in diameter and amounts to as much as 8 percent in some amphibolites. In some rocks, it is concentrated with the plagioclase in light-colored lenses. Postkinematic biotite, as much as 1.5 mm long, occurs in some amphibolites, and it makes up as much as 10–15 percent of the rock in the pods at Devils Gate.

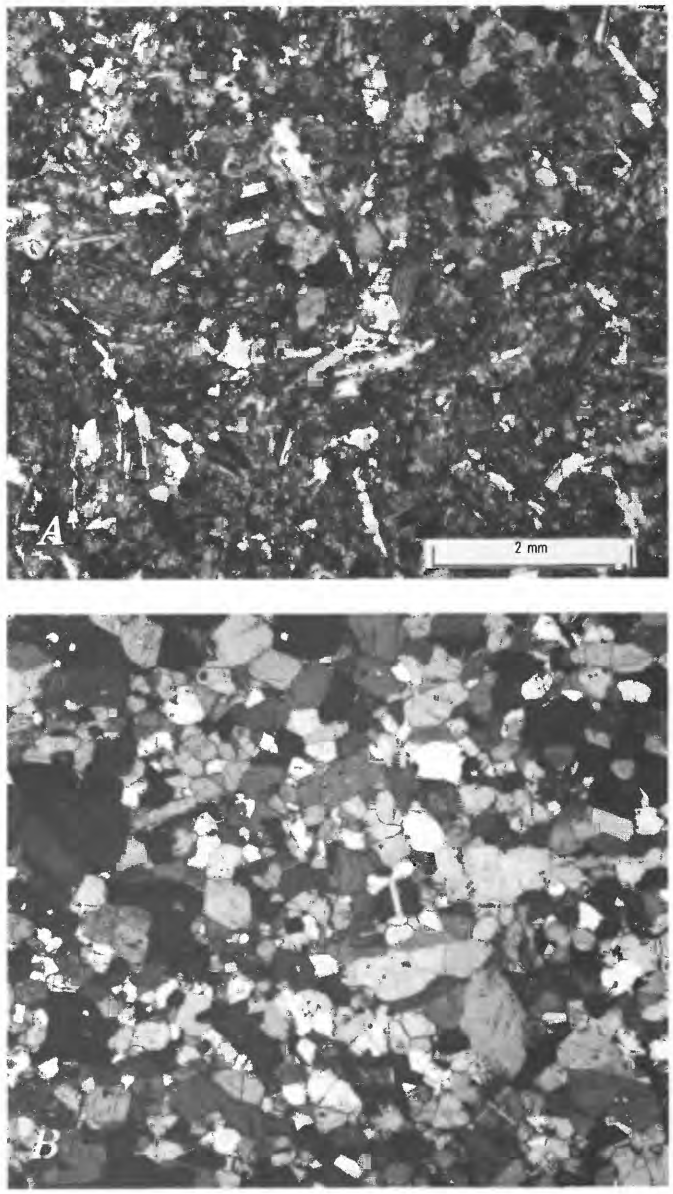


FIGURE 17.—Photomicrographs of amphibolite. *A*, Metadiabase from 10-m-thick discontinuous dike cut by pegmatite (probably an inclusion) in quartz-monzonite gneiss at 7,890 ft (2,405 m) altitude on southeast branch of Taylor Canyon. Relict diabasic texture unusual in amphibolite in Farmington Canyon Complex. Plagioclase is calcic andesine and sodic labradorite. Mafic minerals are monoclinic pyroxene and hornblende that appear to have recrystallized during metamorphism. Analyzed sample 3, table 3. *B*, Typical granoblastic amphibolite from 10-m-thick lens in quartzite at 8,350 ft (2,545 m) altitude on pipeline near the center of sec. 24, T. 2 N., R. 1 E., on ridge between Holbrook and Ward Canyons. Mosaic-textured hornblende and labradorite. Analyzed sample 5, table 3.

Apatite, sphene, and opaque minerals are common accessories. Much of the opaque mineral is nonmagnetic under a hand magnet and is probably ilmenite; however, a few amphibolites contain magnetite. Zircon occurs sparingly as tiny, round grains of 0.05 mm or less in diameter in a few amphibolites. Allanite was found in two samples. Secondary epidote and chlorite are common, and they make up a substantial proportion of the amphibolites in zones of shearing and retrogressive metamorphism.

COMPOSITION AND ORIGIN

Analyses of amphibolites (table 3) show that they have basaltic compositions; SiO_2 ranges from 47 to 52 percent. Alkali contents are somewhat variable and samples 1 and 4 fall into the alkali-basalt field of Macdonald and Katsura (1964) (fig. 18*C*). Samples 1–4 have olivine in their norms, whereas samples 5–7 have normative quartz. TiO_2 contents are less than that of most basalt, but they are within the range given for tholeiites (Winchester and Floyd, 1977) and close to the average of Nockolds' (1954) "central basalt." Zirconium and cerium contents are low, but in the range of values obtained from tholeiites (Winchester and Floyd, 1977).

On a TiO_2-K_2O-P_2O_5 diagram (fig. 18*A*), amphibolites from the Farmington Canyon Complex fall into two groups. Analyses 1–4 fall well within the field designated continental basalts, whereas 5–7 are in the field of oceanic basalts. Pearce and others (1975) suggested that metamorphism will tend to enrich basalts in K_2O and thus tend to move them toward the field of continental basalt. In the Farmington Canyon Complex that is, indeed, what may have happened, for samples 1–4 are either in or near the area where the rock is migmatized or occur as inclusions in the quartz monzonite gneiss, whereas samples 5–7 are from farther south in the schist and gneiss unit. In addition to being from areas of migmatization and melting, these rocks have more secondary minerals due to retrogressive alteration than samples 5–7, as shown by the actinolite and chlorite in their modes. The hypothesis that the samples of amphibolite from the northern part of the area differ from those in the southern part, due to metamorphic effects, is supported by their tight grouping on a FeO^*-MgO-Al_2O_3 diagram (fig. 18*B*), even though some of the analyses do not pass the screen used in that study (51–56 percent SiO_2 content and a tholeiitic classification based on the Macdonald and Katsura alkali-silica diagram). FeO, MgO and A_2O_3 are immobile relative to K_2O during metamorphism in the amphibolite facies, although they could be mobile in greenschist-facies metamorphism during which numerous epidote segregations may form in mafic igneous rocks. The analyses plot in a field of oceanic basalts on the diagram, but whether the detailed division of tectonic environments given, based on Cenozoic igneous rocks, should be applied to Archean rocks is uncertain (Pearce and others, 1977).

In the alkali-silica diagram (fig. 18*C*), samples 6, 7, and 8 from the less migmatized and melted rocks in the southern part of the area are well within the tholeiite field, whereas the ones from the migmatite and quartz monzonite are near the line separating tholeiite from alkali basalt, probably for the same reasons as discussed above.

The analyses of the amphibolites from the southern part of the area fall in field B in the Ti-Zr diagram (fig. 18*D*). That field is not particularly diagnostic, since it includes all three possible rock categories. However, three of the four samples that fall in the field of calc-alkali basalts are from the northern part of the area, so that the differences between the amphibolite from areas of migmatization and melting, and those from areas lacking evidence for action of these processes, are apparent also in that diagram.

The chemical and modal compositions of the amphibolites indicate that they were derived from igneous rocks of basaltic composition. Relict textures in a few bodies show that they were originally gabbro or diabase. The lenticular shape and small size of many of the amphibolites suggest that they may be tectonically disrupted sills or dikes emplaced before deformation and metamorphism. Amphibolites in the quartz monzonite gneiss are interpreted as undigested relicts from the country rock, which was melted to form the quartz monzonite. As in many parts of the Archean crust, the mafic intrusives may be of several ages, but crosscutting relations between amphibolites were not found during this study. However, the gabbro pods south of Ogden Peak may be younger than the amphibolites included in the quartz monzonite gneiss and may have intruded during or just after the formation of the foliation in the quartz monzonite gneiss.

MICA SCHIST

A discrete body of mica schist that occurs in the Farmington Canyon Complex below the Ogden thrust from Waterfall Canyon south about 1.5 km consists of muscovite-biotite-plagioclase-quartz schist and contains mica flakes as much as 5 cm in diameter. Some layers of garnet-biotite-feldspar-quartz gneiss and some lenses of amphibolite and poorly foliated metadiorite are present. The rocks of this unit differ from many of the schist layers in the schist and gneiss unit and the migmatite unit in that muscovite is a conspicuous mineral, and other aluminum-silicate minerals were not found in the samples studied.

The mica schist contains 20–50 percent mica in synkinematic flakes as much as several centimeters

TABLE 3.—*Chemical analyses, modes, and norms of amphibolite and metadiabase*

[Major oxides determined by X-ray spectroscopy by J. S. Wahlberg, J. T. Taggart, and James Baker, U.S. Geological Survey 1980 and 1981. FeO determined by chemical means by Edythe Engleman, U.S. Geological Survey 1981 and 1982, and Fe_2O_3 calculated by difference with Fe_2TO_3. Ten minor elements determined by quantitative emission spectroscopy by Carol Gent, U.S. Geological Survey, 1981, and Paul Briggs, U.S. Geological Survey, 1983. Other minor elements determined by ICP direct reading semiquantitative emission spectroscopy by George Riddle, U.S. Geological Survey, 1981. Looked for but not found: Bi, Cd, U, Pr, Sm, Eu, Tb, Dy; (---), not present; N, not detected; <, less than. Major oxides and CIPW norm in weight percent; minor elements in parts per million; mode in volume percent. CIPW norms calculated on basis of analysis recalculated to 100 percent after deduction of volatiles]

Sample No.-	1	2	3	4	5	6	7
Lab No.----	D-227764	D-234331	D-234327	D-234326	D-227770	D-234334	D-227767
Field No.--	0-57-1-a	P-90-A	0-13	0-38-B	B-122-A	B-120-A	B-118-A
Major oxides (percent)							
SiO_2	47.2	48.2	48.0	50.1	51.1	51.1	51.7
Al_2O_3	15.2	13.5	16.4	14.0	12.6	14.4	14.5
Fe_2O_3	2.42	2.98	1.40	1.88	1.65	2.46	1.42
FeO	9.36	9.07	9.41	9.16	8.92	9.98	9.75
MgO	7.84	7.65	7.64	7.42	9.10	6.53	6.74
CaO	10.0	11.9	10.4	9.79	12.5	10.0	11.4
Na_2O	1.9	2.17	2.10	3.64	1.1	2.15	1.8
K_2O	1.71	1.05	1.53	1.14	0.18	0.59	0.41
TiO_2	1.06	0.90	1.06	0.81	0.72	1.06	0.89
P_2O_5	0.2	0.07	0.1	0.07	<0.1	0.09	<0.1
MnO	0.18	0.22	0.17	0.23	0.19	0.23	0.20
LOI	1.24	0.66	1.27	0.90	0.50	0.63	0.40
Quantitative minor elements (parts per million)							
Ba	253	95	216	280	14	139	105
Cr	185	90	210	90	486	10	42
Nb	<2	<20	<20	<20	<2	<20	<2
Ni	174	80	110	110	140	50	51
Sr	119	90	170	140	95	200	142
Y	14	20	30	20	11	20	14
Zr	77	51	90	70	44	65	56
Ce	19	<30	<30	<30	<10	<30	13

TABLE 3.—*Chemical analyses, modes, and norms of amphibolite and metadiabase*—Continued

Sample No.-	1	2	3	4	5	6	7
Lab No.----	D-227764	D-234331	D-234327	D-234326	D-227770	D-234334	D-227767
Field No.--	0-57-1-a	P-90-A	0-13	0-38-B	B-122-A	B-120-A	B-118-A
Quantitative minor elements (parts per million)							
Nd	11	<40	<40	<40	<10	<40	<10
Rb	70	<50	<50	<50	<10	<50	<10
Semiquantitative trace elements (parts per million)							
Be	1.6	1	<1	<1	0.6	1	0.8
Co	45	53	42	44	42	46	39
Cu	110	120	69	8	47	150	150
La	8	<5	15	12	<3	8	4
Mo	6	<5	<5	<5	<3	<5	5
Pb	51	20	20	30	23	<10	36
Sc	32	64	33	43	44	56	47
V	190	300	220	230	230	330	260
Zn	110	110	95	130	64	130	85
Ga	15	<10	<10	<10	5	10	14
Ge	N	70	60	60	N	50	N
Li	11	<5	12	10	9	5	5
Ta	N	20	20	20	N	20	N
Th	<3	6	<5	<5	4	<5	<3
Yb	2	3	3	2	1	3	2
Gd	20	N	N	N	20	N	20
Modes (percent)							
Quartz	--	--	--	--	--	5.4	2.9
Plagioclase	46.0	36.5	44.5	37.6	32.9	28.3	39.0

TABLE 3.—*Chemical analyses, modes, and norms of amphibolite and metadiabase*—Continued

Sample No.-	1	2	3	4	5	6	7
Lab No.----	D-227764	D-234331	D-234327	D-234326	D-227770	D-234334	D-227767
Field No.--	0-57-1-a	P-90-A	0-13	0-38-B	B-122-A	B-120-A	B-118-A
Modes (percent)							
Hornblende	29.1	52.5	28.6	53.1	66.6	64.6	59.7
Biotite	--	--	1.1	--	--	P	--
Pyroxene	--	3.2	19.7	8.3	--	--	--
Opaque	0.2	P	1.4	--	0.5	0.6	--
Actinolite	11.4	6.4	3.9	0.6	--	--	--
Epidote	P	1.3	--	P	--	0.8	P
Chlorite	12.9	--	0.5	0.3	--	0.2	--
Apatite	P	0.1	--	0.1	--	0.1	P
Sphene	--	P	0.3	--	P	P	0.4
Prehnite	--	--	--	--	--	P	--
Zircon	--	--	--	--	P	--	--

LOCATION AND DESCRIPTION OF SAMPLES

1. Amphibolite lens from exposure of layered swirly biotite-quartz-feldspar gneiss containing pegmatite layers from outcrop just south of U.S. Interstate Highway 80 in Weber Canyon, 780 m S. 87° E. of power plant dam in Weber Canyon, Ogden 7.5-minute quadrangle. Partly retrogressively metamorphosed amphibolite containing tannish-green hornblende to 2 mm long with $Z \wedge C = 18°$ and partly altered to actinolite, chlorite, epidote, and sphene; and mostly saussuritized plagioclase (An 28) as much as 1.5 mm in diameter. Dated sample (Hedge and others, 1983).
2. Amphibolite from 0.5-m-thick pod cut by pegmatite in biotite-quartz-feldspar gneiss from roadcut on Farmington Canyon road, 180 m S. 58° W. from Halfway Creek crossing, Peterson 7.5-minute quadrangle. Brownish-green, anhedral to subhedral hornblende to 3 mm long, locally saussuritized anhedral plagioclase (An 38) to 0.5 mm in diameter, very light tannish green anhedral to subhedral monoclinic pyroxene to 1 mm in diameter, and accessory opaque mineral and sphene. Some light-green secondary amphibole after hornblende and pyroxene.
3. Metadiabase from discontinuous mafic dike about 10 m thick cut by pegmatite from 7,890 ft (2,405 m) altitude in gully, 1.17 km N. 45° W. from summit of Mount Ogden, Ogden 7.5-minute quadrangle. Subhedral plagioclase (An 45–60) as much as 1.7 mm long; anhedral monoclinic pyroxene, mostly 0.1–0.3 mm in diameter but as large as 1 mm in aggregates and between plagioclase grains; anhedral to subhedral brown hornblende as much as 0.3 m long in aggregates and at least in part derived from pyroxene; biotite to 1 mm in diameter; light-green secondary amphibole from hornblende, and opaque mineral. Sericite from plagioclase and chlorite in late fracture.
4. Two-meter-thick amphibolite lens in quartz monzonite gneiss at 8,250 ft (2,515 m) altitude on ridge east of Strongs Peak, Ogden 7.5-minute quadrangle. Anhedral to subhedral, yellowish-green hornblende as much as 3 mm long locally altered to light-bluish-green to colorless amphibole; partly saussuritized anhedral plagioclase (An 27) as much as 2 mm in diameter; and very light green anhedral monoclinic pyroxene as much as 3 mm in diameter but mostly less than 2 mm.
5. Amphibolite about 10 m thick at 8,350 ft (2,545 m) altitude on pipeline on west side of second saddle west of 8,705 ft (2,653 m) knob between the heads of Ward and Holbrook Canyons, Bountiful Peak 7.5-minute quadrangle. Fairly light green, anhedral to subhedral hornblende as much as 1.5 mm long but mainly in mosaic texture aggregates of grains less than 1 mm in diameter, and locally saussuritized plagioclase (An 60) with some reverse zoning. Dated sample (Hedge and others, 1973).
6. Amphibolite about 10 m thick, 380 m S. 66° W. summit of Bountiful Peak, Bountiful Peak 7.5-minute quadrangle. Olive-green anhedral hornblende as much as 3 mm long, anhedral plagioclase (An 50) as much as 1 mm in diameter, and anhedral quartz. Secondary epidote, prehnite, and chlorite concentrated in a fracture.
7. Amphibolite layer 15 m thick in roadcut 140 m west of BM-7946 on Skyline Drive west side crest of Wasatch Mountains east of Centerville. Bountiful Peak 7.5-minute quadrangle. Anhedral brownish-green hornblende as much as 3 mm long, anhedral plagioclase (An 40–45) to 1 mm in diameter, and anhedral quartz. Dated sample (Hedge and others, 1983).

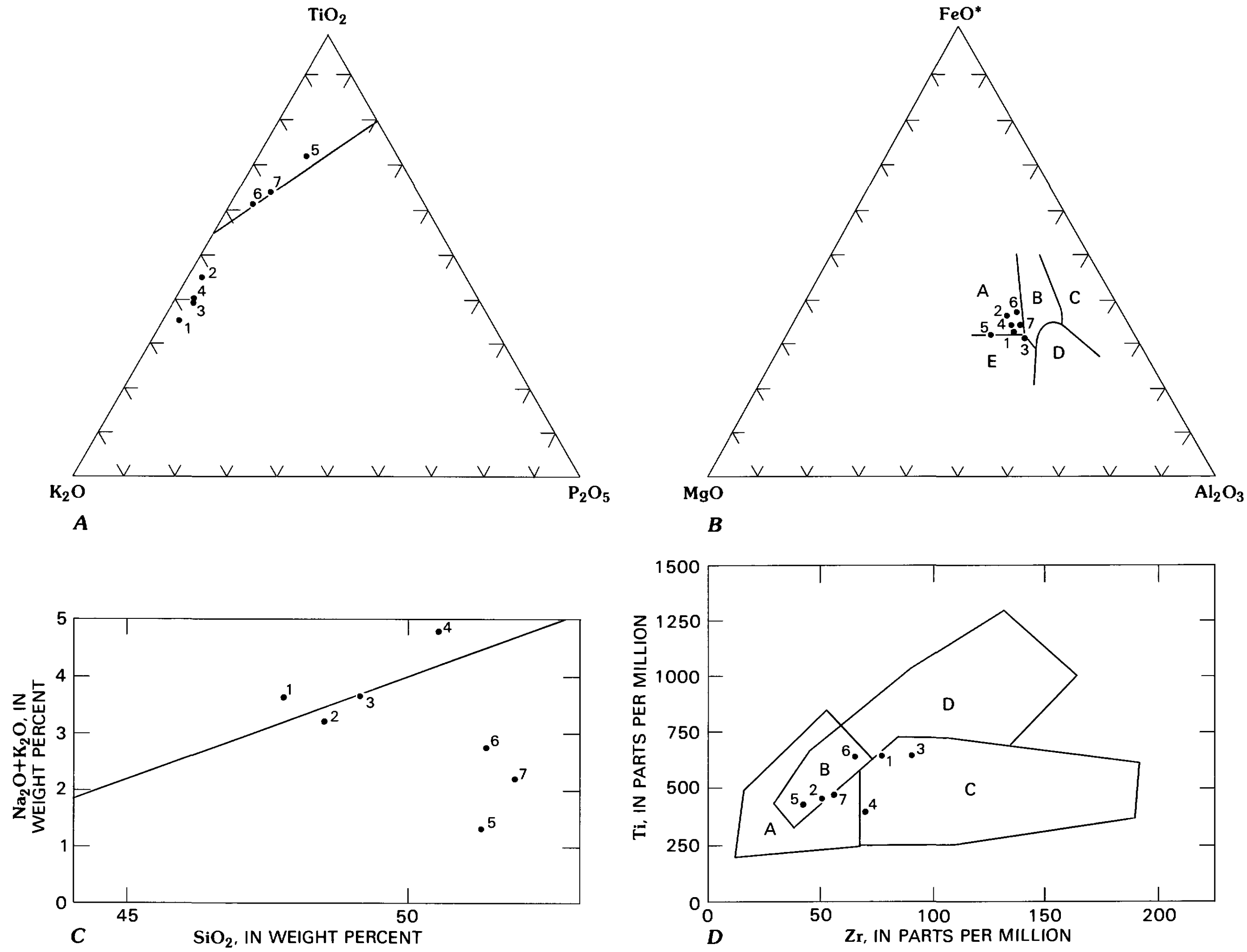


FIGURE 18.—Chemical composition diagrams for amphibolites. Numbers refer to analyzed samples in table 3. *A*, TiO_2-K_2O-P_2O_5 diagram. Line divides oceanic (above) from nonoceanic basalts (below) (Pearce and others, 1975). *B*, FeO*-MgO-Al_2O_3 diagram. Fields, A, ocean island; B, continental; C, spreading center island; D, orogenic; E, ocean ridge and floor (Pearce and others, 1977). *C*, Alkali-silica diagram. Line separates tholeiites below from alkali basalts above (Macdonald and Katsura, 1964). *D*, Ti-Zr diagram. Ocean-floor basalts plot in fields D and B, low-potassium tholeiites in fields A and B, and calc-alkali basalts in fields C and B (Pearce and Cann, 1973).

long. Biotite is about three times as abundant as muscovite. Oligoclase and quartz are 1–2 mm in diameter. Oligoclase is about An 30 where it is not sericitized. Zircon and opaque mineral are accessories.

PEGMATITES

Pegmatites are very abundant throughout the complex as mentioned in the descriptions of the major rock units, but they are generally more sparsely distributed in the quartz monzonite gneiss than in the units to the south. In the migmatites and the somewhat migmatitic-appearing more southern unit of quartz monzonite gneiss, pegmatites are very numerous; concordant ones with gradational contacts make up a significant proportion of the rock, but concordant and discordant pegmatites with sharp contacts are numerous. In the southern part of the area, sharply bounded concordant and discordant pegmatites dominate, and individual pegmatite bodies tend to be larger (as much as 200 m thick and 700 m long; see pl. 1).

The pegmatites contain various proportions of quartz, plagioclase, and microcline and, locally, contain biotite books as much as 10 cm in diameter. In the migmatite unit and, less commonly, in the gneiss and schist unit, the pegmatites contain garnets as much as 2 cm in diameter where garnet-rich schist or gneiss is the country rock. Also, some of them contain hornblende as much as 10 cm long where the country rock is hornblende-quartz-feldspar gneiss. Some hornblende-bearing pegmatites are well exposed at Devils Gate. Muscovite, usually ruled, is more widespread in pegmatites in and near the quartzite unit in the southern part of the area, and some crystals are as much as 4 cm in diameter. Overall grain size of the pegmatites ranges from a few millimeters to several centimeters.

Foliation in the pegmatites ranges from well developed to undetectable. Biotite books, several centimeters in diameter, are broken down into aggregates of individual biotite grains, which are each a few millimeters in diameter. Many pegmatites lack biotite, and in them foliation is not obvious. Although I did no petrographic study of the pegmatites, I judge on the basis of their foliation that at least some of them were emplaced synkinematically or pre-kinematically. The pegmatites were variously affected by subsequent cataclastic metamorphism in the shear zones discussed in the next section.

CATACLASTIC ROCKS

About 20 percent of the rocks of the Farmington Canyon Complex in the Wasatch Mountains have been noticeably sheared and retrogressively metamorphosed to some degree following the main phase of dynamothermal metamorphism. The effects of cataclasis and retrogressive metamorphism can be read into some of the earlier descriptions, such as those of Eardley and Hatch (1940), but Bell (1951, 1952) was the first to point out the retrograde character of some of the metamorphism. He believed that the cataclasis and retrogressive metamorphism was confined to narrow zones, which he interpreted as thrust faults. However, other areas of low-grade rock were believed to be progressively metamorphosed.

Zones of shearing and retrogressive metamorphism in Weber Canyon and to the north are a meter to a few tens of meters thick; one accessible thick zone is at Devils Gate in Weber Canyon. Other zones on the south wall of Weber Canyon are well exposed (fig. 19*B*) and they can be mapped there. To the south, especially in the central part of the complex between Weber Canyon and Farmington Canyon, zones of cataclasis and retrogressive metamorphism are a kilometer or more thick (fig. 19*A*). Thin, anastomosing phyllonite zones on the south side of Weber Canyon appear to grade along strike in a southward direction into great thicknesses of cataclastic rocks, which are shown by an overprint on plate 1. The rocks in those thick zones are well exposed in Hobbs Canyon, in Webb Canyon, the crest of the range east of Hobbs Canyon, and the gorge of Line Creek on the east side of the range.

The rocks of the shear zones are cataclastic gneiss, mylonitic gneiss, blastomylonitic gneiss, phyllonitic gneiss, chlorite-sericite phyllonite (fig. 19*C*), and chlorite phyllonite. Colors are generally greenish gray, light greenish gray, gray, and medium bluish gray. All gradations between slightly sheared gneiss and phyllonite, or mylonite, occur. The less sheared gneisses usually have altered mafic minerals, which give their foliation planes a dull-greenish sheen.

An excellent place to examine many of the rock types seen in zones of sheared rock in the Farmington Canyon Complex is at Devils Gate in Weber Canyon. The east-facing cliff beside the Weber River displays anastomosing zones of more strongly sheared rock separating lenses of less strongly sheared rock, and local discontinuities along the zones of stronger shear are obvious in certain lighting. At the margin of the shear zone, migmatitic gneiss is sheared and brecciated. Even the less sheared rock has shiny green foliation planes formed by chlorite derived by retrogressive alteration of biotite and hornblende. Rocks exposed in the shear zone at Devils Gate include

chlorite-epidote-mylonitic gneiss,
biotite-chlorite-epidote blastomylonite,
mylonite,
cataclastic gneiss,
sericite-chlorite and chlorite-sericite phyllonite,
blastomylonitic gneiss,
silicified blastomylonite, and
chlorite-mylonite phyllonite.

Some lenses and layers of light-greenish-gray quartzite in the Hobbs Canyon area resemble tectonic lumps of quartzite, but field relations indicate that they are derived from pegmatite and gneiss rich in quartz and feldspar. However, they must have been silicified to various degrees during the cataclastic metamorphism.

Cataclastic and retrogressively metamorphosed rocks are common outside the mapped zones of those rocks. Some of the contacts of the zones are quite abrupt, but most of them are gradational and difficult to place except in a diagrammatic fashion. Many of the narrow, unmapped shear zones do have sharp contacts.

The sheared rocks are as varied in thin section as they are megascopically (fig. 20). Many of the rocks are only partly granulated, and they have porphyroclasts of minerals of the older, high-grade metamorphism in a matrix of mortar or partly recrystallized mortar. Some are thoroughly metamorphosed, and all the feldspar is converted to sericite and all the mafic minerals to chlorite. In some, felsic minerals are just mechanically granulated, whereas mafic minerals are retrogressively recrystallized.

Mafic minerals are chloritized without showing any effects of cataclastic deformation in a few rocks in the

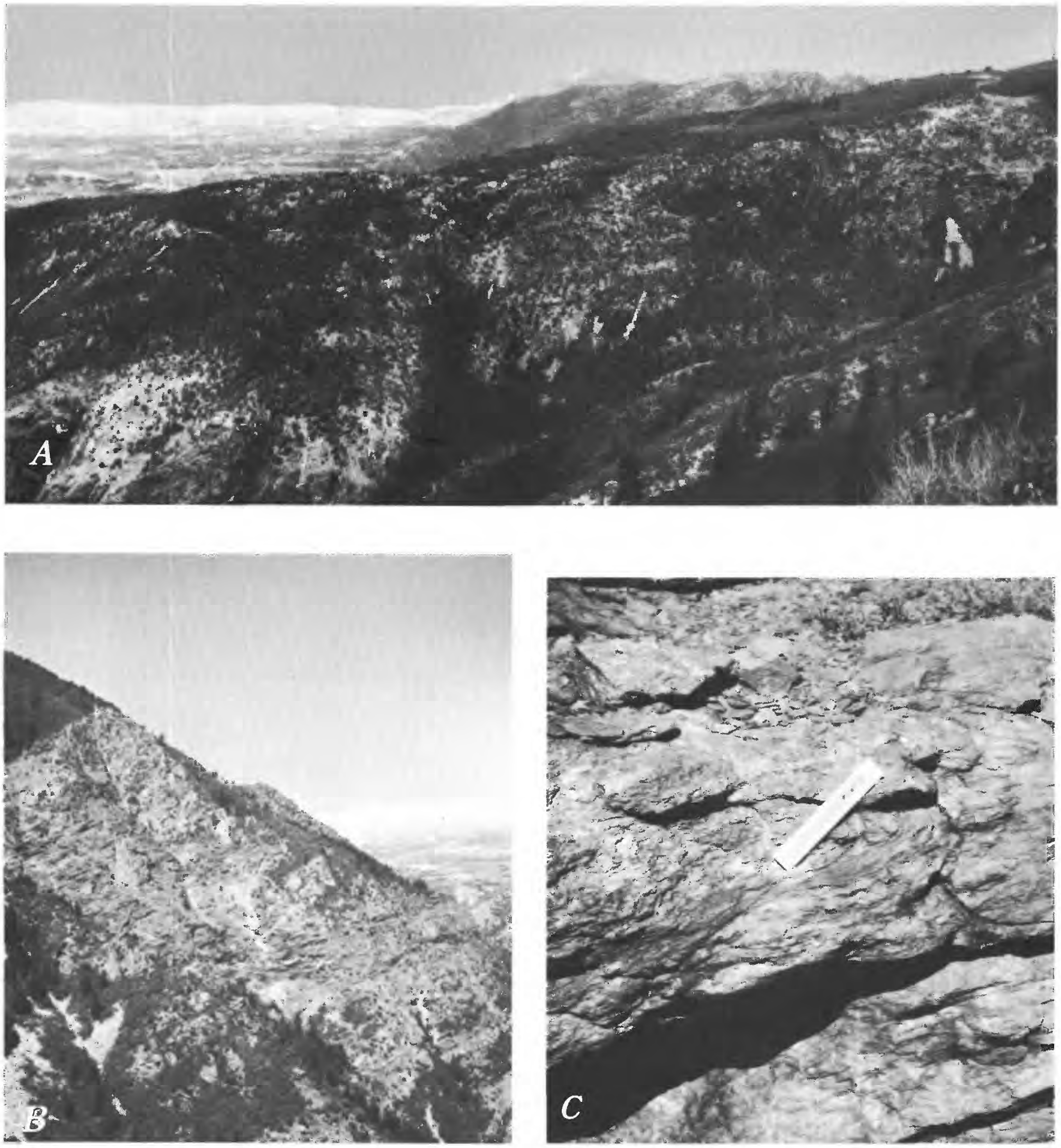


FIGURE 19.—Outcrops of cataclastic rock. *A*, Panorama of north side of Middle Fork of Kays Creek from 8,000 ft (2,440 m) altitude on ridge to south. Mixed retrograded and sheared gneiss and less sheared gneiss in area mapped as predominantly cataclastic rocks. Apparent dip slope of foliation and layering is not a true dip slope. Where measured on the ridge crest, the dips are variable, but the strikes are parallel with the trend of the valley side. The apparent dip here parallels the plunge of folds formed during the main metamorphism. *B*, View to west of south wall of Weber Canyon showing gently dipping shear zones cutting gently dipping foliation and layering in migmatitic gneiss. From 6,600 ft (2,010 m) altitude on ridge south of Devils Gate. *C*, Phyllonite derived from gneiss on 7,520 ft (2,290 m) nose north of Adams Canyon. Scale is 18 cm long.

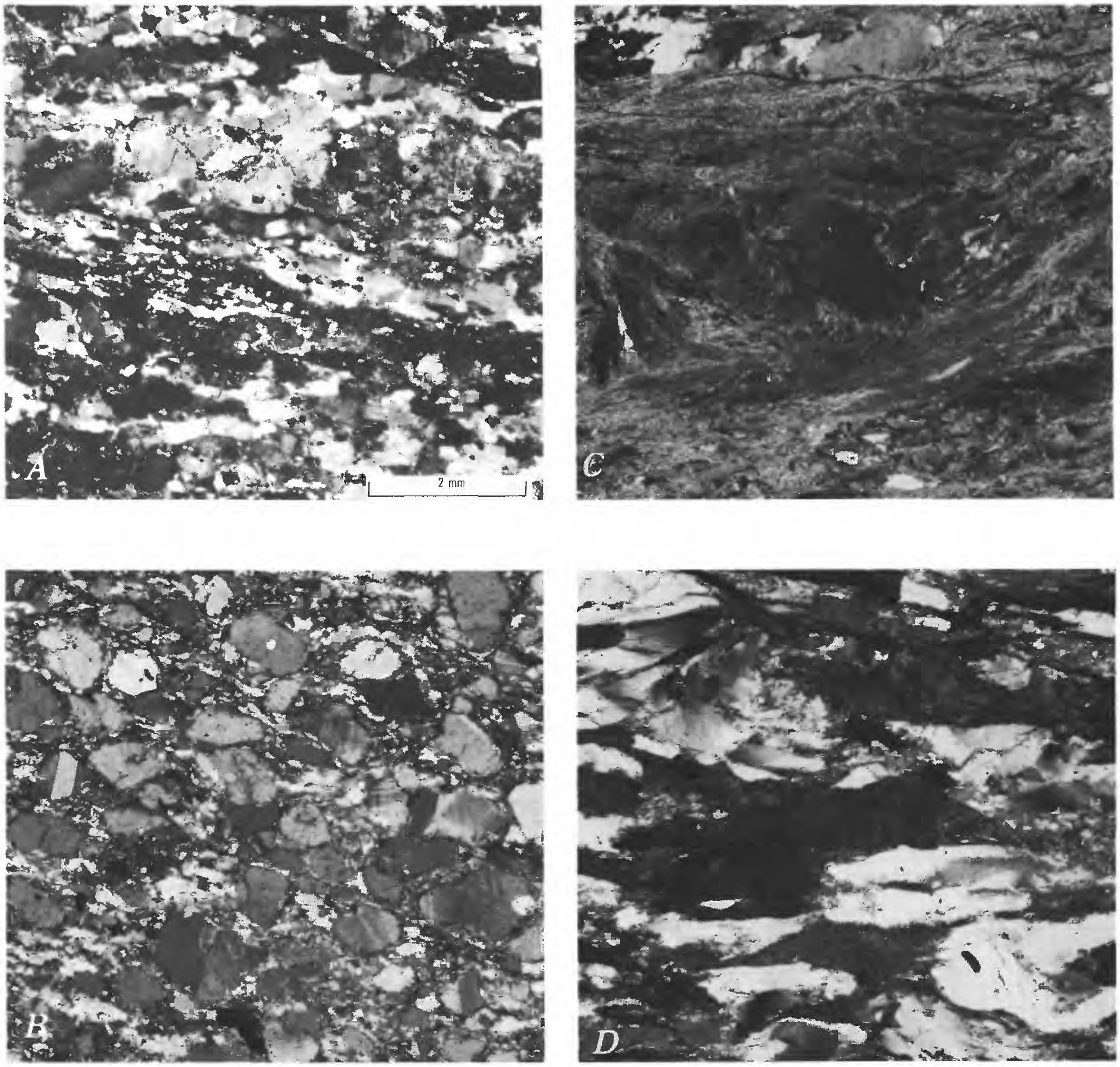


FIGURE 20.—Photomicrographs of cataclastic rocks in which shearing and retrogressive metamorphism have not destroyed evidence of their earlier history. *A*, Blastomylonitic garnet-bearing epidote-hornblende-biotite-plagioclase-microcline-quartz gneiss from 7,700 ft (2,350 m) altitude on ridge north of 8,000 ft (2,440 m) knob north of Corbett Creek. Porphyroclasts of microcline and plagioclase. Lenses of mosaic-textured quartz probably represent former large grains of quartz that were granulated by shearing and recrystallized. Biotite 0.1–0.3 mm long in matrix probably formed at that time. Rock cut by later shears not accompanied by recrystallization. *B*, Blastomylonitic granite gneiss from 6,440 ft (1,965 m) altitude on Deep Creek. Rounded porphyroclasts of microcline in a matrix of fine-grained mosaic-textured quartz, biotite, muscovite, and minor plagioclase. Zones of mortar indicate later shearing without recrystallization. *C*, Porphyroclastic phyllonite from outward corner of road northeast of Farmington Guard Station. Large porphyroclast of biotite partly altered to chlorite in a matrix of chlorite and sericite. Quartz concentrated in lens. Nearby schists contain sillimanite. *D*, Phyllonitic biotite-garnet-feldspar-quartz gneiss, from 5,760 ft (1,755 m) altitude on ridge south of Shepard Creek. Quartz is in elongate, strained, and brecciated grains and mortar. A few porphyroclasts of saussuritized plagioclase. Garnet in crushed and strung-out grains. (One such grain forms dark area in center of photo.) Fine-grained biotite, epidote, sericite, and chlorite formed contemporaneously with shearing. Rock contains a few porphyroclasts of biotite (not shown).

mapped sheer zones. Similar partly altered rocks are widespread in the southern part of the area, where relatively few sheer zones occur.

Effects of the variability in the interplay between cataclasis and recrystallization helps identify these sheared rocks as being the products of retrogressive metamorphism rather than progressive metamorphism, for relict minerals and structures resulting from earlier high-grade dynamothermal metamorphism can be found here and there in even the most severely sheared rocks.

Quartz is the most sensitive indicator of strain, and in over 70 percent of the samples studied in the entire Farmington Canyon Complex, the quartz shows well-defined undulatory extinction. In the cataclastic rocks, quartz is strongly strained, strung out, broken into smaller grains, or converted to mortar. In some rocks, large quartz grains were drawn out to double or triple their original length (fig. 19*D*). In other rocks, quartz is partly recrystallized into mosaic-textured aggregates.

Large plagioclase, biotite, or muscovite grains are commonly bent. Plagioclase is saussuritized or sericitized, and muscovite is sericitized. Microcline may also be bent and sericitized. Garnet is commonly fractured and altered to chlorite, sericite, epidote, and, locally, biotite along the fractures. In more highly sheared rocks, the garnet tends to get broken into fragments and strung out in the plane of the foliation (fig. 20D). Hornblende is converted to chlorite and epidote; locally, it is altered to actinolite. Sillimanite is especially sensitive to retrogressive metamorphism, and it is sericitized in many rocks including ones lacking cataclastic effects.

Neomineralization in the cataclastic rocks is dominated by the formation of sericite and chlorite. New albite or microcline is much less abundant. A fine-grained, biotite-appearing mineral is intergrown with chlorite in about 20 percent of the cataclastically metamorphosed rocks examined. This mineral is similar in aspect and occurrence to oxidized chlorite described by Chatterjee (1966). Epidote is locally abundant. It is concentrated in zones of late shearing in some rocks, it occupies later, crosscutting fractures in others, and in a few, it forms segregations with or without quartz.

Features in a few rocks sampled suggest the possibility of two stages in the shearing and retrogressive metamorphism. These rocks contain lenses of mosaic-textured quartz with a grain size of 0.1–0.2 mm. Synkinematic biotite, 0.1–0.2 mm long, may be contemporaneous with the crystallization of the mosaic-textured quartz. That quartz has been strongly strained, broken, and made into mortar, and the fine-grained biotite has been bent and mostly altered to chlorite during a later phase of the metamorphism (fig. 19*A* and *B*).

At least half the samples studied from the schist and gneiss and the migmatite units, not classified as cataclastic rocks, show significant cataclastic effects in thin section. However, only about 10 percent of the samples from the quartz monzonite gneiss unit studied show those effects.

METAMORPHIC HISTORY

The main metamorphic assemblage of the whole outcrop area of the Farmington Canyon Complex in the Wasatch Mountains is of amphibolite grade. Layers of sillimanite-garnet-biotite schist are found throughout the major rock units, except the quartz-monzonite gneiss. Some of the schist layers contain microcline, but they generally lack muscovite north of Farmington Canyon. Between Farmington Canyon and Ward Canyon, rocks containing sillimanite and microcline have small quantities of muscovite which appears to have formed contemporaneously with the sillimanite and microcline. South of Ward Canyon, contemporaneous muscovite and sillimanite occur together, and microcline is absent in the sillimanite-bearing rocks I studied. A similar wide zone, in which pelitic rocks contain muscovite with sillimanite and potassic feldspar, has been described by Evans and Guidotti (1966), and Lundgren (1966). Evans and Guidotti explained the assemblage by assuming that during dehydration P_{H_2O} slowly increases from initial values less than total rock strength, under conditions of low permeability, and is buffered by the assemblage muscovite-sillimanite-potassic-feldspar-plagioclase-quartz and controlled by local values of P and T. Subsequent mineralogical studies by Cheney and Guidotti (1979) support this conclusion. Application of this explanation to the wide zone of muscovite-sillimanite-potassic-feldspar-facies rocks in the Farmington Canyon Complex is uncertain because of the probable major metamorphism of these rocks during the Archean, so that the Early Proterozoic metamorphism probably did not involve dehydration and could even have involved hydration. The isograd of Evans and Guidotti (1966) would correspond to the boundary near Ward Canyon between rocks to the south having sillimanite and muscovite but lacking microcline and those to the north containing all these minerals, whereas Lundgren's isograd (1966) would correspond to the boundary at Farmington Canyon between rocks to the south with contemporaneous sillimanite, microcline, and muscovite and those to the north in which muscovite is lacking in the main mineral assemblage.

The layered metamorphic rocks north of Farmington Canyon are migmatitic like those in the sillimanite-microcline zone in Connecticut (Lundgren, 1966). South of Farmington Canyon, the rocks just barely reached the temperature necessary for the beginning of anatexis at about 650°C and 3 or more kb, assuming $P_{H_2O} \approx P_{total}$ (Winkler, 1979, p. 247). This incipient melting produced liquids rich in feldspar and quartz, components that segregated, or were injected into, planes of weakness in the schist and gneiss where they crystallized to form pegmatites. North of Farmington Canyon a somewhat larger proportion of the rock was melted to produce the migmatitic gneiss and schist. Experiments show that after the inception of melting only a very small increase in temperature can lead to a great increase in the proportion of rock melted (Winkler, 1979, p. 310). Pegmatitic liquids were available over a considerable interval of time during the main metamorphism, for condordant pegmatites are locally folded and cut by discordant ones.

The occurrence of hastingsite, both in quartz monzonite gneiss and in some layers of migmatitic gneiss in the vicinity, and its textural relations, indicate that it crystallized during the metamorphism following emplacement of the quartz monzonite gneiss. Experimental study of hastingsite (Thomas, 1982) suggests that it crystallizes at relatively low-oxygen fugacities. Thomas points out that Fe:(Mg+Fe) ratio in the rock also governs the occurrence of hastingsite. All the analyzed rocks in the Farmington Canyon Complex containing hastingsite (table 1, table 2, samples 5-7) have a higher Fe:(Mg+Fe) ratio than rocks lacking that mineral. However, all the analyzed rocks have low Fe_2O_3:FeO ratios, suggesting that they were metamorphosed at low-oxygen fugacities.

Bell (1951, 1952) stated that the highest grade rocks in the Farmington Canyon Complex are in the granulite facies. In his detailed report (1951), he did not mention the presence of pyroxene of any kind in granulite-facies rocks, but in his later summary article (1952), he mentioned hypersthene without giving any details about its occurrence. I found little definitive evidence for granulite-facies rocks in the Farmington Canyon Complex. A few rocks do contain orthorhombic pyroxene, but it is not in equilibrium with the main mineral assemblage. Although the complex has some layers of rocks rich in garnet and lacking biotite, biotite is very widespread. Pyroxene, even monoclinic pyroxene, is rare; little was seen in the rocks of the Farmington Canyon Complex. Monoclinic pyroxene was found in one sample of calc-silicate rock from the schist and gneiss unit, in two amphibolites from the migmatite north of Farmington Canyon, and in several amphibolites included in the quartz monzonite gneiss. One sample of quartz monzonite gneiss contains monoclinic pyroxene, and two have clumps of secondary minerals probably representing altered pyroxene. Relict hypersthene was found in one sample of migmatitic gneiss adjacent to the quartz monzonite gneiss and in an inclusion of garnet-plagioclase-quartz rock and one of ultramafic rock in the quartz monzonite gneiss.

The sparse evidence for an earlier metamorphism under granulite-facies conditions suggests that the terrane may have attained granulite-facies grade during an early stage of the main metamorphism or, alternatively, was first metamorphosed at granulite-facies conditions and later remetamorphosed under amphibolite-facies conditions.

Shearing and retrogressive metamorphism took place under very different temperature and pressure conditions than the main metamorphism. Some rocks were sheared and recrystallized under conditions of the biotite zone, indicating temperatures of greater than 400°C. In a few rocks, mosaic-textured quartz aggregates have been cataclastically deformed, and fine-grained synkinematic biotite is altered to chlorite. This suggests that shearing and retrogressive metamorphism occurred more than once under slightly different temperatures. Much of the shearing and retrogressive metamorphism took place under conditions of the chlorite zone. A mineral resembling biotite is intergrown with, and intergrades with, chlorite in about 20 percent of the cataclastically metamorphosed rock; this mineral could be oxidized chlorite (Chatterjee, 1966) rather than biotite and, if so, would not be diagnostic of temperature conditions.

AGE OF THE ROCKS AND METAMORPHIC EVENTS

Rb-Sr and Sm-Nd studies of rocks of the Farmington Canyon Complex indicate that the layered metamorphic rocks are Archean (Hedge and others, 1983). Sm-Nd data from migmatites in Weber Canyon give model ages ranging from 2,740 m.y. for amphibolite to 3,430 m.y. for a gneiss layer. The age of the amphibolite of 2,740 m.y. may be the age of that rock, which was originally intrusive into the pile of stratified rocks. The older Sm-Nd ages could be inherited from detritus derived from older crust and incorporated into sediments in the stratified sequence.

Rb-Sr data on whole rock, from samples of migmatite in Weber Canyon and from gneiss and amphibolite in the gneiss and schist unit south of Bountiful Peak, do not define a simple line on a Rb-Sr isochron diagram; instead, they define a triangular area with the lower boundary much sharper than the upper boundary. The lower boundary corresponds to an age of 2,600 m.y. and the upper boundary to an age of 3,600 m.y. The favored explanation for this relationship is that a regional metamorphism equilibrated the Rb-Sr ratios in most, but not all, of the rocks 2,600–2,400 m.y. ago, and that the older apparent ages are inherited from the premetamorphic rocks.

Rb-Sr studies of whole rock samples and U-Pb studies of zircon from the quartz monzonite gneiss samples, from both above and below the Ogden thrust, gave ages of 1,790±20 m.y. for that unit (Hedge and others, 1983), which was emplaced during an early stage of the metamorphism that produced the main metamorphic mineral assemblage in the Farmington Canyon Complex. That this metamorphism did not seriously affect the Rb-Sr ratios in the migmatites is additional evidence that those ratios were set during an earlier metamorphism, rather than being inherited in sedimentary detritus from older crust. The initial ^{87}Sr:^{86}Sr ratio of 0.769 for the quartz monzonite gneiss is very high, indicating that it probably formed by partial melting of crustal rocks like those of the surrounding migmatite, gneiss, and schist.

U-Pb data for zircons from the layered metamorphic complex yield a spectrum of ^{206}Pb:^{207}Pb ages from 1,770 to 2,271 m.y. (Hedge and others, 1983). The oldest ages are from the gneiss and schist unit south of the area of migmatization. These data are interpreted to indicate that the zircons were severely to totally reset during the main metamorphism about 1,790 m.y. ago.

Other evidence for Early Proterozoic metamorphism in the region is found just above the Willard thrust in a sequence of quartzite, muscovite, and chlorite schist called the Facer Formation (fig. 1; Crittenden and Sorenson, 1985a). Rb-Sr age of muscovite from the schist is 1,660±50 m.y. (Crittenden and Sorensen, 1980). Hornblende from a metamorphosed diorite (interpreted by Crittenden and Sorensen (1980) as unmetamorphosed, but by me as metamorphosed based on their fig. 10) has a K-Ar age of 1,681±12 m.y. (Crittenden and Sorensen, 1980). These rocks of much lower metamorphic grade were derived from tens of kilometers west to northwest of the Farmington Canyon Complex, which tectonically underlies them.

Ages of minerals from the Farmington Canyon Complex reflect cooling after the metamorphism 1,790 m.y. ago and resetting to varying degrees by subsequent events. K-Ar ages of hornblende range from 1,364 to 1,700 m.y. (Hashad, 1964). A Rb-Sr age of 1,580 m.y. was obtained on muscovite from a pegmatite in rock above the Ogden thrust (Gilletti and Gast, 1961). K-Ar ages of biotite range from 1,079 m.y. below the Ogden thrust to 559–487 m.y. above the fault (Hashad, 1964), suggesting differing Phanerozoic thermal histories between the two tectonic units. Even greater contrasts exist between the biotite ages in the Wasatch and a single biotite age of 224 m.y. on Antelope Island to the west. These dates suggest that the biotite may have undergone partial argon loss during the Sevier orogeny of Mesozoic age, for the history of the region, as shown by Paleozoic sedimentary rocks, suggests no Paleozoic thermal or dynamothermal event, nor a time of uplift sufficiently great to bring deeply buried crustal rocks closer to the surface and set K-Ar ages of biotite.

The shearing and retrogressive metamorphism have usually been attributed to the Sevier orogeny (Bell, 1951; Bruhn and Beck, 1979; Hollet and others, 1978). In a few places where relative movement directions on the shear zones have been determined, upper rocks moved southeast in relation to underlying rocks (Bell, 1951; Hollet and others, 1978), a reasonable direction for movement during the Sevier orogeny. All the major areas of sheared rock are overlain by Tertiary or Quaternary sedimentary rocks. Critical relations of the shearing to the Paleozoic rocks that stratigraphically overlie the Farmington Canyon Complex are not exposed. However, southeast of Strawberry Peak, the Cambrian Tintic Quartzite overlies a few mylonite zones about 1 m thick. In the cirque north of De Moisy Peak, a 3- to 5-m-thick shear zone, composed of mylonite gneiss, is overlain by the Tintic. However, nearby, the Tintic has been offset about 75 m in map plan along a fracture parallel with the shear zone. These relations suggest renewed movement during the Phanerozoic along a shear zone of Precambrian age.

On Antelope Island, Larsen (1957) reported considerable shearing older than diamictite and dolomite of Late Proterozoic age. Reconnaissance by Bryant and Graff (1980) has confirmed this relation; a shear zone hundreds of meters thick is overlain by the younger Precambrian rocks. On the other hand, shearing along the Ogden thrust, a fault of Sevier age involving the Farmington Canyon Complex and Paleozoic rocks, is much less intense. The quartz monzonite gneiss is well fractured near the fault, but it is not ground down and reconstituted like the rock in many of the shear zones in the Farmington Canyon Complex. I suggest that most of the shearing and retrogressive metamorphism in the complex in the Wasatch Mountains is like that on Antelope Island and probably of Middle Proterozoic age. A logical time for this Precambrian deformation would be later than the main metamorphism 1,790 m.y. ago and before the deposition of rocks of the Big Cottonwood Formation and Uinta Mountain Group 900–1,000 m.y. ago (Bressler, 1981), for the Uinta Mountain and Big Cottonwood rocks show no signs of similar deformation or metamorphism. Some renewed movement may have occurred along the shear zones during the Sevier orogeny.

ORIGIN AND HISTORY

The protoliths of the rocks of the Farmington Canyon Complex were a layered sequence composed principally of arkosic sandstone, quartz sandstone, and shale, as Eardley and Hatch (1940) and Bell (1951) concluded. The gneisses generally contain more quartz than feldspar. Abundance of quartz and microcline in the gneisses and the presence of quartzite layers (fig. 8) indicate that the sediments were derived from continental crust. However, some feldspar-rich layers could have been felsic tuffs. Whether there are any flows of felsic volcanic rock is unknown. The whole terrane was intruded by small bodies of tholeiitic gabbro or basalt before the metamorphism 1,790 m.y. ago, and at least some of these bodies intruded during Archean time. The "ocean floor" characteristics of the mafic-igneous rocks may indicate that no older continental crust lay beneath the deposits that were metamorphosed to form the Farmington Canyon Complex.

The entire complex was highly deformed, in part migmatized, and metamorphosed to high grade, probably in the Late Archean and definitely in the Early Proterozoic metamorphism. Some of the migmatite, schist, and gneiss were melted and intruded and formed the quartz monzonite of the northern part of the complex.

The potassic-feldspar content of the gneiss and schists may have been increased by the conversion of muscovite to orthoclase in the rocks north of Ward Canyon. Lundgren (1966) pointed out that potassium feldspar forms from muscovite in the proportion of 7 percent feldspar for every 10 percent muscovite converted. However, this process does not seem sufficient to account for the compositional differences, because, although some schists in the southern part of the area are rich in muscovite, the gneisses in that area are not particularly so. If the numerous thick pegmatites in the gneiss and schist unit represent locally derived material, some potassic feldspar may have been subtracted from the gneisses by formation of the pegmatite. Alternatively, the composition of the protoliths for the metamorphic rocks may change in a north-south direction.

STRUCTURE

The structures in the Farmington Canyon Complex are the result of a long complex history, and they formed in a great range of environments ranging from the temperature and pressure of high-grade metamorphism to near-surface conditions of brittle fracturing. A brief study such as this one is insufficient for deciphering all the details of the structures and their origins, but the most obvious events can be demonstrated.

FOLDS

Lack of stratigraphic marker horizons precludes determination of large-scale folds and offsets on shear zones within the Farmington Canyon Complex. The only map unit that could possibly serve as a marker horizon is the schist, gneiss, and quartzite north of Mill Creek. No facing structures were found in the rocks of that area, but compositional layering is parallel to bedding, and those rocks appear to occur in a complex, doubly plunging antiform. This possible antiform would be the oldest structure detected in rocks of the Farmington Canyon Complex, and it might have formed during the Late Archean metamorphic event 2,600 m.y. ago. Another possible interpretation of the structural configuration of these rocks, based on a downplunge view of the map relations, is that they form a thick and complexly deformed lens in northwest-plunging structural sequence. If that interpretation were correct, the age of the structure need not be older than the metamorphism during which the present mineral assemblages, lineations, and minor folds formed.

The next oldest possible structure also has to be inferred by viewing the map down the plunge of the mineral lineation. If this method is valid in this complex area, the quartz monzonite gneiss in the Dry Canyon-Spring Creek area occurs in a westward-plunging isoclinal synform overturned to the north. Because I know the age of the quartz monzonite, which has deformed after its emplacement during the Early Proterozoic metamorphism, I infer that the synform is of that age. Both my data and those of Bell (1951) indicate that south of the Weber River another west-plunging synform crosses the range.

The oldest minor structures obvious in individual outcrops are generally tight to isoclinal folds that range to open folds, gentle undulations, and sharp drags with axes plunging gently to moderately west, although many variations occur owing to later differential rotation or tilting (figs. 8*C*, 10*A*). Some folds are best described as contorted. Axial planes are generally parallel with the foliation and layering. The range in the shapes of the folds suggests that the rocks deformed in a non-uniform, plastic manner. No unequivocal evidence for superposition by folds of different orientation was found in rocks outside the zones of cataclastic deformation. Nevertheless, it is possible that earlier, more attenuated folds may have been refolded and transposed by the deformation that formed these obvious folds. Petrographic evidence indicates that these folds formed during the metamorphism responsible for the main mineral assemblage now preserved in the rocks. Minerals, such as sillimanite and hornblende are aligned parallel with the axes of the minor folds. Biotite and, in the southern part of the area, muscovite form polygonal arcs around the noses of folds indicating crystallization of these minerals synchronous with folding. In some samples, some of the biotite parallel to compositional layering is bent, and in a few, some biotite is aligned with the axial plane of the fold, suggesting local lack of synchroneity between folding and crystallization.

Map distribution of foliation attitudes suggests larger scale folds trending parallel with the lineation (Bell, 1951; pl. 1); however, few of these folds have much lateral continuity, reinforcing the impression derived from individual outcrops of a complex fold pattern. An open anticline and syncline, trending N. 60° W. and plunging N. 15°–20° through Bountiful Peak, is reported by Eardley and Hatch (1940). More detailed data indicate that a number of folds of differing tightness have that trend through the ridge at Bountiful Peak and vicinity.

SHEAR ZONES

The rocks of the Farmington Canyon Complex have undergone significant shearing and retrogressive metamorphism under conditions of the greenschist facies, as described in the section on cataclastic rocks (p. 31). Individual shear zones were mapped as thrust faults by Bell (1951) in the area between Weber and Farmington Canyons. In my work, I was not able to trace these zones satisfactorily because of the dense brush and because they merge with great thicknesses of cataclastic rocks. Some of the uncertainties in the mapping of these zones are suggested by Bell (1951).

In many places, the later shearing cuts the older foliation, especially where the foliation dips steeply. These relations are best developed in the large areas of sheared and retrogressively metamorphosed rock along the mountain front between Farmington Creek and the Weber River. In gross aspect, the sheared rocks appear to form a gentle, west-dipping mass (fig. 20*A*), but in detail their configuration is more complex, for they also form areas of sheared rock that cut east through the range. In many places, the shearing was controlled by the preexisting layering. Lineation formed by stretched minerals, aggregates of minerals, or slickensides on the shear planes is commonly oriented close to the older lineation in rocks of high metamorphic grade.

I found no certain way to tell one phyllonite zone from another. In Weber Canyon, exposures are good enough so that individual phyllonite zones can be mapped, especially on the south wall of the canyon west of Devils Gate. East of Devils Gate, the zones are cut by high-angle faults, some of which may have substantial Neogene displacement, based on apatite fission-track dating (Naeser and others, 1983). Correlating the closest phyllonite zones on either side of these faults may be an erroneous procedure, and to map the zones would require detailed fieldwork and considerable interpretation.

No offsets in metamorphic grade or mapped rock units were detected across zones of sheared rocks, but there are local sharp contacts between the retrogressively metamorphosed rocks of the greenschist facies and the high-grade rocks of the amphibolite facies where the shear zones have sharp boundaries.

STATISTICAL SUMMARIES OF MINOR STRUCTURES

Statistical plots of structures in the Farmington Canyon Complex (fig. 21) show some of the relations mentioned previously. Data collected from a small area form the simplest case. Column 1 is a summary of data from an area about 2 km square adjacent to the lower part of Farmington Canyon. There, poles to foliation form a fairly well developed girdle with an axis plunging 25° N. 70° W. (diagram 1*A*). This axis is close to a statistical concentration of mineral-lineation trends (diagram 1*B*) and coincides with a statistical concentration of the axes of minor folds (diagram 1*C*). Poles to foliation of cataclastic rock and phyllonite (diagram 1*D*) do not form a well-defined girdle; many of them lie on the girdle formed by the poles to foliation of the noncataclastic rocks, but they fall on a part of the girdle formed by gentler dips than those represented on the statistical high of the diagram for noncataclastic rocks (diagram 1*A*). Gentler dipping foliation planes, thus, were favored as loci for the later shearing. Mineral lineation in cataclastic rock (diagram 1*E*) is generally similar in trend to that in the noncataclastic rock (diagram 1*B*), but the distribution of the cataclastic lineations is somewhat more diffuse.

Data from larger areas (columns 2–4) form more diffuse patterns, but the same elements can be seen in them. The data from the Bountiful Peak quadrangle (column *B*) are most analogous to the data from lower Farmington Canyon. Statistical maxima of lineation (diagram 2*B*) and fold axes (diagram 2*C*) coincide with one another and with the statistical axis of a girdle formed by the poles to foliation in noncataclastic rocks (diagram 2*A*). Poles to cataclastic foliation form a diffuse pattern and show that the foliation generally has a gentle dip (diagram 2*D*). Lineation in the cataclastic rocks trends northwest and west, like that in the noncataclastic rocks, but has a somewhat more diffuse pattern (diagram 2*E*).

The diagrams for the Kaysville and Peterson quadrangles (column 3) are more diffuse than for the other areas, perhaps due to more rotation by late Tertiary high-angle faulting. However, the maxima of the fold axis and lineation diagrams coincide (diagrams 3*B* and *C*) and have a N. 70° W. trend. Foliation of the

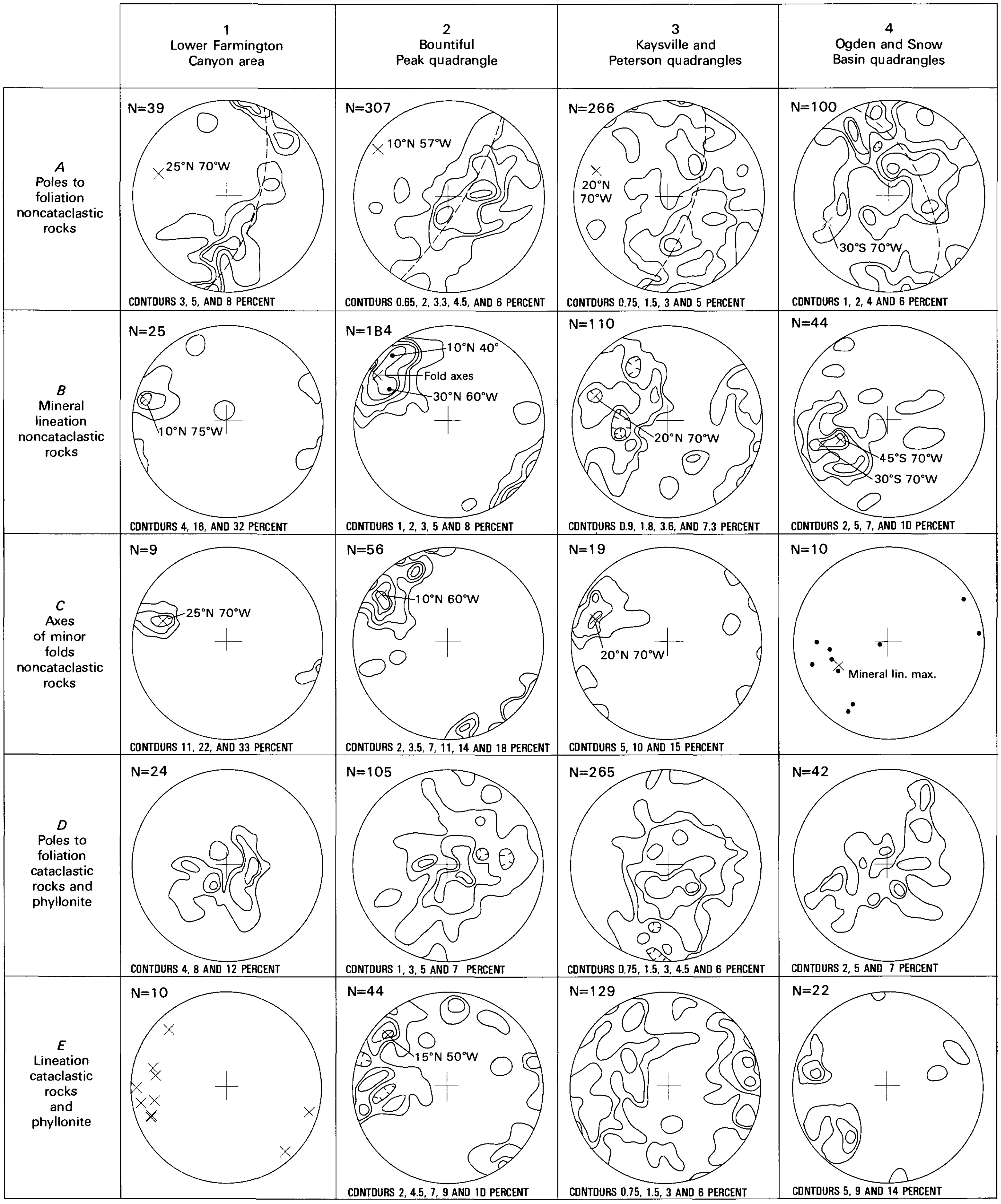


FIGURE 21.—Summary of minor structures measured in Farmington Canyon Complex. Equal area projections in lower hemisphere with planes of projection horizontal and north at the top. Contours show percentage of points falling within 1 percent of area of diagram. In column *A* X indicates the pole to the girdle shown by the dashed line through the contoured poles to foliation. In other diagrams the X is the statistical maximum of points contoured, unless otherwise labelled.

cataclastic rocks (diagram 3*D*) dips more gently overall than that in the noncataclastic rock (diagram 3*A*), but many of the attitudes in the cataclastic rock coincide with attitudes in noncataclastic rock. Lineation in the cataclastic rock (diagram 3*E*) has a range in trend, but has a statistical high in a westerly direction and various gentle to moderate plunges.

The diagrams for the Ogden and Snow Basin quadrangles (column 4) show a somewhat less diffuse pattern than those for the Kaysville and Peterson quadrangles (column 3). The mineral lineation maximum (diagram 4*B*) is parallel with a number of the axes of minor folds (diagram 4*C*) and is the axis of a possible girdle formed by the poles to foliation in the noncataclastic rocks (diagram 4*A*). The cataclastic rocks (diagram 4*D*) statistically have a gentler dip than the noncataclastic rock (diagram 4*A*); their strikes are variable. They do tend to dip gently north and west. Lineation in the cataclastic rocks (diagram 4*E*) forms two highs and trends west and southwest.

THRUST FAULTS

Some of the faults of the Ogden thrust zone pass south from the Paleozoic rocks in Ogden Canyon into the Farmington Canyon Complex. These thrust faults were recognized long ago by Blackwelder (1910) and Eardley (1944). Eardley mapped two separate folded thrust faults, which he called the Ogden and Taylor thrusts. South of Ogden Canyon, my mapping agrees with that of Temple (1969), who showed the Ogden thrust cutting an earlier east-trending fault (fig. 22C) and continuing south to the base of the range.

The main strand of the Ogden thrust zone, labeled the Ogden thrust on the map (pl. 1), cuts down into the Farmington Canyon Complex at the head of Warm Water Canyon and disappears beneath Quaternary rocks of the Salt Lake valley south of Strongs Canyon (fig. 22*A*). The fault cuts earlier, more deformed structures that appear to be related to the thrust fault system because of their spatial association with the thrust along strike to the north (Crittenden and Sorensen, 1985b) and their involvement with the same part of the stratigraphic section as the Ogden thrust. At its southernmost exposures, the Ogden thrust cuts down into the Farmington Canyon Complex rocks in the lower plate.

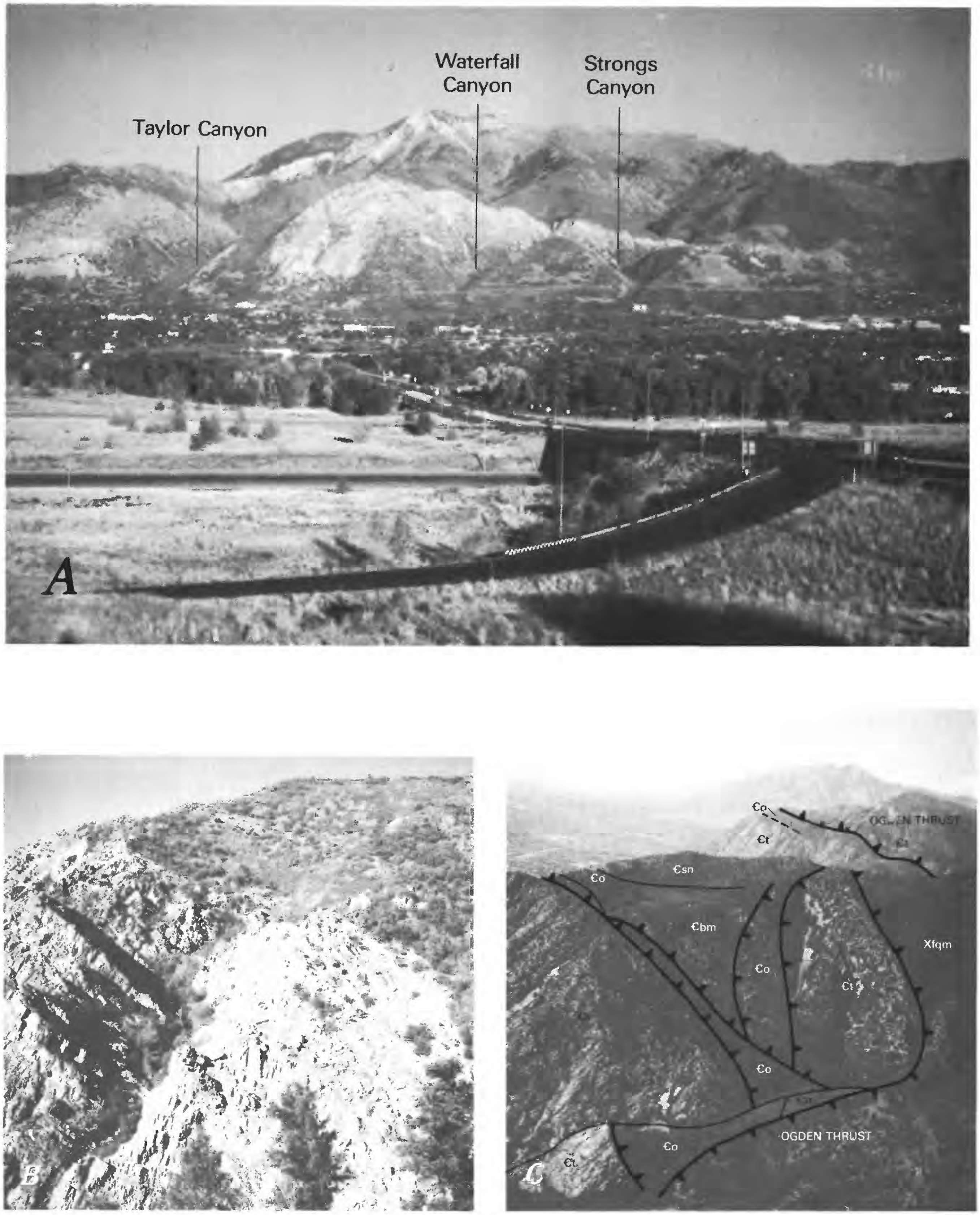
The Ogden thrust is well exposed on the ridge between Strongs Canyon and Waterfall Canyon. There, the overlying quartz monzonite gneiss is highly fractured and somewhat altered, but it is not pervasively sheared as are the rocks within the shear zones in the Farmington Canyon Complex. The fractured rather than sheared aspect of the quartz monzonite gneiss of the upper plate is characteristic of exposures near the Ogden thrust, such as those on the south side of Taylor Canyon (fig. 21*B*).

A thin section of quartz monzonite gneiss from the ridge between Waterfall and Strongs Canyons, from about a meter from the Ogden thrust, shows that quartz is strongly strained but not made into mortar (fig. 23). The quartz is only locally broken. Microcline is bent and locally broken. Biotite is bent and mostly altered to sericite and chlorite. Sericite occurs in fractures in grains and between the grains of the gneiss.

Below the fault on this ridge is a slice of the Lower and Middle Cambrian Tintic Quartzite, about 30 m thick, overlying about 6 m of limestone which could be from either the Middle Cambrian Maxfield Limestone or the Middle Cambrian Ophir Formation, another 6 m of Tintic, 6 m of limestone with phyllitic partings, probably from the Ophir, and then the main body of Tintic Quartzite, which stratigraphically overlies mica schist of the Farmington Canyon Complex.

Where the Ogden thrust approaches the floor of the Salt Lake valley in the northwest corner of sec. 11, T. 5 N., R. 1 W., the underlying Tintic Quartzite has a fault at its lower contact, as shown by very steep south dips where it is still sandwiched between the rocks of the Farmington Canyon Complex. South of where the Tintic is pinched out along the faults, the Ogden thrust abruptly changes strike and dip from north-south and gentle to the east, to east-west and steep to the south. The quartz monzonite gneiss is intensely fractured, but mylonite or phyllonite was not found. Similar abrupt variations in attitude of the thrust also were detected by careful tracing of the thrust on the south side of Taylor Canyon.

FIGURE 22 (facing page).—Views of the Ogden thrust. *A*, View of Ogden thrust where it passes southward into the Farmington Canyon Complex. West face of Wasatch Mountains from west side of Ogden. Left to right are Taylor, Waterfall, and Strongs Canyons. Bands of light-colored rocks are outcrops of Cambrian Tintic quartzite. The Tintic forms Mount Ogden in center of photo and is stratigraphically underlain by darker colored rocks of the Farmington Canyon Complex. Farmington Canyon rocks lie above lower, light-colored Tintic outcrop belt along Ogden thrust. On right side of photo, Tintic is cut out along Ogden thrust. At left, in Taylor Canyon, the Tintic is in an anticline and overlain by highly deformed rocks of the Ophir and Maxfield Formations. A fault in Taylor Canyon puts Farmington Canyon rocks against Maxfield Limestone and is cut by the Ogden thrust higher in the canyon. *B*, Detail of Ogden thrust on north side of Waterfall Canyon. Fault is irregular in detail with small slices of Tintic and Ophir along it above main body of the Tintic. Farmington Canyon Complex above thrust is highly fractured. *C*, View of Ogden thrust zone from 7,710 ft (2,350 m) altitude on ridge south of Taylor Canyon. Across valley, main thrust dips steeply, but changes to a gentler attitude in foreground. Xfqn, Farmington Canyon Complex; Ꞓt, Tintic quartzite; Ꞓo, Ophir Formation; Ꞓm, Maxfield Limestone; Ꞓbm, Bloomington Formation and Maxfield Limestone; Ꞓsn, St. Charles Formation and Nounan Dolomite.


Taylor Canyon
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A
C
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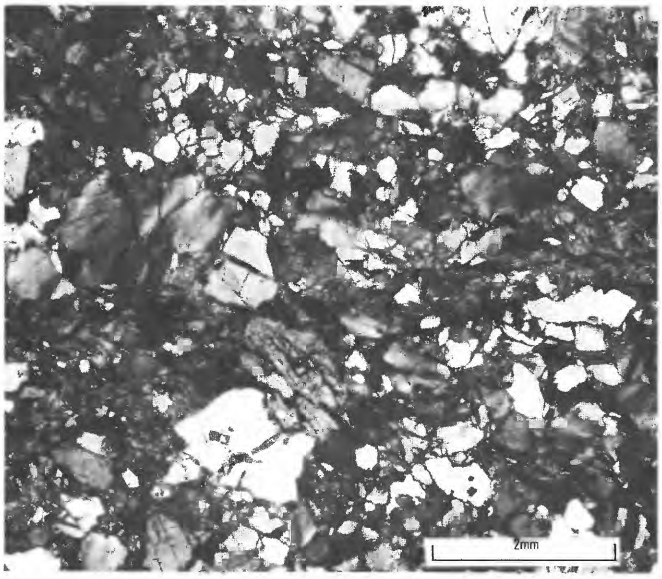


FIGURE 23.—Photomicrograph of somewhat cataclastic quartz monzonite gneiss sampled 1 m from Ogden thrust on ridge between Waterfall Canyon and Strongs Canyon. Quartz, plagioclase, and microcline are crushed somewhat, but are not milled down. Biotite is bent and mostly altered to sericite. Sericite occurs in fractures and between quartz and feldspar grains. Compare with rock from cataclastic zones (fig. 20).

At the head of Warm Water Creek, slices of limestone, either from the Maxfield or the Ophir, occur between the Tintic above and the Middle and Upper Cambrian Nounan Dolomite below. North of Taylor Canyon, a larger slice of Tintic occurs along the Ogden thrust.

Bruhn and Beck (1981) in a detailed study in Taylor Canyon, have interpreted the fault down Taylor Canyon (the Taylor thrust of Eardley, 1944) as north dipping. I have previously interpreted it as an isoclinally folded thrust (Bryant, 1979), but, in this report, I adopt their interpretation of the fault in the bottom of Taylor Canyon. I connect the north-dipping thrust there with a south-dipping one just south of Ogden Canyon (Sorensen and Crittenden, 1972). That south-dipping thrust fault is exposed on a sharp ridge near the south edge of sec. 23, T. 6 N., R. 1 W. south of the point where the "O" in Ogden Canyon appears on the base of plate 1, and Ophir Formation in the upper plate is against quartz monzonite gneiss in the lower. A 3-m-thick slice of Tintic Quartzite occurs in the Ophir, adjacent to the fault. The quartz monzonite gneiss is altered and fractured below the fault. On the next nose east, in the very southeast corner of sec. 23, a slice of Tintic along the fault dips 15° to the south.

Because of variations in the thickness of the Ophir and its continuation along the fault south of Ogden Canyon, I interpret a thrust fault above the one in Taylor Canyon that is also folded into an east-trending syncline and cut by the Ogden thrust, as is the thrust in the bottom of Taylor Canyon (sec. *A-A'*, pl. 1).

The continuation of this strand of the Ogden thrust zone to the south in the subsurface is a matter of interpretation. It may die out, or it may be cut by an inferred thrust fault between the Wasatch Mountains and Antelope Island, as shown in figure 1.

Reconnaissance mapping in the Cold Water Canyon area indicates that strands of the Ogden thrust zone pass above the Farmington Canyon Complex. Tintic Quartzite is sliced into several lenses with differing dips, and faults separating these lenses trend southward into outcrop belts of deformed Ophir. Folds and cleavage are found in the scattered outcrops of the Ophir and the adjacent part of the Maxfield south to east of Strawberry Bowl.

On Hardscrabble Creek, in the Porterville quadrangle east of the main outcrop area of the Farmington Canyon Complex, schist and gneiss of the complex on the east are faulted against the Upper Devonian Pinyon Peak Limestone on the west. The adjacent Maxfield Limestone has been thinned to about 100 m by faulting. Late Tertiary high-angle faults are abundant in the area, but I think about 500 m of the contact between the Farmington Canyon and Pinyon Peak is a thrust unmodified by the later faulting. The steep dip of this fault and of the Paleozoic rocks on the west would seem to preclude the juxtaposition of those units by high-angle faulting. This fault is on trend with faults north of Morgan, which put Precambrian basement rocks, assigned to the Farmington Canyon Complex, over Cambrian rocks (Mullens and Laraway, 1973; R. J. Hite, written commun., 1977). Along strike south of Hardscrabble Creek is a zone of complex structure in the Fort Douglas quadrangle (Van Horn and Crittenden, 1988) that may connect with the Mount Raymond and Absaroka thrusts (M. D. Crittenden, written commun., 1977). An alternative interpretation favored now by me and based on analysis of the structure of the thrust belt to the east and seismic data (Royse and others, 1975) is that the Odgen fault system is cut by the Absaroka and Mount Raymond thrusts, which must pass beneath the exposures of the Farmington Canyon Complex in the Wasatch Mountains. The faults north of Morgan may in some complicated manner connect beneath Morgan and Ogden valleys with the branches of the Ogden thrust that pass above the Farmington Canyon Complex in the Wasatch Mountains. The presence of Cambrian limestone and dolomite in the southeast corner of Ogden valley and their trends immediately below the Willard thrust suggests structural complexity in the rocks that underlie the valley. An outcrop area of very fractured Tintic Quartzite in secs. 8 and 17,

T. 5 N., R. 2 E., suggests similar complexity beneath the valley and the possible continuity of fault zones exposed along the mountains on either side.

HIGH-ANGLE FAULTS

The youngest structures in the Farmington Canyon Complex are many high-angle fractures and faults, most of them associated with late Tertiary and Quaternary uplift of the Wasatch Range. My reconnaissance mapping confirms and extends the general north-south pattern Bell (1951) obtained and adds a few east-west- and northwest-trending faults. The rocks of the Farmington Canyon Complex are highly fractured and, no doubt, more faults exist than are shown on the map, especially in areas of poor exposure. However, the lessening of the intensity of faulting at both the north and south ends of the area is confirmed by the continuity of the Paleozoic rocks covering the Farmington Canyon Complex in both places.

This stretch of the Wasatch Range is a complex horst with greatest displacement on the Wasatch fault along its west side. The zone of latest movement on the Wasatch fault, which has had as much as 11 m of Holocene displacement at a site between Baer and Shepard Creeks (Swan and others, 1980), lies a few meters to several hundred meters west of the most obvious topographic break formed by the wave-cut bench and cliff at the position of the high stand of Lake Bonneville at an altitude of about 1,585 m (5,200 ft). These Holocene faults also mark the western limit of exposures of the Farmington Canyon Complex in gullies eroded through the Lake Bonneville sediments west of and below the shoreline of the high stand.

Between Hobbs Canyon and Baer Creek, a number of north-south-trending fracture or fault zones were mapped parallel with the trend of the Wasatch fault. Displacement on them is unknown (fig. 24*B*).

The east side of the range is bounded by a complex system of high-angle faults separating blocks of Precambrian metamorphic rock overlain by the Paleocene and Eocene Wasatch Formation. The Wasatch in these blocks dips east at angles of 20°–40°, dips similar to those of the Wasatch on the west margin of Morgan Valley, just east of the zone of faulting. The faults have displacements as much as 800 m, with the west side down, and the blocks are not significantly rotated in relation to the Tertiary rocks along the west margin of Morgan Valley. (See secs. *C–C′*, *D–D′*, and *E–E′*, pl. 1.)

The principal fault of this system crosses the crest of the range at a low angle north of Francis Peak. To the south, it separates Precambrian rocks on Gold Ridge from the inlier of Wasatch at Farmington Flats.

FIGURE 24.—High-angle faults. *A*, View south along Francis Peak fault. Zone of crushed rock forms thick soil. Radar dome is on summit of Francis Peak. *B*, High-angle fault or fracture zone along west flank of range. View north from 8,120 ft (2,475 m) shoulder on ridge south of Adams Canyon. Fault zone marked by gullies, notches, and vegetation. Outcrops are migmatitic gneiss lacking many effects of later shearing and retrogressive metamorphism.

In the Francis Peak area, this fault, which forms a zone of crushed rock 200–300 m wide (fig. 24*A*), probably accounts for the "area of tan soil, as much as 100 ft thick" reported by Bell (1951) to occur on the crest of the range. The importance of the Francis Peak fault is shown by the discordance across it in the fission-track ages of apatite determined by Naeser and others (1983). About 2–3 km south of Francis Peak, apatite ages are 56.7 m.y. on the upthrown side and 74.3 m.y. on the downthrown side. At Francis Peak, the ages are 55 m.y. on the upthrown side and 63 m.y. on the downthrown side. Two kilometers north of Francis Peak, apatite ages are 17.7 m.y. on the upthrown side and 46.1 m.y. on the downthrown side. Apatite ages indicate that the rocks in this block are even more uplifted in secs. 8, 9, and 16, T. 4 N., R. 1 E., where apatite ages of 7–10 m.y. occur on the crest of the range; these ages are similar to those obtained in rocks just east of the Wasatch fault, some 1.5 km lower than the crest of the range.

Fission-track ages indicate that this strongly uplifted block within the range is terminated by east-trending faults south of Weber Canyon, for the apatite fission-track ages in the bottom of Weber Canyon are greater than on the crest of the range in that block. One of the faults in secs. 8 and 9, T. 4 N., R. 1 E., has a slice of Wasatch along it. That slice must have worked its way down the fracture progressively, for it is in rock in which the apatite fission-track ages are 7–8 m.y. Similar slices of sediment far below their level of deposition are found along fault zones in the Colorado Front Range (for example, Scott, 1963).

At the latitude of Weber Canyon, the high-angle faults and fractures swing to a northwest trend and die out. However, the main fault bounding the east side of the mountain block continues north and is en echelon with a fault system that preserves downfaulted portions of the Willard thrust sheet north of Wheeler Creek (pl. 1). This fault system dies out to the north, but another en echelon fault forms the east margin of the range north of the Ogden River (Sorensen and Crittenden, 1979).

Displacement along the fault system on the east side of the range is still substantial in the southern part of the area (See secs. *D-D'* and *E-E'*, pl. 1), but it diminishes to the south near the head of Hardscrabble Creek, 3–4 km south of the Porterville quadrangle.

On the east side of the range, in the southeastern part of the Peterson quadrangle and the eastern part of the Bountiful Peak quadrangle, the pattern of faulting is relative movement down on the west toward the range crest. This pattern may be explained if these faults dip west and are listric in relation to a possible basal decollement of the thrust belt 8 or 10 km below the surface, as suggested by Royse and others (1975). Hopkins and Bruhn (1983) have interpreted the fault bounding the Morgan valley on the east as a listric fault merging into that basal decollement. Based on this hypothesis, many of the faults on the east side of the Wasatch Mountains could dip west and also merge into that

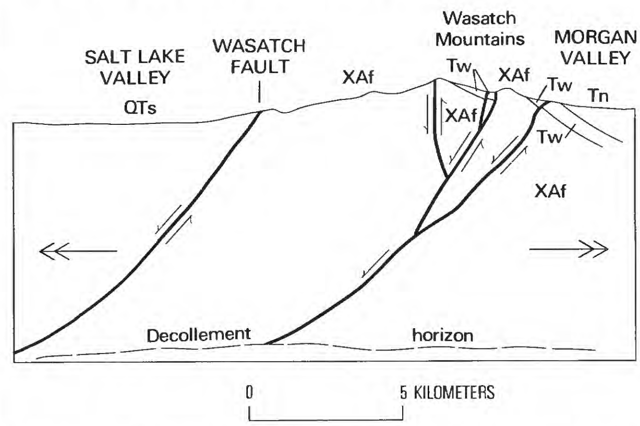


FIGURE 25.—Diagrammatic sketch of possible explanation of down-towards-the-range faulting on the east side of the Wasatch Mountains during extension of brittle crust above a decollement near the transition zone between brittle and ductile crust. XAf, Farmington Canyon Complex; Tw, Wasatch Formation; Tn, Norwood Tuff; QTs, Quaternary and Tertiary sedimentary rocks in the Salt Lake Valley. Single barbed arrows show direction of relative movement on faults. Double barbed arrows show direction of regional extension.

decollement, as shown in fig. 25. That decollement would form the boundary between brittle extension of the crust above and ductile extension below.

RELATION OF THE FARMINGTON CANYON COMPLEX IN THE WASATCH MOUNTAINS TO OTHER BASEMENT ROCKS OF NEARBY AREAS

ANTELOPE ISLAND

The second largest area of exposed basement rock in north-central Utah is on Antelope Island in Great Salt Lake, west of the Wasatch Mountains. In a detailed study of Antelope Island, Larsen (1957) concluded that the basement rocks there represent a stratigraphic section 6 km thick, composed entirely of clastic sediments metamorphosed to medium grade. Some of these sediments were dolomitic shales and were metamorphosed to amphibolite. Others were extraordinarily feldspar-rich sandstones, now quartz-feldspar schist. He divided the rocks into three units, which he treated as stratigraphic units: a lower unit of quartz-feldspar schist containing amphibolite beds; a middle unit of various rock types, such as microcline schist, mica schist, quartz-feldspar schist, and amphibolite; and an upper unit of quartz-feldspar schist, migmatite, and amphibolite. The upper and lower designations were based on physical superposition across foliation planes, rather than on any stratigraphic-facing criteria in the succession. Larsen recognized considerable cataclastic and retrogressive metamorphic effects in the rocks, and he attributed them to slippage during folding of the strata, for he believed that the main metamorphism took place under static conditions.

I participated in a brief reconnaissance of Antelope Island that led to a different interpretation of the geology (Bryant and Graff, 1980). The lower unit is a distinctive, uniform, nonlayered-granite gneiss that has sharp contacts with the middle unit, which is a belt of sheared rock greater than 1 km thick. Shearing lessens to the east, and the rock grades to the upper unit. The upper unit consists both of migmatitic schist and gneiss, which are at least in part of metasedimentary origin, and a body of granite gneiss like that composing the lower unit.

The granite gneiss is medium-grained, grayish-orange-weathering, homogeneous gneiss with a few pegmatite stringers and lenses, a few greater than 20 cm thick (fig. 26*A*). Amphibolite occurs in the granite gneiss scattered sparsely as pods and discontinuous dikes tens of centimeters to a few meters thick. The grain size of the gneiss is 1–2 mm (fig. 27*A*). Mafic minerals compose 5–10 percent of the rock and consist of light-green monoclinic pyroxene, which is partly to completely altered to amphibole and a brown micaceous mineral, perhaps stilpnomelane. Potassic feldspar is somewhat perthitic microcline. Plagioclase is generally saussuritized, but some fresher grains have a composition in the sodic-oligoclase range. Weakly developed cataclastic effects and veinlets of secondary epidote are common. Apatite, zircon, allanite, and magnetite are common accessory minerals.

Similar, but not identical, rocks were cored in Amoco Production Company's Antelope Island, State of Utah, well about 8 km west of the island. Through the courtesy of Amoco Production Company and Vince Matthews, III, we obtained a piece of that core. It is a garnet-biotite-granite gneiss with a texture indicating that it was sheared and metamorphosed under somewhat lower grade conditions than the granite gneiss exposed on Antelope Island (fig. 27*C*).

Semiquantitative modal compositions of several samples are relatively uniform and qualify as granite, as shown on a Q-Pl-Kf diagram (fig. 28). It seems most reasonable to interpret these rocks as an orthogneiss rather than as a metasediment.

Chemical analyses (table 4) show that the granite gneiss differs from the quartz monzonite gneiss of the Farmington Canyon Complex on the Wasatch Mountains in having more SiO_2 and K_2O and less CaO and FeO. Samples 1 and 2 have normative corundum, whereas the samples of quartz monzonite gneiss have none. Both those samples have significant amounts of modal epidote, and sample 1 has both biotite and muscovite.

The few Pb-U data obtained from zircons indicate that the granite gneiss is about 2,020 m.y. old (Hedge and others, 1983).

The east contact of the granite gneiss is abrupt and the adjacent rock is dark-green-gray to gray phyllonitic gneiss and phyllonite that (fig. 26*C*) contrasts with adjacent light-colored granite gneiss. The first several hundred meters are intensely sheared (fig. 26*B*), but farther east the degree of shearing and retrogressive metamorphism is variable (fig. 27*D*) and the rock ranges from schist and gneiss of high metamorphic grade to phyllonite. The phyllonite contains new sericite and chlorite and, locally, biotite (fig. 27*B*). Porphyroclasts of muscovite and chloritized biotite occur in rocks derived from mica schist, and porphyroclasts of quartz and feldspar in rocks derived from quartz-feldspar gneiss. Partially altered, relict garnet occurs in some rocks.

The rest of the rocks east of and overlying the shear zone are migmatitic gneisses, including garnet-biotite-quartz-feldspar gneiss, granite gneiss, and amphibolite, that contain many concordant stringers and lenses of pegmatite. Some pegmatite pods are as much as a few tens of meters thick. Pegmatites are much more numerous than in the granite-gneiss unit. Some amphibolites are as much as 10 m thick and have a relict-igneous texture formed by areas of saussuritized plagioclase between amphibole grains.

Sillimanite occurs in one sample of quartz-biotite schist we collected, thus allowing a better estimate of the grade of the main metamorphism than Larsen was able to make. Because this rock contains neither muscovite nor microcline, what part of the sillimanite zone it was in is unknown.

The amphibolites found in both the granite gneiss and the migmatite are composed predominantly of hornblende and plagioclase, which is usually saussuritized. They form sharply bounded, concordant lenses and discontinuous discordant dikes (fig. 26*B*), are locally cut by small pegmatites, and are not associated with calc-silicate rocks or marble. Judging from their field relations, modal compositions, mineralogies, and textures, the amphibolites are best interpreted to be mafic igneous-intrusive rocks.

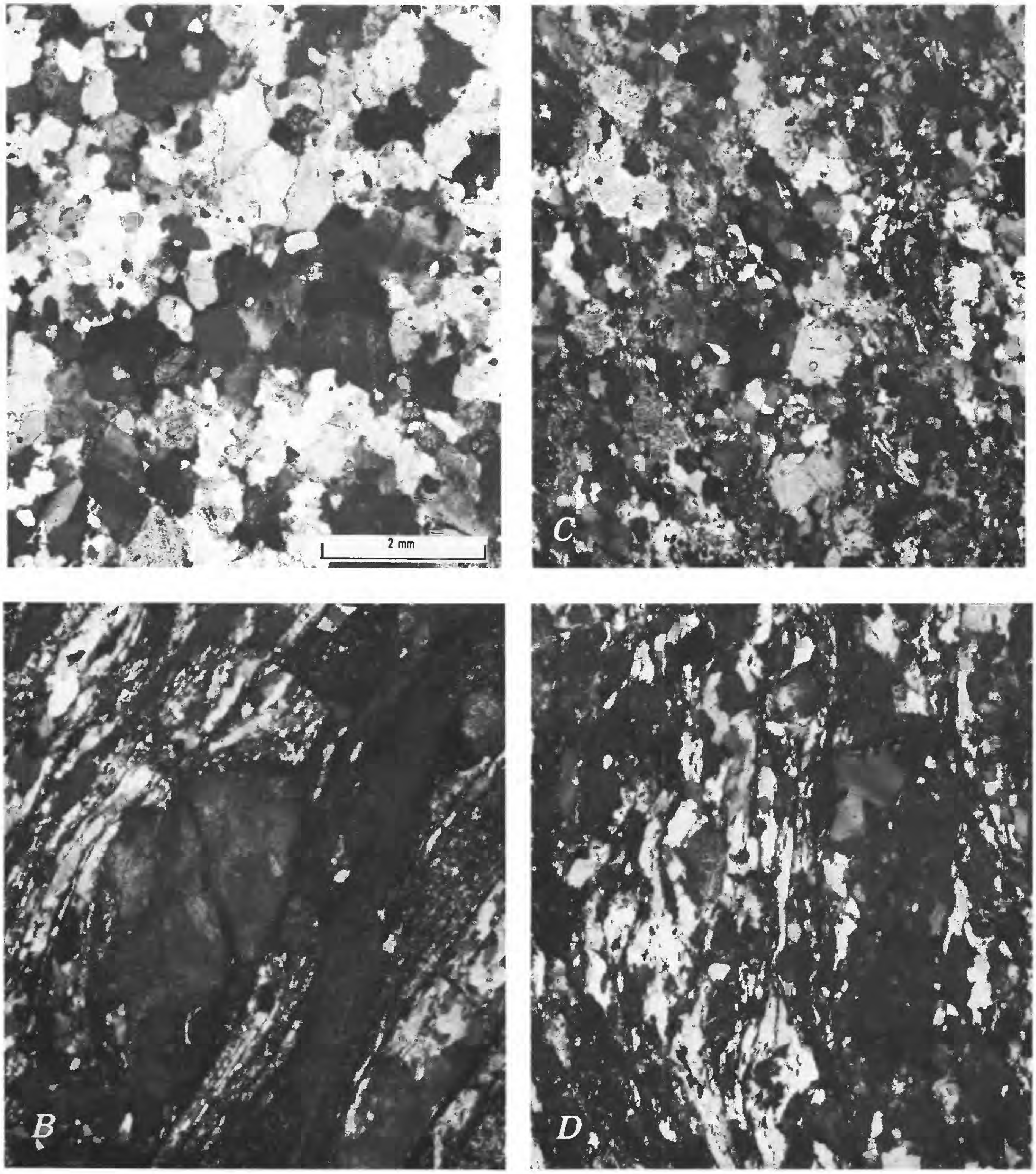
The cataclastic metamorphism, during which the rocks were milled down and minerals of the greenschist facies formed, must have taken place under significantly lower temperature and pressure than the main metamorphism. Steeply dipping rocks, deformed during this metamorphism, are overlain by almost flat-lying

FIGURE 26.—Outcrops of Farmington Canyon Complex on Antelope Island. Hammer is 31 cm long. *A*, Granite gneiss with pegmatite lenses, stringers, and pods. Just east of spot elevation 5,835 ft (1,778 m), sec. 31, T. 3 N., R. 3 W. *B*, Amphibolite pods in felsic gneiss south of Mollies Nipple. Such amphibolite pods are widespread in the Farmington Canyon Complex in the Wasatch Mountains. *C*, Phyllonite east of 4,635 ft (1,413 m) knob in sec. 7, T. 2 N., R. 3 W., in shear zone east of the main body of granite gneiss.

diamictite and dolomite of late Precambrian age, as Larsen (1957) observed. Therefore, this cataclastic and retrograde metamorphic event must have occurred in Precambrian time.

The rocks on Antelope Island probably originally constituted a sedimentary terrane, but it was intruded by diabase and by granite and then deformed and

FIGURE 27 (facing page).—Photomicrographs of rocks on Antelope Island and vicinity. *A*, Pyroxene-quartz monzonite gneiss from granite-gneiss unit from west side of knob, east of spot elevation 6,198 ft (1,889 m), west of center of sec. 31, T. 3 N., R. 3 W. Typical texture of granite-gneiss unit, but mafic minerals are less altered than in most samples examined. Plagioclase is sodic-oligoclase and albite. Potassic feldspar very fine textured perthite. Mafic minerals are monoclinic pyroxene and amphibole derived from it. Pyroxene is partly altered to brown micaceous mineral. Analyzed sample 3, table 5. *B*, Porphyroclastic chlorite-sericite phyllonite from main shear zone on 4,600 ft (1,402 m) knob east of 4,635 (1,413 m) spot elevation in center of east part of sec. 7, T. 2 N., R. 3 W. Porphyroclast of saussuritized plagioclase in a matrix of quartz-mortar, lenses of mosaic-textured quartz, and sericite and chlorite. From outcrop shown in figure 24*C*. *C*, Garnet-bearing biotite-muscovite-granite gneiss from Amoco, Antelope Island, State of Utah, well-core depth 10,391 ft (3,167 m), SW NW¼ sec. 19, T. 3 N., R. 4 W. Somewhat perthitic microcline and saussuritized plagioclase. Quartz in mosaic-textured aggregates. Texture suggests that the rock was partly sheared and recrystallized under medium-grade conditions. Analyzed sample 1, table 5. *D*, Phyllonitic quartz-plagioclase gneiss



from east slope of island north of summit 6,596 ft (2,002 m), west of center of sec. 29, T. 3 N., R. 3 W., about 0.5 km from west edge of shear zone. Saussuritized plagioclase and very strongly strained, strung out, and broken grains and mortar of quartz. Fractured and partly altered garnet present but not shown in this photomicrograph. Fine-grained biotite and chlorite formed during shearing.

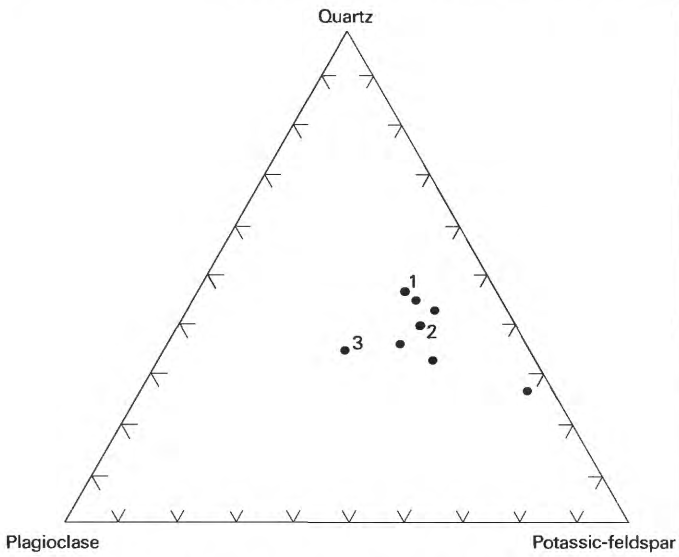


FIGURE 28.—Proportions of quartz, plagioclase, and potassic feldspar in samples of granite gneiss from Antelope Island and core from nearby drill hole. Numbers refer to analyzed samples in table 5.

metamorphosed under sillimanite-grade conditions. Later, in Precambrian time, shearing and retrogressive metamorphism occurred along zones a few meters to hundreds of meters thick under greenschist-facies conditions, a sequence of events similar to that which occurred in the Wasatch Mountains. However, the character and age of the granite gneiss have no counterpart in the Farmington Canyon Complex in the Wasatch Mountains, and I have no firm information on the age of the main metamorphism or the protoliths of the rocks on Antelope Island.

One K-Ar age of biotite of 224 m.y. from Antelope Island (Hashad, 1964) suggested that that mineral was more degassed in the Mesozoic than biotite in the Wasatch Mountains, which has K-Ar ages of 500–1,000 m.y. Perhaps a thrust fault emerging in the Salt Lake valley between the Wasatch Mountains and Antelope Island has telescoped these two parts of the basement complex, bringing rocks from the west, more severely heated during the Sevier orogeny of Mesozoic age, against less heated, more easterly ones.

AREAS TO EAST AND SOUTH

The Little Willow Formation (Crittenden and others, 1952; Neff, 1962), which crops out southeast of Salt Lake City (fig. 29), consists of quartz schist, quartzite, biotite-feldspar-quartz gneiss, and quartz-pebble schist. It contains concordant bodies of amphibolite as much as 100 m thick. Pegmatites are small and few. The gneisses and quartzites are much finer grained and appear to be of lower metamorphic grade than the Farmington Canyon Complex near Bountiful. Compositionally, the Little Willow resembles the Farmington Canyon Complex in the Mill Creek area, and it lacks the distinctive rock types of the Facer Formation (Crittenden and Sorensen, 1980). A thermal overprint from the nearby middle Tertiary Little Cottonwood stock (Hashad, 1964) makes it difficult to isotopically determine the age of these rocks and may be responsible for the post-kinematic micas and porphyroblasts of andalusite in pelitic rocks in the area.

East of the Wasatch Mountains in Cottonwood Creek (fig. 29), rocks mapped by R. J. Hite (written commun., 1977) as Farmington Canyon Complex consist of biotite-garnet-plagioclase gneiss and muscovite-biotite schist and gneiss of medium or high metamorphic grade and derived from sedimentary protoliths. No amphibolites have been found there.

At Santaquin (fig. 29) the basement rocks have the same stratigraphic relations as the Little Willow Formation, for they lie beneath a sequence of quartzite correlated with the Middle Proterozoic Big Cottonwood Formation. The Santaquin rocks are dominantly of metasedimentary origin and of medium and high metamorphic grade. Biotite-quartz-plagioclase gneiss, garnet-biotite-quartz-plagioclase schist, and biotite-microcline-augen gneiss are the main rock types; amphibolite, pegmatite, and a body of very light colored granite gneiss are also found. No definitive isotopic data are yet available from these rocks.

Age distribution of upper crustal rocks based on a study of ore leads (Stacey and others, 1968) suggests that the rocks at Santaquin are not correlative with the Farmington Canyon Complex. A sample of ore lead from the Mount Nebo Mine, south of Santaquin, fits the lead isotope data for the Oquirrh Mountains, and the study indicates that upper crustal rocks under the Oquirrh Mountains and in the Tintic region are of Proterozoic age (fig. 29). A possible interpretation of the 2,415 m.y. age, obtained from ore leads in the Cottonwood-Park City region in the same study, is that the area lies astride the margin of the Archean continent and that both Archean and Proterozoic materials form the basement there. If so, the rocks of the Little Willow Formation could be derived from very early Proterozoic sediments and deposited on the margin of the Archean continent rather than being derived from Archean sediments.

The apparent upper-crustal age based on ore leads in the Tintic region resembles the age obtained from granite gneiss on Antelope Island and vicinity (Hedge and others, 1983), perhaps indicating that additional bodies of that rock form the basement beneath the Tintic district.

TABLE 4.—*Chemical analyses, modes, and norms of granite gneiss from Antelope Island and vicinity*

[Major oxides determined by X-ray spectroscopy by J. S. Wahlberg, J. T. Taggart, and James Baker, U.S. Geological Survey 1980 and 1981. FeO determined by chemical means by Edythe Engleman, U.S. Geological Survey 1981 and 1982, and Fe_2O_3 calculated by difference with Fe_2TO_3. Ten minor elements determined by quantitative emission spectroscopy by Carol Gent, U.S. Geological Survey, 1981, and Paul Briggs, U.S. Geological Survey, 1983. Other minor elements determined by ICP direct reading semiquantitative emission spectroscopy by George Riddle, U.S. Geological Survey, 1981. Looked for but not found: Bi, Cd, U, Ge, Tb. Mode by count of 1,000 points. P, present but not intersected in counting; leaders (---), not present; N, not determined; <, less than. Major oxides and CIPW norm in weight percent; minor elements in parts per million; mode in volume percent. CIPW norms calculated on basis of analysis recalculated to 100 percent after deduction of volatiles]

Sample No.—	1	2	3
Lab No.------	D-227771	D-234340	D-234341
Field No.----	Am-1	A-11	A-5
Major oxides (percent)			
SiO_2	73.1	72.3	69.2
Al_2O_3	11.6	12.1	12.1
Fe_2O_3	2.65	3.40	3.90
FeO	1.69	1.46	2.34
MgO	0.1	0.21	0.42
CaO	0.97	1.20	1.91
Na_2O	2.1	2.65	3.15
K_2O	5.07	4.82	4.40
TiO_2	0.27	0.33	0.44
P_2O_5	<0.1	0.06	0.15
MnO	0.07	0.09	0.15
LOI	0.21	0.47	0.35
Quantitative minor elements (parts per million)			
Ba	1,410	N	1,490
Cr	2	N	<10
Nb	13	N	30
Ni	2	N	<10
Sr	52	N	80
Y	108	N	110
Zr	725	705	550
Ce	229	N	200
Nd	108	N	90
Rb	115	N	140

Sample No.—	1	2	3
Lab No.------	D-227771	D-234340	D-234341
Field No.----	Am-1	A-11	A-5
Semiquantitative minor elements (parts per million)			
Be	3.7	5	3
Co	<1	2	3
Cu	9	15	19
La	120	140	110
Mo	7	<5	<5
Pb	44	20	20
Sc	6	15	17
V	<3	6	5
Zn	95	150	140
Ga	19	20	10
Li	<3	8	<5
Ta	N	10	10
Th	34	31	24
Yb	11	16	10
Pr	30	<20	<20
Sm	20	20	<20
Eu	3	<5	<5
Gd	30	N	N
Dy	20	N	N
Ho	4	<10	<10
Er	13	N	N
Modes (percent)			
Quartz	41.0	36.4	32.2

TABLE 4.—*Chemical analyses, modes, and norms of granite gneiss from Antelope Island and vicinity*—Continued

Sample No.—	1	2	3
Lab No.------	D-227771	D-234340	D-234341
Field No.----	Am-1	A-11	A-5
Modes (percent)			
Plagioclase	13.8	15.1	29.9
Potassic feldspar	33.2	39.3	29.6
Biotite	1.9	--	--
Muscovite	4.7	--	--
Amphibole	--	--	1.4
Pyroxene	--	--	2.5
Garnet	0.3	--	--
Chlorite	0.1	0.6	--
Epidote	2.6	5.8	0.1
Stilpnomelane	--	--	0.8
Opaque	1.3	2.6	3.5
Allanite	0.8	0.1	P
Apatite	0.1	P	P
Sphene	0.1	--	--
Zircon	0.1	0.1	P
CIPW Norms (percent)			
q	39.93	36.22	30.58
C	0.92	0.49	0.00
or	30.69	26.88	26.49
ab	18.20	22.74	27.15
an	4.95	5.64	5.99
di-wo	0.00	0.00	1.11
di-en	0.00	0.00	0.66
di-fs	0.00	0.00	0.48
hy-en	0.26	0.53	0.41
hy-fs	0.61	0.00	0.24
ol-fo	0.00	0.00	0.00
ol-fa	0.00	0.00	0.00
mt	3.94	4.10	5.76
hm	0.00	0.62	0.00
il	0.53	0.64	0.85
ap	0.00	0.14	0.36

LOCATION AND DESCRIPTION OF SAMPLES

1. Core from 10,391 ft (3,167 m) depth in Amoco Production Co., Antelope Island State of Utah drill hole, SW¼NW¼, sec. 19, T. 3 N., R. 4 W., in Great Salt Lake. Blastomylonitic garnet-bearing biotite-muscovite granite gneiss. Anhedral porphyroclasts of somewhat perthitic microcline as much as 1.5 m in diameter and of saussuritized plagioclase (now albite) 1 mm in diameter in a matrix of mosaic textured quartz 0.1–0.3 mm in diameter. Muscovite and biotite in aggregates of anhedral grains as much as 0.7 m long. Aggregates are alined but the micas in them are not. Sample dated by Hedge and others, 1983.
2. Epidote granite gneiss from 4,430 ft (1,350 m) altitude on ridge NE center sec. 7, T. 2 N., R. 3 W. Anhedral, strongly strained quartz to 2 mm in diameter, saussuritized plagioclase (now albite) as much as 1 mm in diameter, and perthite potassium feldspar as much as 2 mm in diameter. Epidote accompanied by chlorite in aggregates as much as 2 m in diameter, in veinlets, and disseminated. A little mortar. Sample dated by Hedge and others, 1983.
3. Pyroxene quartz monzonite gneiss from ridge crest, 150 m N. 70° E. from spot elevation 6,198 ft (1,889 m) 370 m east of center of sec. 31, T. 3 N., R. 3 W. Anhedral quartz, plagioclase (An 8–17) and very fine textured perthitic potassium feldspar in grains as large as 2 mm. Anhedral to subhedral, light-green monoclinic pyroxene partly altered to a reddish-brown micaceous mineral (stilpnomelane?). Hornblende is from pyroxene.

The Farmington Canyon Complex lies at the margin of the Archean continent (fig. 29). The ancient margin was subject to metamorphic and thermal events during the Early Proterozoic, when new crust apparently formed south of it. Similar histories have been deciphered along this margin in southern Wyoming (Hills and Houston, 1979) and northeastern Utah (Sears and others, 1972), 400 and 250 km east, respectively, of the Farmington Canyon Complex.

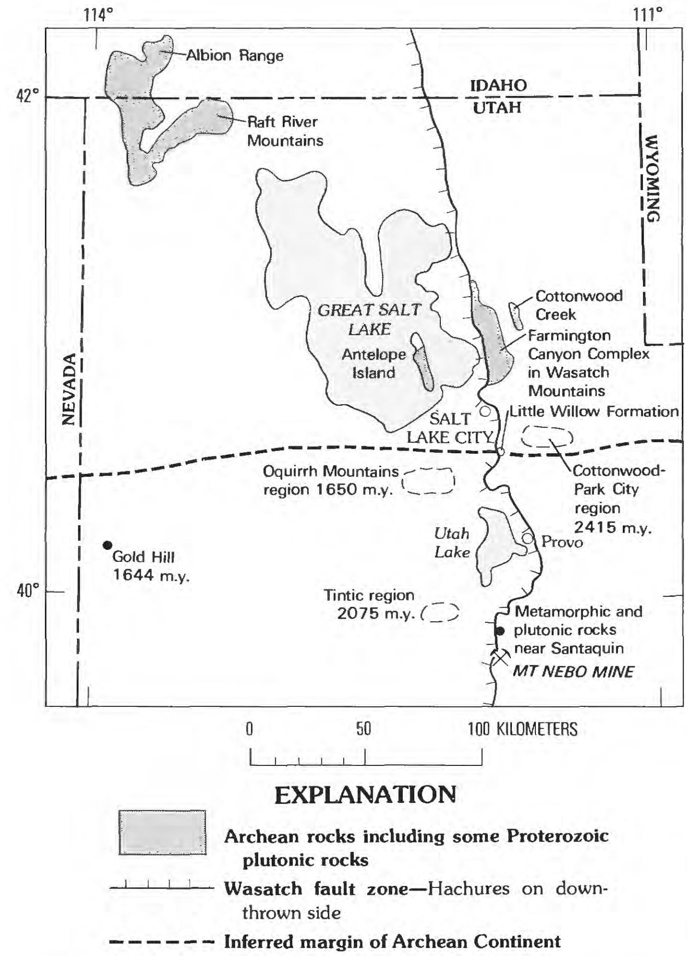


FIGURE 29.—Inferred margin of Archean continent in central and western Utah, showing areas of study and ages of crustal rocks based on interpretation of lead isotopic compositions (Stacey and others, 1968; Stacey and Zartman, 1978).

ECONOMIC GEOLOGY

Although no significant metal production is reported from rocks of the Farmington Canyon Complex in the Wasatch Mountains, they have been prospected extensively in some areas. The most numerous prospects, and some of the largest, are in Farmington Canyon. Other areas with several prospects are the north side of lower Hornet Creek valley, the north slope of lower Parrish Canyon, and the south slope of the Sessions Mountains.

Most of the prospects visited have little evidence of mineralization. Some have quartz veins a few centimeters thick, and some veins contain malachite and azurite. Some veins are in sheared rock and phyllonite, and rock adjacent to the vein contains some malachite. Where the veins are not weathered intensely, chalcopyrite is locally visible; some veins have been leached, and a limonitic boxwork has replaced the sulfides. Sphalerite and galena occur in quartz vein material on a large prospect dump north of Mill Creek.

Butler and others (1920) mentioned some of the prospects in the Farmington Canyon area near the mountain front. There, quartz-chalcopyrite veins lie along foliation in the metamorphic rocks, and a 20–25 ton shipment of copper-gold ore was said to have been made from one opening. Other prospects in the Farmington district have been reported to contain copper and a little gold. The annual reports of the U.S. Bureau of Mines mention a little mining activity in the Farmington Canyon area during the 1920's.

The principal vein described by Butler and others (1920) trends west on the north side of Parrish Creek. The prospects they visited are 3 km from the mouth of the canyon. Prospect dumps I saw from a distance in that canyon are about 1.5 km from the mountain front, and perhaps are in the same vein system. Near-surface material was said to average 10 percent copper and a little silver. The vein is 1–3 m (4–12 ft) thick and consists of chalcopyrite, pyrite, and specularite in a gangue of quartz and pyrite.

Analyses of malachite-bearing quartz veins (table 5, nos. 1–4) show that these selected samples contain as much as 1.4 percent Cu, but no gold or silver or significant amounts of other base metal. One vein (no. 3) has a high bismuth content. The selected sample from the galena and sphalerite-bearing veins (no. 5) contains 18 percent Pb, 1.7 percent Zn, and amounts of Cu and Cd above background as indicated by the content of these elements in the rocks analyzed (tables 1–3).

The veins are certainly of Precambrian age, for they do not cut the Paleozoic rocks overlying the Farmington Canyon Complex. Some of them cut phyllonite zones and perhaps formed during a late stage of the cataclastic metamorphism.

A prospect on Arthurs Fork is reported to show a trace of molybdenite associated with monazite, uraninite, and xenotime in biotite-rich layers and pods in biotite gneiss (U.S. Geological Survey, 1964). A few pegmatites have been prospected for muscovite.

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TABLE 5.—*Semiquantitative analyses of veins*

[Determined by ICP direct reading semiquantitative emission spectroscopy by George Riddle, U.S. Geological Survey, 1981. <, less than. Looked for but not found: Ag, Au, P, Be, U, Pr, Nd, Sm, Eu, Gd, Tb, Dy, Ho, Er]

Sample No.—	1	2	3	4	5
Lab No.----	D-234891	D-234892	D-234893	D-234894	D-234895
Field No.—	B-29	B-61	MB-25	F-22	F-23
			Percent		
Al	1.9	9.1	0.47	1.2	1.2
Fe	4.1	4.5	1.2	2.6	0.30
Mg	0.75	1.3	0.17	0.92	0.12
Ca	0.19	0.69	<0.02	0.08	0.04
Na	<0.14	7.5	0.04	0.03	13
K	<0.4	<0.4	<0.4	<0.4	1
Ti	0.12	0.21	<0.02	0.04	0.04
			Parts per million		
Ba	60	44	69	73	220
Bi	<100	<100	800	<100	<100
Cd	<10	<10	<10	<10	<40
Ce	<40	<40	<40	50	<40
Co	7	70	<4	10	7
Cr	50	100	10	50	90
Cu	14,000	14,000	3,800	4,900	220
Ga	<20	70	<20	<20	<20
La	<10	20	<10	20	<10
Li	<10	<10	<10	20	<10
Mn	170	420	60	160	40
Mo	10	10	<10	<10	<10
Nb	<20	80	<20	<20	<20
Ni	10	30	<10	<10	<10
Pb	<30	50	80	<30	180,000
Sc	<10	10	<10	<10	<10
Sn	<20	40	<20	<20	<20
Sr	20	60	<10	10	<10
Th	<10	30	<10	<10	<10
V	50	50	<10	30	20
Y	20	10	<10	<10	<10
Zn	50	70	20	50	17,000
Yb	<4	4	<4	<4	<4

LOCATION AND DESCRIPTION OF SAMPLES

1. Quartz vein containing malachite and sulfide in bluish-gray sheared rock. From prospect at 8,060 ft (2,557 m) altitude on ridge, 1.6 km N. 69° W. of Bountiful Peak, Bountiful Peak 7.5-minute quadrangle.
2. Biotite quartz feldspar gneiss cut by small quartz veinlet containing malachite. Prospect at 6,480 ft (1,975 m) altitude on ridge between Parrish Creek and Centerville Canyon, Bountiful Peak 7.5-minute quadrangle.
3. Quartz vein containing malachite. Prospect at 5,740 ft (1,450 m) altitude north of Hornet Creek, and 150 m S. 28° W. from center sec. 20, T. 3 N., R. 1 E., Bountiful Peak 7.5-minute quadrangle.
4. Sheared pegmatite gneiss and quartz veinlet containing pyrite and local malachite from dump of adit at 7,560 ft (2,304 m) altitude, 420 m S. 19° W. from NE corner sec. 35, T. 2 N., R. 1 E., Fort Douglas 7.5-minute quadrangle.
5. Quartz vein containing galena and sphalerite from dump of adit at 7,530 ft (2,295 m) altitude, 550 m S. 39° W. of NE corner sec. 35, T. 2 N., R. 1 E., Fort Douglas 7.5-minute quadrangle.

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