Geologic Map of the Warm Spring Canyon Area, Death Valley National Park, Inyo County, California

With a Discussion of the Regional Significance of the Stratigraphy and Structure

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Pamphlet to accompany

Scientific Investigations Map 2974, Version 1.1

View of lower Warm Spring Canyon, southern Panamint Range, Death Valley National Park, California. On right (south) side of canyon, dark diabase sill (age about 1 Ga), one of many in area, intrudes light-colored dolomite of the Mesoproterozoic Crystal Spring Formation, locally forming talc deposits (white patches) by contact metamorphism. South-dipping strata in left foreground are basal beds of the Crystal Spring Formation. View is east-southeast into Death Valley. In distance, Black Mountains are on left, and Owlshead Mountains are on right. Photograph by C.T. Wrucke, 1951.

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Introduction

Warm Spring Canyon is located in the southeastern part of the Panamint Range in east-central California, 54 km south of Death Valley National Park headquarters at Furnace Creek Ranch (fig. 1). For the relatively small size of the area mapped (57 km^2), an unusual variety of Proterozoic and Phanerozoic rocks is present. The outcrop distribution of these rocks largely resulted from movement on the east-west-striking, south-directed Butte Valley Thrust Fault of Jurassic age (Wrucke and others, 1995). The upper plate of the thrust fault comprises a basement of Paleoproterozoic schist and gneiss overlain by a thick sequence of Mesoproterozoic and Neoproterozoic rocks, the latter of which includes diamictite generally considered to be of glacial origin (Miller, 1987; Abolins and others, 2000). The lower plate is composed of Devonian to Permian marine formations overlain by Jurassic volcanic and sedimentary rocks. Late Jurassic or Early Cretaceous plutons intrude rocks of the area, and one pluton intrudes the Butte Valley Thrust Fault. Low-angle detachment faults of presumed Tertiary age underlie large masses of Neoproterozoic dolomite in parts of the area. Movement on these faults predated emplacement of middle Miocene volcanic rocks in deep, east-striking paleovalleys. Excellent exposures of all the rocks and structural features in the area result from sparse vegetation in the dry desert climate and from deep erosion along Warm Spring Canyon and its tributaries.

Access to the area is provided by an unpaved road that enters Warm Spring Canyon from Death Valley on the east, passes Warm Spring, and continues westward across the range to Panamint Valley (figs. 2, 3). Another unpaved road from Death Valley leads to the Queen of Sheba Mine in the northeast corner of the map area. A few kilometers west of Warm Spring, an abandoned road, usable as a trail, extends north from Warm Spring Canyon to abandoned talc mines (fig. 3). An unmapped trail to the Queen of Sheba Mine begins about 1 km east of Warm Spring and winds northeastward across a high ridge.

Most of the Warm Spring Canyon map area was mapped originally by Wrucke (1966). Later Stone (1984) mapped the westernmost part of the area, with emphasis on the upper Paleozoic rocks. This report is based primarily on those original studies but also incorporates much additional stratigraphic, structural, and paleontologic data acquired by the authors at various times through 2005. The map area lies directly east of an area mapped by Johnson (1957); directly south of an area mapped by Hunt and Mabey (1966); and also southeast of the Telescope Peak 15' quadrangle, which was mapped by Albee and others (1981).

Figures 4 and 5 summarize the stratigraphy of the Proterozoic, Paleozoic, Mesozoic, and Tertiary sedimentary and volcanic rocks of the Warm Spring Canyon map area.

Paleoproterozoic Rocks

The oldest rocks in the Warm Spring Canyon map area are schist, monzogranite gneiss, pegmatite, and aplite of Paleoproterozoic age. Biotite quartz-mica schist and associated metamorphic rocks, for brevity referred to here collectively as mica schist (unit Ᵽ$_p$ms), are the oldest and most extensive of these rocks. The rocks were metamorphosed mostly from sandy shale and argillaceous sandstone interbedded with lesser amounts of porphyritic volcanic flows and tuffs of mafic to silicic composition. The mica schist hosts Paleoproterozoic intrusive rocks, the largest of which is a pluton of monzogranite gneiss (unit Ᵽ$_p$mg). Conversion of the monzogranite to an augen gneiss is here interpreted to have been contemporaneous with the development of schistosity in the mica schist, as the

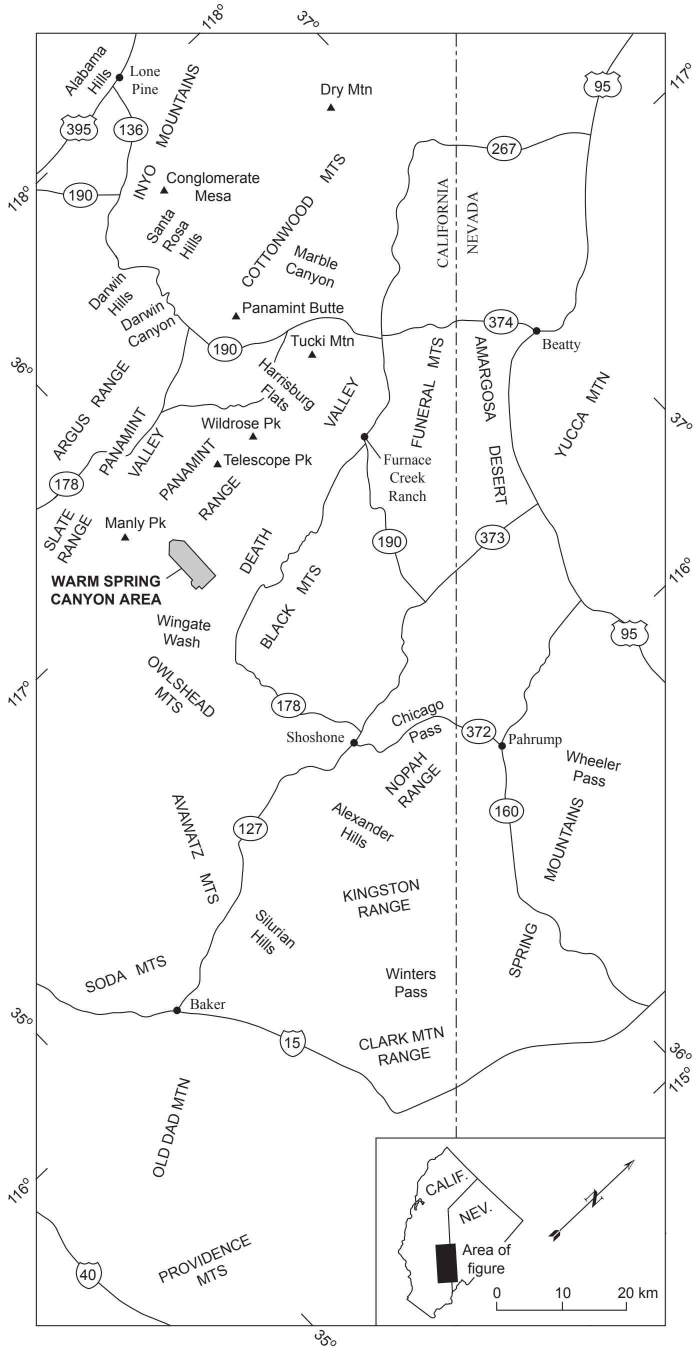

Figure 1. Map of region in eastern California and western Nevada, showing location of Warm Spring Canyon map area (shading) and other geographic features.

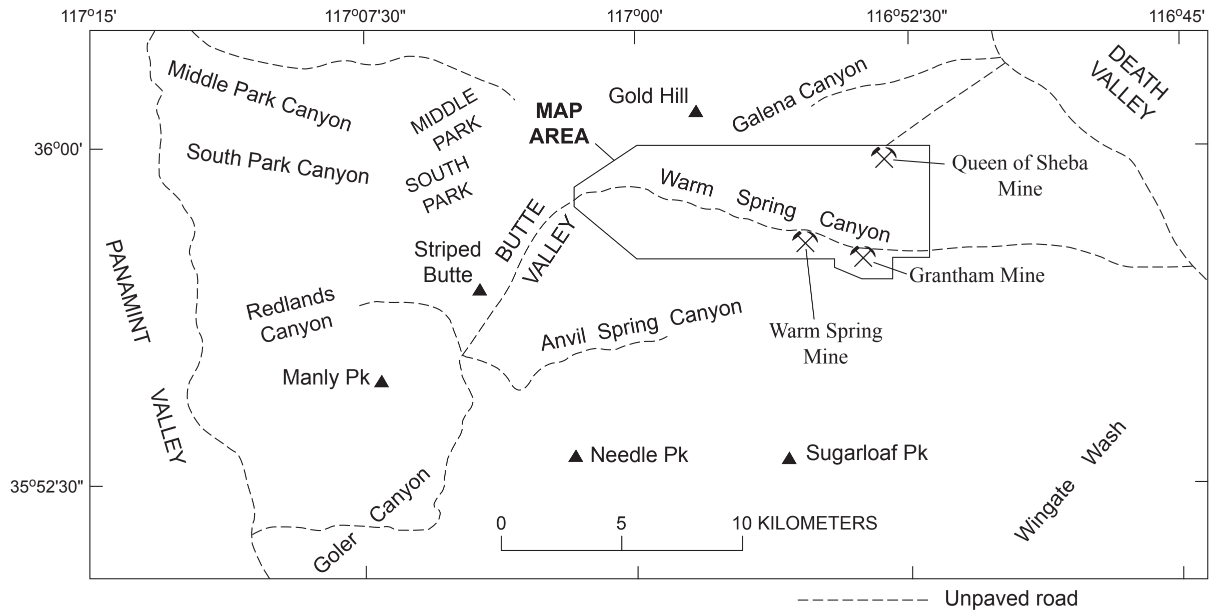

Figure 2. Map of southern Panamint Range, showing location of Warm Spring Canyon map area and other geographic features.

metamorphic fabric crosses their mutual contacts. Dikes of the monzogranite gneiss (Ᵽpmg) and of aplite and pegmatite (unit Ᵽppa) cut the mica schist north of the monzogranite pluton. The dikes were folded along with the host mica schist. Regional metamorphism of the original sedimentary and igneous rocks to mica schist and the monzogranite to gneiss produced a mineral assemblage of the quartz-albite-epidote-biotite subfacies of the greenschist facies. Green biotite in the schist suggests that the metamorphism probably was at the lower end of this subfacies.

Zircons from the monzogranite gneiss and from porphyritic metarhyolite and metarhyodacite in the mica schist of the Warm Spring Canyon area have yielded minimum $^{207}Pb/^{206}Pb$ ages of 1,780 ± 20 Ma and 1,720 ± 20 Ma, respectively (Silver and others, 1961). Rb-Sr analyses have provided ages of 1,700 to 1,570 Ma from pegmatite in the mica schist and 1,480 Ma from the mica schist host (Wasserburg and others, 1959). The discordance resulting from isotopic ages that are older for the crosscutting pegmatites than for the mica schist itself may have resulted from regional metamorphism of Mesozoic age (Labotka and others, 1980).

Outside the map area, zircons from the World Beater Complex, an assemblage of metamorphic basement rocks centered about 15 km northwest of Warm Spring, have yielded U-Pb ages of about 1,800 Ma (Lanphere and others, 1964; Labotka and others, 1980); these ages were later reported as about 1,700 Ma (Albee and others, 1981). K-Ar ages of 143 to 103 Ma for micas from a number of rocks in the World Beater Complex (Lanphere and others, 1964) indicate Mesozoic (probably Cretaceous) metamorphism.

A study of the Nd isotopic compositions of cratonic basement rocks in the southern Death Valley region, including mica schist from Galena Canyon 1.5 km north of the map area (fig. 2) and monzogranite gneiss 0.75 km northwest of Warm Spring, shows that these rocks contain as much as 30 to 40 percent Archean crustal material (Rämö and Calzia, 1998). Presumably this material was derived from Archean sedimentary detritus that in some cases has been recycled by tectonic and magmatic processes (Rämö and Calzia, 1998). Bennett and DePaolo (1987) showed that an Archean component has been found in Paleoproterozoic rocks in a wide area of southeastern California.

Mesoproterozoic and Neoproterozoic Rocks

Resting unconformably on the contorted Paleoproterozoic mica schist and augen gneiss are sedimentary rocks of the Pahrump Group, originally named the Pahrump Series by Hewett (1940) for rocks exposed in the Kingston Range (fig. 1). In the map area, the Pahrump Group consists of the Mesoproterozoic Crystal Spring Formation and the Neoproterozoic Kingston Peak Formation. A Mesoproterozoic diabase that intrudes the Crystal Spring Formation locally is overlain depositionally by the Kingston Peak Formation. The diabase is not considered to be part of the Pahrump Group. A third formation of the Pahrump Group, the Beck Spring Dolomite, which lies between the Crystal Spring and Kingston Peak Formations in other parts of the Death Valley region, including an area farther north in the Panamint Range (Hunt and Mabey, 1966; Albee and others, 1981), is absent in the Warm Spring Canyon area, presumably because of erosion prior to deposition of the Kingston Peak Formation.

The Pahrump Group has been interpreted as deposited in a northwest-striking intracratonic basin that Wright and others (1976) named the Amargosa aulacogen. More recent studies,

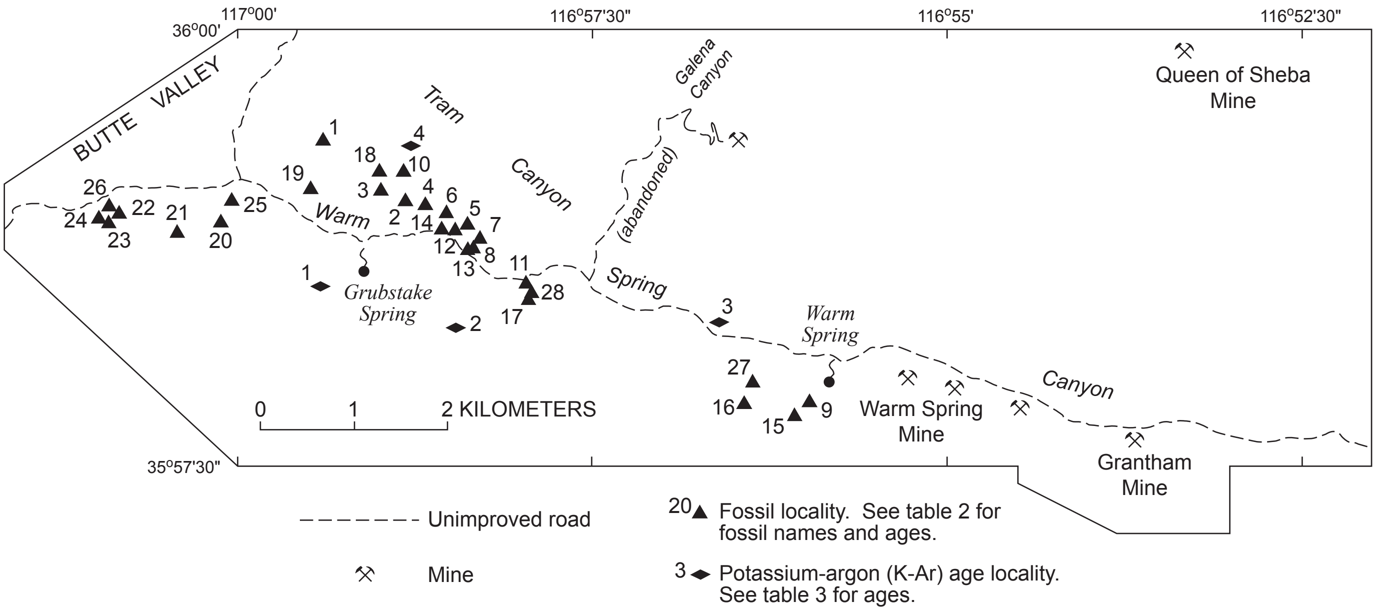

Figure 3. Location map of Warm Spring Canyon area, showing named geographic features, fossil localities, and potassium-argon (K-Ar) age localities.

however, suggest that the Crystal Spring Formation consists of continental-platform deposits (Heaman and Grotzinger, 1992; Timmons and others, 2001; Corsetti and others, 2002), although at least the upper parts of the Kingston Peak Formation accumulated in fault-bounded basins (Walker and others, 1986). The original extent of the Pahrump Group is unknown, but exposures extend from the Funeral Mountains in the Death Valley region as far south as Old Dad Mountain in the eastern Mojave Desert (Wright and others, 1976; Roberts, 1982; Wright and Prave, 1993) (fig. 1). Similar coeval deposits elsewhere in the western United States are represented by the Grand Canyon Supergroup of northern Arizona and by the Apache Group of southeastern Arizona (Stewart, 1972; Elston and others, 1993; Elston and Link, 1993; Link, 1993). These units have stratigraphy that is broadly similar to the Pahrump Group and, like the Pahrump Group, are intruded by Mesoproterozoic diabase.

Crystal Spring Formation

The stratigraphy, depositional environment, and tectonic setting of the Crystal Spring Formation in the Warm Spring Canyon area have been studied in detail by Wrucke (1966) and Roberts (1976, 1982). The studies of Roberts are based on numerous measured sections that were made from the Kingston Range (fig. 1) to the southern part of the Panamint Range.

The Crystal Spring Formation in the Warm Spring Canyon area comprises, in ascending order, mappable members of sandstone (unit Ᵽ$_M$cs), argillite (unit Ᵽ$_M$ca), cherty dolomite (unit Ᵽ$_M$cc), stromatolitic dolomite (unit Ᵽ$_M$cst), and chert (unit Ᵽ$_M$cch), all in a generally fining-upward, subhorizontal to moderately dipping sequence about 580 m thick, exclusive of intrusive diabase sills. According to Roberts (1976), these members are lithologically similar throughout the Kingston Range–Panamint Range area. Stratigraphically higher siltstone, sandstone, conglomerate, and dolomite of the Crystal Spring Formation, which crop out elsewhere in the Death Valley region, are referred to as the upper units of the formation (Wright and others, 1976; Maud, 1983; Calzia, 1997). The most accessible and complete section of the Crystal Spring Formation in the map area crops out along Warm Spring Canyon, from Warm Spring east to near the mountain front of the Panamint Range.

Delicate sedimentary structures are preserved in the cherty dolomite member of the Crystal Spring Formation. Small amounts of chlorite and white mica in the argillite member, however, indicate weak regional metamorphism to the chlorite-sericite subfacies of the greenschist facies. Contact metamorphism in cherty dolomite, mainly in narrow zones adjacent to diabase intrusions, has produced calcium-magnesium-silicate rock, including talc. Talc deposits in altered dolomite of the Crystal Spring Formation are known at many localities in the Death Valley region (Wright, 1952, 1963, 1968; Chidester and others, 1964). The largest talc deposits in the map area are at the Grantham Mine, near the east end of Warm Spring Canyon (Franklin, 1965); at an open pit mine about 1 km west of the Grantham Mine; and near the crest of the highest ridge to the north-northwest of Warm Spring (fig. 3).

Sandstone Member

The (basal) sandstone member (Ᵽ$_M$cs) of the Crystal Spring Formation is equivalent to the arkose and feldspathic sandstone members of Roberts (1976, 1982) and the feldspathic quartzite member of Wright (1952). In the map area, arkose makes up about the lower three-fourths of the member; feldspathic sandstone, the upper one-fourth. A conglomerate about 0.5 m thick, consisting of rounded pebbles of milky quartz,

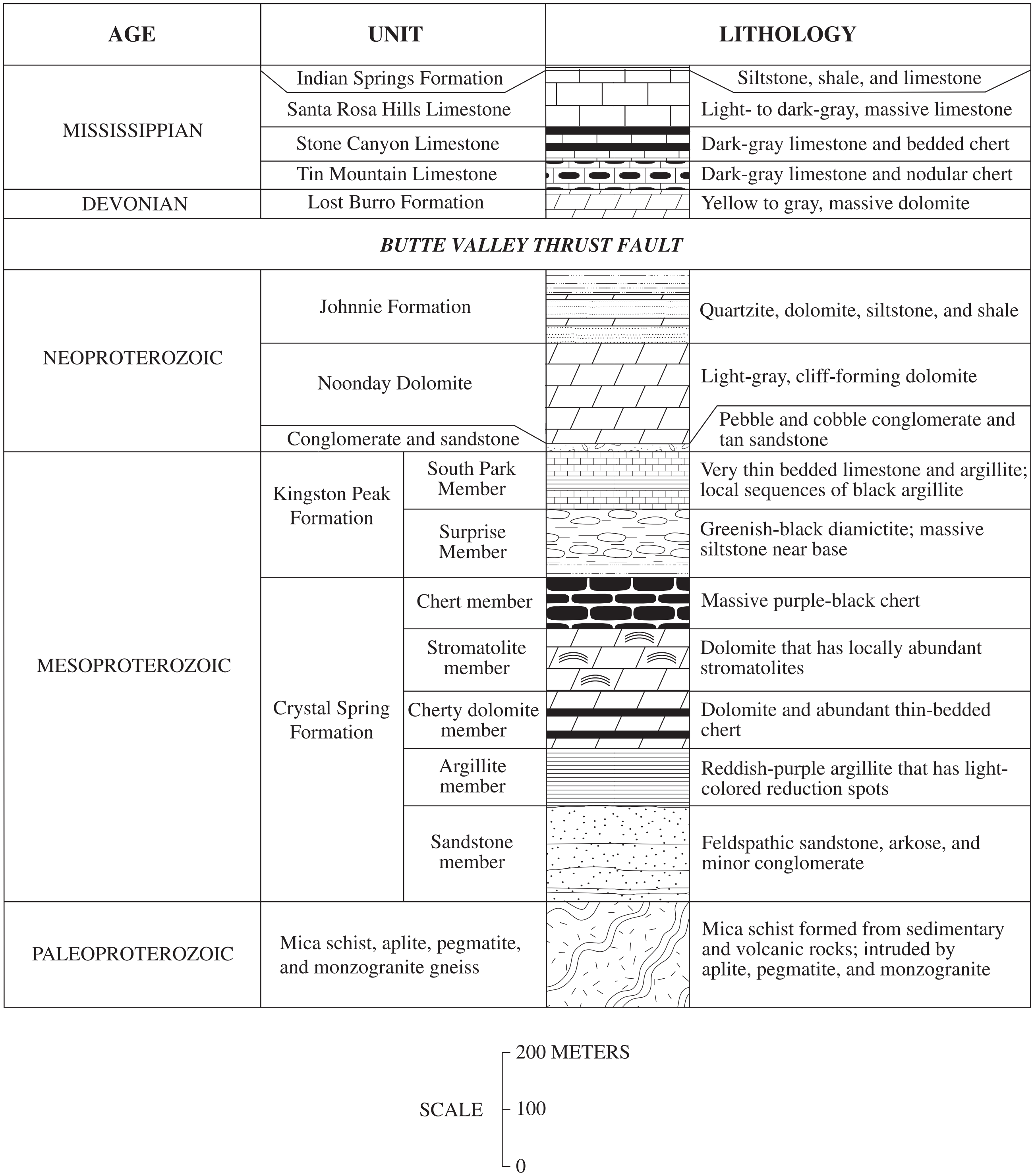

Figure 4. Chart showing stratigraphy of Proterozoic to Mississippian units in Warm Spring Canyon area.

schist, and gneiss derived from the Paleoproterozoic basement rocks, is present locally at the base of the member. The measured thickness of the sandstone member at a locality 2.5 km northeast of Warm Spring is 192 m.

The arkose is brownish gray in the lower part, becoming light yellowish gray and pinkish gray upward. It contains subordinate amounts of siltstone. The arkose mainly is medium to coarse grained and forms tabular beds that commonly are 15 cm to l m thick and less commonly are as much as 3 m thick. Beds typically contain abundant trough-crossbedded strata in sets 10 to about 50 cm thick, but many beds are planar laminated. Roberts (1976) reported an average composition of 35 percent quartz, 20 percent lithic fragments, 17 percent feldspar, 6 percent minor constituents, and 22 percent fine-grained matrix, mainly illite. Pebbles and cobbles of milky quartz, quartzite, schist, and gneiss are scattered throughout the unit but form thin beds and lenses of poorly sorted conglomerate at and near the base.

The feldspathic sandstone in the upper part of the sandstone member is pale red to light gray, weathering to shades of reddish brown and brownish gray. The strata commonly are fine to medium grained, but the uppermost few tens of centimeters are very coarse grained, a feature of regional extent. Beds typically are 5 to 40 cm thick and separated by shale partings. Internally they are laminated to cross-stratified. A few beds contain ripple marks. Locally the unit contains lenses of brownish-gray-weathering dolomite 8 cm thick and 30 cm long.

Roberts (1976), using measurements of cross-strata in conglomerates, interpreted the sandstone member in the southern part of the Panamint Range to have been deposited by south-flowing braided streams. The streams may have flowed into nearshore environments where the sediments were reworked by waves and currents. Herringbone cross-strata in the sandstone suggest deposition from tidal currents (Roberts, 1976).

Argillite Member

Overlying the sandstone member is the argillite member (Ᵽ_{M}ca), which consists of highly indurated mudstone (argillite) and subordinate amounts of siltstone, sandstone, and dolomite. The argillite typically is reddish purple, poorly bedded and nonfissile, and very fine grained. Small amounts of white mica and chlorite result from greenschist-facies metamorphism of presumed Mesozoic age. The member has a thickness of 107 m, as measured 1.6 km north of the Grantham Mine.

Argillite and silty argillite composing the lower two-thirds of the argillite member commonly have indistinct bedding and abundant light-gray to greenish- and yellowish-gray reduction spots, commonly 2 to 3 cm in diameter. Beds, where evident, average about 10 cm thick. Mud cracks and ripple marks are common. The upper one-third of the member is thin-bedded siltstone, silty argillite, and minor dolomite. Beds in much of this upper part are cross-laminated. Reduction spots are lacking. A 5-m-thick interval near the top of the unit, however, has bleached lenses commonly 5 cm thick and 15 cm long that contrast markedly with the argillite host, which here is very dark gray, almost black.

At the top of the member is a 9-m-thick section of pale grayish-purple dolomitic siltstone, sandstone, and subordinate dolomite. Wright (1952, 1968) distinguished these rocks as a separate quartzite member, and Roberts (1976) included them in the basal part of the overlying cherty dolomite member; however, they are included here with the argillite member because of the abundance of sand and silt.

The argillite member is here considered to have been derived from a deeply weathered source of low relief and to have accumulated in a low-energy environment, probably as mud flats.

Cherty Dolomite Member

Dolomite containing conspicuous beds and nodules of chert, which gradationally overlies the argillite member through a 5-m-thick interval, is here designated the cherty dolomite member (Ᵽ_{M}cc). It is part of the massive carbonate member of Wright (1952) and is equivalent to the dolomite member of Roberts (1976, 1982). Much of the member has been converted to talc. A measured thickness of the member northeast of the Grantham Mine is 115 m.

The cherty dolomite member consists of two distinctive types of dolomitic rock. The more abundant of these is grayish-orange dolomite in beds a few centimeters thick containing nodules and beds of chert commonly 1.5 to 2.5 cm thick. Dolomite is slightly more abundant than chert. The cherty dolomite commonly is crossbedded, and some of the chert is brecciated. The other common rock type is brown-weathering dolomite or dolomitic limestone containing abundant silt and microcrystalline quartz in fine wavy laminae and also in very thin disseminated masses that produce a crepe-like appearance. This crepe-like dolomite is present everywhere at the base of the member, as well as at higher positions, in layers commonly as much as 6 m thick that alternate with sequences of cherty dolomite 10 to 15 m thick.

The cross-stratification and the brecciated chert that typify the cherty dolomite suggest a high-energy environment, possibly the intertidal zone. The crepe-like dolomite, in contrast, probably accumulated in lower energy zones close to shore or in the supratidal zone. A supratidal environment is further suggested by molds resembling halite hoppers in some of the chert layers.

Stromatolite Member

Dolomite of the stromatolite member (Ᵽ_{M}cst) is equivalent to the algal member of Roberts (1976, 1982). The member, which forms distinctive brownish-red cliffs in the eastern part of Warm Spring Canyon, is about 110 m thick, some of it in sections greatly disrupted by diabase.

The stromatolite member consists of medium-gray, reddish-purple, and purplish-gray dolomite that weathers light brown to brown and reddish brown. Typical of the member are sequences of stromatolitic dolomite 1 to 18 m thick. The locally abundant stromatolites commonly consist of laterally linked stacked hemispheres 5 to 10 cm in diameter. Wrucke (1966) considered these stromatolites to be *Collenia* forms (Rezak,

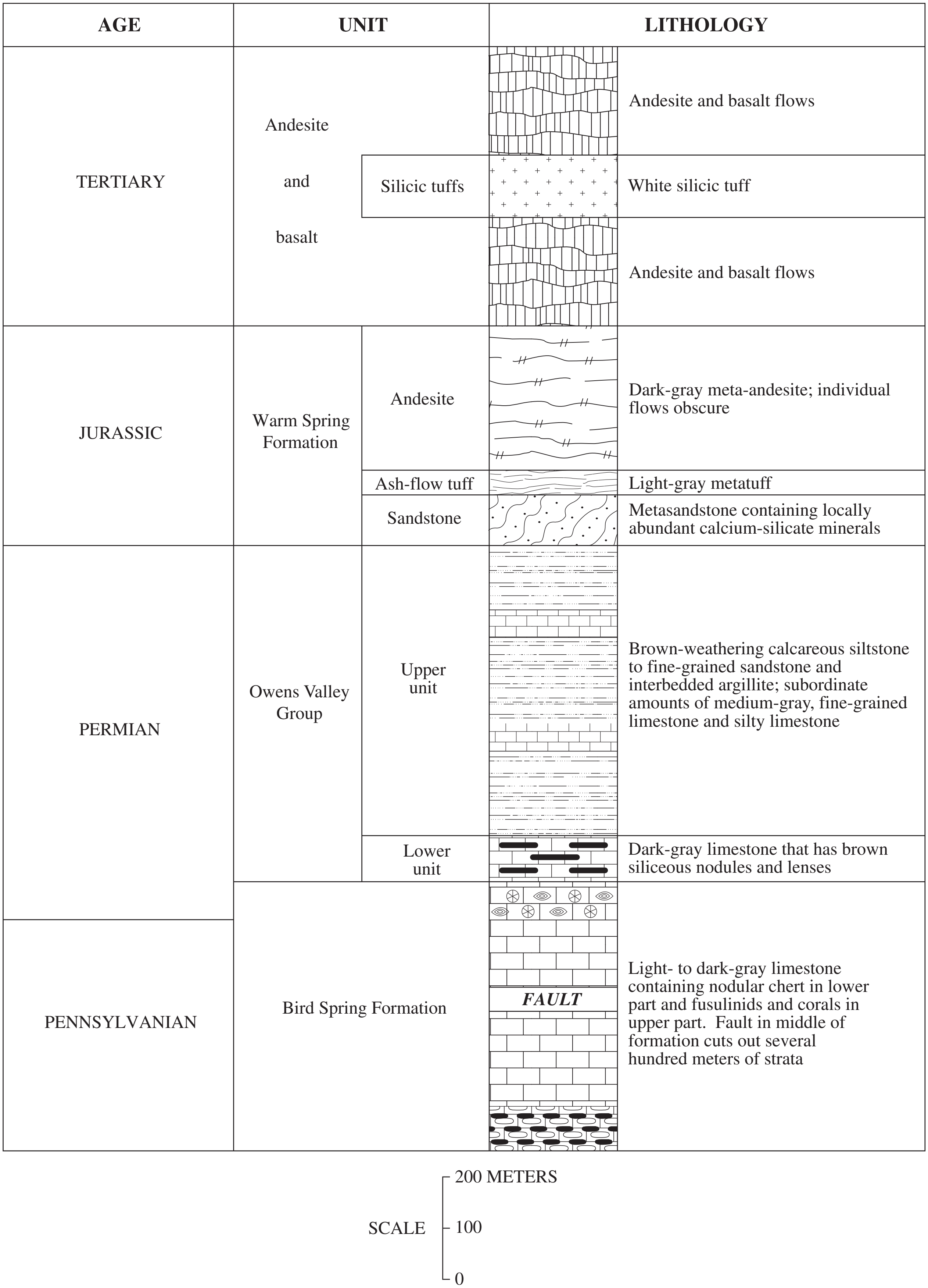

Figure 5. Chart showing stratigraphy of Pennsylvanian to Tertiary units in Warm Spring Canyon area.

1957), which may be the same as the *Conophyton* forms of Roberts (1982). A wavy hemispheroidal form identified as *Baicalia* (Howell, 1971) also has been found elsewhere in the Panamint Range. Some of the stromatolites are inclined, probably from tidal action. The stromatolitic dolomite alternates with planar-laminated to massive dolomite that lacks stromatolites and locally contains small amounts of chert. The massive layers locally contain sparse, poorly defined laminations suggestive of soft-sediment deformation.

Chert Member

Black and locally brownish-red chert of the chert member (Ᵽ$_{M}$cch) forms the highest member of the Crystal Spring Formation in the Warm Spring Canyon area. It rests on the stromatolite member (Ᵽ$_{M}$cst). The chert crops out only at a few localities: south of the Grantham Mine; on the north side of Warm Spring Canyon, near the east end of the map area; and at scattered localities in the north-central part of the map area. It is equivalent to the massive chert member of Wright (1952). Although the chert commonly is massive, at a few localities it is faintly bedded or laminated and contains limestone laminae. Dolomite rhombs and clots as large as 2 cm locally are conspicuous. Typically the chert forms steep slopes and ledges. At its thickest locality south of the Grantham Mine, an unfaulted sequence of the member is 80 m thick. The lower contact with the underlying stromatolite member is planar at most localities; the upper contact, also commonly planar, is against intrusive diabase.

Thin sections show that the chert contains abundant disseminated, very fine grained iron oxide, which accounts for the dark colors of the member. Some thin sections show faint laminations containing variable amounts of silt-size quartz; others reveal variable amounts of dolomite in clots and rhombs. One thin section shows irregular masses of coarse-grained plagioclase that may represent recrystallized tuff.

The chert member is here interpreted as originating during Proterozoic weathering, largely from silicification of carbonate strata that at one time made up the upper parts of the stromatolite member. Strong suggestion for the weathering can be found northeast of the Grantham Mine, where a massive body of chert several tens of meters in diameter appears to have filled a former solution cavity in the underlying stromatolitic dolomite. This chert is contiguous with the overlying massive member. The abundant iron oxide in the chert could be from iron that accumulated during Proterozoic weathering.

Evidence of silicification can be seen in the upper 5 to 10 m of the stromatolite member, adjacent to the chert infilling, where dolomite laminations pass laterally into chert laminations. Lenses of unreplaced dolomite as large as a meter in size are present in basal parts of the chert member, and scattered remnant dolomite beds remain higher up in the member, as do the scattered dolomite rhombs and clots found locally.

Age and Regional Relations

The age of the Crystal Spring Formation is not known precisely. It is, however, bracketed by the 1,400 Ma granite in the World Beater Complex that underlies the Crystal Spring Formation northwest of the Warm Spring Canyon area (Albee and others, 1981) and by the 1,080 Ma diabase (Heaman and Grotzinger, 1992) that intrudes the formation in Warm Spring Canyon and elsewhere (Hammond, 1990).

Shride (1967) commented on the remarkable similarity of the Crystal Spring Formation to rocks of the Apache Group in Arizona but thought it presumptuous to suggest correlation. Wrucke (1966, 1989, 1993), however, presented various lines of evidence for their possible correlation. Particularly noteworthy is that the Crystal Spring Formation of the Pahrump Group and the Mescal Limestone (actually a metamorphosed dolomite) of the Apache Group have closely similar sequences of cherty dolomite overlain by stromatolitic dolomite. Proterozoic weathering and silicification after deposition of carbonate units, which are strongly suggested in the Crystal Spring Formation, are well documented in the Mescal Limestone of the Apache Group. Yet another notable similarity is that diabase sills intruded similar stratigraphic horizons in both formations. Finally, the Crystal Spring Formation and the Apache Group both contain feldspathic arenites and deep-red argillite that has reduction spots, though in different stratigraphic order. Because of the close lithologic similarities of the two units, both intruded by diabase, Stewart (1972) suggested that they might have accumulated in different parts of the same basin. Zircons from feldspathic arenite of the Dripping Spring Formation, which lies beneath the Mescal Limestone in the Apache Group, have yielded an age of 1.26 Ga, indicating a maximum age for the feldspathic rock, as well as possibly an approximate depositional age because of the suggestion that the zircons resulted from contemporaneous volcanism (Stewart and others, 2001). If correlation with the Apache Group is correct, the Crystal Spring Formation may also be about 1.26 Ga.

Diabase

Dark greenish-black diabase (unit Ᵽ$_{M}$d) intruded strata of the Crystal Spring Formation as sheets and sills as thick as about 90 m and locally as multiple tabular bodies having a total thickness of 120 m. Six sills, principally in the dolomite units, are well exposed in the eastern part of Warm Spring Canyon. One sill (unit Ᵽ$_{M}$ds) above the Grantham Mine on the south side of Warm Spring Canyon was emplaced in the middle of another sill and is identified by fine-grained, chilled contacts of the younger body against coarse-grained central parts of the host. A few diabase dikes connect the sills, and dikes also are found in the Paleoproterozoic mica-schist basement. Many of the diabase bodies intrusive into Crystal Spring strata are interconnected and, therefore, resulted from a single magmatic injection. The presence of the one sill (Ᵽ$_{M}$ds) in another, however, indicates that more than one intrusion did occur.

Diabase in the Warm Spring Canyon area invaded strata as high as the chert member of the Crystal Spring Formation (Ᵽ$_{M}$cch); however, elsewhere in the Death Valley region, diabase intruded horizons in the overlying upper units of the Crystal Spring (Hammond, 1983). The diabase, therefore, is younger than the Crystal Spring Formation, but it is older than the middle member of the Kingston Peak Formation of Calzia and Troxel (1993), which contains diabase clasts. Because the Crystal Spring Formation, the overlying Beck Spring Dolomite, and the lower member of the Kingston Peak Formation in the Kingston Range form a continuous stratigraphic section, Calzia and Troxel (1993) concluded that the diabase probably was intruded in post-lower but pre-middle Kingston Peak time. Dehler and others (2001), however, reported that the diabase may be older than the Beck Spring Dolomite; no diabase has been found intrusive into the Beck Spring Dolomite (Calzia, 1997).

Several petrographic studies (Wrucke, 1966; Hammond, 1983, 1986) have shown that the diabase crystallized as a coarse-grained ophitic rock of basaltic composition. Before final solidification, most of the diabase was intensely altered, mainly to albite, actinolite, and chlorite that clearly preserve evidence of the original ophitic texture. Many plagioclase crystals, now mostly albite, contain remnants of original labradorite (as high as An_{66}) and andesine. Remnant original augite is sparse. Pyroxene grains have been converted to actinolite, much of which has been further altered to chlorite, green biotite, and sparse quartz; the small amount of quartz may have resulted from contamination. Minor amounts of brown biotite, partly altered to green biotite, are associated with ilmenite. Other accessory and secondary minerals include apatite, sphene, zircon, kaersutite, ferrohastingsite, and hydrogarnet. No olivine was found, but a few rounded granular accumulations of magnetite and chlorite suggest altered olivine in the margin of one sill. The amount of chlorite correlates with the thoroughness of alteration, which varies from sill to sill.

The unusually thorough alteration and the extensive development of hydrous minerals suggest that the diabase crystallized under conditions that allowed retention of water. The high water content might have come from connate sources in adjacent strata of the Crystal Spring Formation, a reasonable possibility considering that the highest diabase sill in the Kingston Range invaded soft, wet sediments in upper strata of the formation (Hammond, 1983). Another possible source may be from dehydration of clay minerals in Crystal Spring strata during contact metamorphism. Diabase dikes 100 m or more below the lowest beds of the Crystal Spring Formation, however, were as rich in water as the sills high in the stratigraphic section, suggesting that the original diabase magma may have been water rich. Whatever the source, water was abundant during deuteric alteration of early-formed minerals.

Despite the abundance of silica, especially in the cherty dolomite member of the Crystal Spring Formation, metamorphic minerals other than talc are sparse more than a few meters from the Crystal Spring–diabase contact, and in some localities they are absent more than 15 cm from the contact. Because siliceous dolomite is highly sensitive to thermal metamorphism, the general absence of metamorphic minerals indicates that metamorphism did not take place in an open system where CO_2 could be liberated during genesis of calcium-magnesium silicate minerals. If egress of CO_2 was prevented, metamorphic minerals would not have developed beyond an initial stage, and late-aqueous solutions available for alteration of the diabase would have been concentrated (Wrucke, 1966). Conditions appear to have changed sufficiently late in the metamorphic process to allow this water to play a role in developing talc in contact-metamorphic zones in the siliceous dolomite adjacent to diabase. Talc is especially concentrated and unusually abundant in areas where diabase has broken across the stratigraphic section, as at the Grantham Mine and also near the north-central border of the map area. High concentrations of late-magmatic water in these areas, together with numerous fractures that likely developed in the siliceous-dolomite host during diabase emplacement, would have provided ideal conditions for development of abundant talc.

A notable feature of the diabase in the Warm Spring Canyon area is the large size of original crystals in the central parts of sills. In particular, the coarsest parts of two sills have plagioclase crystals 5 to 7 mm long, as well as augite crystals commonly 10 mm long and some as long as 4 cm. These crystal dimensions are comparable to those in the Arizona sills (Wrucke, 1966, 1989; Shride, 1967). The high water content, determined to have been commonly over 3 percent total H_2O in diabase intrusions in the Crystal Spring Formation (Wrucke, 1966), may be the reason for the large crystal size, reflecting an increase in fugative components as the diabase crystallized and approached deuteric alteration.

Wrucke (1966), using the basalt tetrahedron of Yoder and Tilley (1962), described the diabase at Warm Spring Canyon as an olivine tholeiite, whereas Hammond (1986) concluded that the diabase is mildly alkaline olivine basalt, on the basis of a detailed study of the major- and minor-element chemistry, including rare-earth-element patterns. However, Hammond (1986) also noted that iron enrichment in the diabase indicates a tholeiitic affinity. The alkaline nature of the rock is revealed by any of the several possible diagrams of alkalis versus SiO_2, and Hammond (1986) found that the original augite contained 1.56 percent TiO_2, a value comparable to the augite in alkaline basalts. Hammond's (1986) work has provided considerable information on the differentiation of the diabase, as well as on its source and melting history.

Examination of the petrography and chemistry of diabase from sills in the Apache Group of southern Arizona, the Unkar Group in the Grand Canyon, and the Pahrump Group of the Panamint Range led Wrucke (1966) to conclude that the diabases in all three areas are identical and, therefore, correlative. Hammond (1986, 1990) also concluded that the diabase of the Death Valley region is chemically identical to the Arizona diabases. Additional information on the chemistry of the diabase, as well as on the correlation of diabase in southern Death Valley with diabase sills in the Apache Group and the Unkar Group, is supplied by Shride (1967), Wrucke and Shride (1972), Elston (1989), Wrucke (1989, 1993), Hammond and Wooden (1990), Howard (1991), and Elston and others (1993). Ophitic coarse-grained Mesoproterozoic diabase also is widely exposed in discordant horizontal sheets in crystalline basement rocks, commonly granitic plutons, in a large area in southeastern California and western Arizona (Fitzgibbon and Howard, 1987; Fitzgibbon, 1988; Hammond, 1990; Howard, 1991). Thus, the Death Valley sills belong to diabase of regional extent in the southwestern United States.

The age of the diabase is known from isotopic data for a number of localities in the southwestern United States, including the southern Death Valley region (table 1). Ages from Arizona and California range from 1,069 ± 3 to 1,153 ± 30 Ma, which suggests that the approximate age of the diabase is 1,100 Ma. Data are too few to determine if the diabase was emplaced in more than one intrusive episode and what the duration may have been for any single intrusive pulse. Although the diabase in the Warm Spring Canyon area has not been dated isotopically, it is correlative with diabase in the Death Valley area that has yielded U-Pb ages of 1,089 ± 3 Ma and 1,069 ± 3 Ma (Heaman and Grotzinger, 1992).

Kingston Peak Formation

The Kingston Peak Formation, the youngest formation in the Pahrump Group, is divided into four members in the Panamint Range (Miller, 1985): they are, in ascending order, the Limekiln Spring, Surprise, Sourdough Limestone, and South Park Members. Of these, only the Surprise Member, composed of diamictite (unit $\mathrm{P}_N$ksd) and siltstone (unit $\mathrm{P}_N$kss), and the South Park Member ($\mathrm{P}_N$ksp), composed of limestone and argillite, are recognized in the Warm Spring Canyon area. The Kingston Peak Formation in the Warm Spring Canyon area is near the midpoint in exposures of the formation in the Death Valley region; from Warm Spring Canyon, it extends southeast to the Kingston Range and Silurian Hills and north to the Funeral Mountains (fig. 1).

Surprise Member

The Surprise Member, originally named as a formation by Murphy (1930) for exposures in the Telescope Peak area, was redefined as a member of the Kingston Peak Formation by Johnson (1957). At Warm Spring Canyon, the Surprise Member primarily is a diamictite composed of pebbly to bouldery metamorphosed mudstone but also contains minor amounts of siltstone. The largest exposure of the member is in Tram Canyon, in the west-central part of the map area (fig. 3); one small exposure is on the east side of Galena Canyon, near the north edge of the map area (at cross section *C–C'*). The diamictite (unit $\mathrm{P}_N$ksd) consists of pebbles, cobbles, and boulders randomly dispersed in a much more abundant, very fine grained, greenish-black matrix. Clasts larger than 1 cm make up an estimated 5 percent of the rock. In general, the pebbles are angular to subangular; larger clasts are subrounded to rounded. The largest clast, far larger than any other, is a limestone block 5 by 5 by 9 m in size, found high in the member on the east side of Tram Canyon. Granite and granitic gneiss are the most abundant clast types, followed by limestone and sparse diabase. Of these, only diabase has a possible source in the immediate Warm Spring Canyon area. The matrix is hornfels, originally a silty to sandy mud that is now composed principally of brownish-green biotite and subordinate amounts of actinolite and sparse calcite. Angular detrital silt and sand of quartz, albite, and quartzite make up about 10 percent of the matrix. Most exposures are structureless, although a vague suggestion of stratification is observed locally.

A sequence of dark purplish-gray and locally greenish-gray siltstone (unit $\mathrm{P}_N$kss) about 10 m thick is present in the lower part of the Surprise Member in Tram Canyon. The siltstone is extremely well indurated and generally massive,

Table 1. Isotopic ages of Mesoproterozoic diabase in Arizona and California.

Location	Source	Mineral or Rock Dated	Method	Age (Ma)
Saratoga Springs, Calif.	Heaman and Grotzinger, 1992	Baddeleyite	U-Pb	1,069 ± 3
Kingston Range, Calif.	Heaman and Grotzinger, 1992	Baddeleyite	U-Pb	1,087 ± 3
Hance Rapids, Grand Canyon, Ariz.	Hendricks, J.D., *in* Hammond and Wooden, 1990	Whole rock	Rb-Sr	1,153 ± 30
Shinumo Creek, Grand Canyon, Ariz.	Elston and McKee, 1982	Whole rock	Rb-Sr	1,070 ± 30
Hualapai Mts., Ariz.	Stensrud, H., *in* Hammond and Wooden, 1990	Whole rock	K-Ar	1,070 ± 25
Little Dragoon Mts., Ariz.	Silver, 1978	Zircon	^{207}Pb/^{205}Pb	1,100 ± 15
Sierra Ancha, Ariz.	Damon and others, 1962	Biotite	K-Ar	1,150[1]
Sierra Ancha, Ariz.	Neuerburg and Granger, 1960	Uraninite[2]	^{207}Pb/^{205}Pb	1,050
Sierra Ancha, Ariz.	Neuerburg and Granger, 1960	Uraninite[2]	^{207}Pb/^{205}Pb	1,104
Sierra Ancha, Ariz.	Silver, 1978	Zircon from diabase pegmatite	^{207}Pb/^{205}Pb	1,100 ± 15
Sierra Ancha, Ariz.	Silver, 1963	Uraninite and zircon from diabase differentiate	^{207}Pb/^{205}Pb	1,150 ± 30[3]
Sierra Ancha, Ariz.	Shastri and others, 1991	Baddeleyite and zircon	^{207}Pb/^{205}Pb	1,100 ± 2

[1]Originally reported as 1,140 Ma but here recalculated using constants of Dalrymple (1979).
[2]Additional uraninite samples that were reported by Neuerburg and Granger (1960) were contaminated.
[3]Rerun of preliminary age that was originally reported as 1,075±50 Ma (Silver, 1960).

although irregular lenses 2 by 10 to 20 cm of slightly different siltstone are present locally. Thin sections show that the rock consists of faint laminations containing 30 percent angular quartz grains, mostly 0.03 to 0.10 mm in diameter, in a very fine grained matrix of brown biotite. A few conspicuous lenses of light-gray and bluish-gray limestone 0.5 to 3 m thick and 10 m long crop out in the siltstone on the east side Tram Canyon, and a few lenses of yellow dolomite of similar size also are present in the northernmost siltstone exposures in the canyon.

Siltstone that underlies the diamictite east of Galena Canyon is about 30 m thick and contains a few beds of gray limestone and yellow dolomite, similar to those in Tram Canyon.

The thickness of the diamictite-siltstone sequence in Tram Canyon probably is about 100 m. An 80-m-thick section (base not present) is exposed from the canyon bottom to the upper contact with the South Park Member. An additional 20 m or so of the sequence crops out to the northeast across a fault where the basal contact is exposed. This basal contact climbs rapidly eastward across diabase and strata of the Crystal Spring Formation.

The basal contact of the Surprise Member on the east side of Tram Canyon may represent the eastern margin of a depositional basin. The diamictite of the Surprise Member thickens west of the Warm Spring Canyon area to more than 750 m in Redlands and South Park Canyons and, northwest of the area, to more than 1,000 m in the Telescope Peak 15' quadrangle (Labotka and others, 1980; Miller, 1985).

South Park Member

Rocks herein assigned to the South Park Member (Ᵽ$_N$ksp) of the Kingston Peak Formation overlie the diamictite of the Surprise Member on a planar contact exposed in Tram Canyon. The South Park Member was named by Johnson (1957) for exposures at South Park in the Manly Peak 7.5' quadrangle west of the Warm Spring Canyon area (fig. 2). The South Park Member crops out widely at other places on the west side of the Panamint Range (Johnson, 1957; Miller, 1985), where it is separated from the Surprise Member by the Sourdough Limestone Member.

In the Warm Spring Canyon area, the South Park Member mainly consists of black to medium-gray, very thin bedded to laminated limestone interlayered with well-defined thin beds and laminae of calcareous argillite and silty argillite. The argillite now is very fine grained biotite(?). Typically the limestone and argillaceous layers are 1 to 5 mm thick in sections composed of about 60 percent limestone and 40 percent argillite. The limestone weathers medium gray to medium-dark gray and the argillite weathers in shades of brown, the argillaceous layers standing out slightly from the carbonate beds.

Dolomite, dominantly light yellowish brown but, in part, medium gray and about 20 m thick, crops out at the base of the South Park Member in the lower reaches of Tram Canyon. This dolomite is massive to thin bedded and lacks the distinctive argillaceous layers of higher parts of the member.

Interspersed with the typical limestone-argillite rock are a few sequences of black to purplish-black argillite as much as 25 m thick. These sequences commonly contain less than 5 percent limestone in beds about 1 cm thick; some argillite sequences contain no limestone. A few isolated beds of light-gray-weathering flat-pebble-limestone conglomerate, 0.5 to 1 m thick, also are interlayered with the thin-bedded limestone and argillite. One of these conglomerate beds is within purplish-black argillite in the upper part of Tram Canyon. The limestone clasts commonly are a few centimeters thick and 10 to 20 cm across, in a limestone matrix. Brownish yellow-orange limestone breccia is present locally in a few beds 0.5 to 1 m thick.

Correlation of the limestone and argillite in Tram Canyon with the South Park Member in the western part of the Panamint Range is based on identical limestones in the two areas. The lower 40 percent of a section in Redlands Canyon southwest of Warm Spring Canyon (fig. 2) contains abundant laminated limestone similar to that at Tram Canyon; the section also has at least one flat-pebble conglomerate like that in Tram Canyon. The Redlands Canyon section, however, differs from the one in Tram Canyon in that it has considerable sandstone interbedded with the limestone. Another difference between the Tram Canyon and Redlands Canyon sections is that, in Redlands Canyon, the Sourdough Limestone Member lies between the South Park Member and the underlying Surprise Member. The Sourdough Limestone Member forms a distinctive marker unit in the western part of the Panamint Range and, though in part laminated, overall does not resemble the limestone in Tram Canyon. Despite the absence of sandstone, the position of the Tram Canyon section above the diamictite of the Surprise Member, as well as the striking similarity of the limestone with the Redlands Canyon section, support correlation with the South Park Member.

The South Park Member in the upper reaches of Tram Canyon is at least 120 m thick. In the western part of the Panamint Range, the thickness of the member, as reported by Johnson (1957), ranges from 87 m in Goler Canyon to 305 m in South Park Canyon (fig. 2).

The consistent thin, laminated, almost varve-like, planar bedding typical of the limestone of the South Park Member is interpreted as indicating deposition in a lake or quiet basin. Black argillite and limestone suggest anoxic conditions. Locally the beds are slightly kinked and crumpled, possibly from soft-sediment deformation. Flat-pebble-limestone conglomerate may have formed when limestone beds exposed at the surface were desiccated or possibly broken by syneresis action and subsequently dismembered by downslope slippage.

Origin and Age of the Kingston Peak Formation

The origin of the Kingston Peak Formation, particularly the Surprise Member, has been controversial for many years, one group contending that it is of glacial origin and another espousing deposition by nonglacial processes. A glacial origin of the diamictite was first mentioned by Hazzard (1939, 1940). Hewett (1940), however, concluded that the diamictite at the type locality in the Kingston Range (fig. 1) is a fanglomerate. Johnson (1957), who studied the diamictite in the western Panamint Range, stated that the depositional mode of the rock is less likely to be from glacial processes than from marine processes. Troxel (1966, 1967) concluded that the diamictite may have had a complex origin, part glacial and part turbiditic. Miller (1982, 1985, 1987) concluded that the diamictite of the Surprise Member is glaciogenic, on the basis of striated stones and dropstones. Calzia (1997) believed the diamictite to be of local origin, possibly under fluvial and marine conditions involving slumping of rock material into tectonically active depositional basins. Other geologists (Prave, 1999; Abolins and others, 2000; Kennedy and others, 2001a) have referred to the rock as glacial, having been deposited during a worldwide snowball-earth glaciation (Hoffman and Schrag, 2000).

The diamictite in the Surprise Member is now generally considered to be glaciogenic (Stewart, 1976; Miller, 1985; Prave, 1999; Abolins and others, 2000; Vogel and others, 2002), although it also may contain some nonglacial deposits (Troxel, 1967, 1968). The overlying Sourdough Limestone Member (not present in the map area) is generally considered to be a cap carbonate (Prave, 1999). Cap carbonates are envisioned (Hoffman and Schrag, 2000) to have been deposited during a time of greatly increased temperatures immediately after melting of global ice. Prave (1999) showed that the $\delta^{13}C$ pattern of the Surprise Member diamictite and Sourdough Limestone Member compare well with the patterns from deposits in the South Congo craton that are considered to be from a worldwide climatic reversal (Kaufman and others, 1997; Hoffman and others, 1998), suggesting correlation of glaciation in these widely spaced areas. This glaciation is considered to be Sturtian, which is bracketed as 760 to 700 Ma (Brasier and others, 2000; Corsetti and others, 2000).

Neoproterozoic rocks that are lithologically similar to diamictite of the Surprise Member crop out from California to British Columbia and may represent a time of widespread glaciation along the western margin of North America (Cloud, 1971; Stewart, 1972). Whatever their origin, they record depositional environments that differ greatly from those of the underlying and overlying strata. The base of the Kingston Peak Formation may mark the beginning of continental separation and the breakup of the Rodinian supercontinent about 750 Ma (Stewart, 1972).

Conglomerate and Sandstone

A thin, unnamed unit of conglomerate and sandstone (unit $\mathrm{P_N}$cs) rests unconformably on diabase at three mapped localities in the eastern part of the Warm Spring Canyon area. Conglomerate dominates at one locality, 2.8 km west-southwest of the Queen of Sheba Mine; only sandstone is present at the other two localities, 1.5 km south-southeast and 1.2 km west-southwest of the Queen of Sheba Mine (fig. 3). The conglomerate and sandstone are here interpreted as remnants of alluvial deposits that accumulated on the unconformity prior to deposition of the overlying Noonday Dolomite.

Although the lower contact of the conglomerate and sandstone unit is mapped as depositional, a thin section of the lowest rock in the locality south-southeast of the Queen of Sheba Mine shows minor brecciation, indicating at least some structural disruption. The upper contact, poorly exposed at all three localities, is here interpreted as having been faulted.

At the conglomeratic locality 2.8 km west-southwest of the Queen of Sheba Mine, a lenticular body as much as 6 m thick and 210 m long consists of 20 to 30 percent clasts in a brown-weathering siltstone and sandstone matrix. Most clasts are angular to subrounded pebbles and cobbles of sandstone, but a few percent are of gray carbonate rock and diabase. A subangular boulder of diabase 45 cm in diameter is present near the base of the conglomerate. The feldspathic sandstone matrix consists primarily of quartz, microcline, and sparse altered plagioclase, all in a groundmass of fine-grained dolomite. The uppermost meter of the conglomerate is reddish brown from abundant iron oxide. This conglomerate differs from the diamictite of the Surprise Member, exposed as close as 1.4 km to the west, in that it has a brown-weathering sandy and silty matrix, sandstone pebbles and cobbles, and as much as 30 percent clasts (compared to 5 percent clasts in the diamictite, which has a green matrix).

At the sandy locality 1.5 km south-southeast of the Queen of Sheba Mine, the unit consists of three beds that have a maximum combined thickness of about 3.5 m. The lowest bed, 0.5 m thick, is microbrecciated dolomitic sandstone that contains abundant irregular quartz veinlets. The middle bed, 1.5 m thick, consists of a felted mass of white mica containing abundant apatite needles, irregular quartz blebs, and skeletal iron oxide; this bed may be an altered tuff. The upper bed, also 1.5 m thick, is fine-grained sandstone that contains sparse, partly resorbed quartz crystals and small pebbles in a very fine grained matrix that contains white mica and as much as 20 percent small blebs of iron oxide. A quartzite cobble 15 cm in diameter was found near the north end of the unit outcrop, where the three separate beds cannot be distinguished.

At the sandy locality 1.2 km west-southwest of the Queen of Sheba Mine, the unit consists of 15 cm of sandstone that has a dolomitic matrix and sparse, partly resorbed quartz grains, similar to those found south-southeast of the mine. The partly resorbed quartz grains at these localities could be of volcanic origin; alternatively, the resorption could have resulted from a postdepositional reaction with the sandstone matrix.

Conglomerate and sandstone similar to that in the Warm Spring Canyon area also are present below the Noonday Dolomite elsewhere in the Death Valley region. Wright (1973) reported discontinuous layers of conglomerate a meter or so thick locally below the dolomite in the northern part of the Alexander Hills and in the Nopah Range (fig. 1). James Calzia (oral commun., 2004) reported a boulder conglomerate 5 m thick that has a pink to tan sandy-siltstone matrix below the dolomite at the War Eagle Mine in the Nopah Range, as well as a pebble to cobble conglomerate 1 m thick that has a yellowish-brown sandy matrix in the eastern Kingston Range. We sampled the sandstone matrix of the conglomerate at the War Eagle Mine and found that it, like that west of the Queen of Sheba Mine, contains a high percentage of feldspars, including microcline. Evidence of tuffaceous material has not been reported in these rocks.

The Neoproterozoic age of the conglomerate and sandstone unit is between that of the Kingston Peak Formation, which is thought to have been deposited in the interval 760 to 700 Ma, and that of the overlying Noonday Dolomite, which may be as young as 590 Ma or younger.

Noonday Dolomite

In the Warm Spring Canyon area, the Noonday Dolomite (Ꝑ$_{N}$n) is here interpreted as resting on a faulted unconformity. The lowermost beds of dolomite generally are parallel to the basal contact but can vary from weakly deformed to highly disturbed. The formation rests on a variety of Mesoproterozoic and Neoproterozoic rocks—the cherty dolomite (Ꝑ$_{M}$cc), stromatolite (Ꝑ$_{M}$cst), and chert (Ꝑ$_{M}$cch) members of the Crystal Spring Formation, the diabase (unit Ꝑ$_{M}$d), and the Surprise (units Ꝑ$_{N}$ksd, Ꝑ$_{N}$kss) and South Park (Ꝑ$_{N}$ksp) Members of the Kingston Peak Formation—but in the eastern Death Valley region the base of the Noonday Dolomite extends down to Paleoproterozoic basement gneiss (Hazzard, 1937).

The Noonday Dolomite is exposed in high parts of the Warm Spring Canyon area, along the north border and near the southeastern corner of the map area. Its pale-yellow and light-gray hues contrast sharply with the darker colors of most underlying and overlying units. This dolomite probably is correlative with the Sentinel Peak Member of the Noonday Dolomite in the Telescope Peak 15' quadrangle to the northwest (Albee and others, 1981).

No undisturbed section of the Noonday Dolomite is present in the Warm Spring Canyon area, and so its thickness cannot be determined accurately. Most of the lower part of the formation (about 100 m thick) consists of pale-yellow to light-gray dolomite that is micritic to fine grained and laminated to very thin bedded. Bedding locally is obscure. The upper part of the formation (about 60 m thick) is light-gray, medium-grained dolomite that commonly is massive but locally has discontinuous beds 0.6 to 4 m thick. Excellent exposures of this part of the formation are present in the northwest corner of the map area. The uppermost 3 m of the formation, which lies beneath the conformably overlying Johnnie Formation, consists of light-brown, sandy cross-laminated dolomite in beds about 10 cm thick. This part of the Noonday Dolomite is transitional with the Johnnie Formation, which contains similar strata interbedded with quartzite.

Locally the lower part of the Noonday Dolomite contains closely spaced, parallel-cylindrical, tube-like structures that are 1 to 3 cm in diameter, commonly 20 cm to more than 1 m long, and normal to bedding. The structures are partly to completely filled with coarse-grained sparry dolomite; at the Queen of Sheba Mine, they are filled with galena (Morton, 1965). Sedimentary layers adjacent to the tubes are undeformed. The tube-like structures are present in the Noonday Dolomite throughout the southern Death Valley region (Hunt and Mabey, 1966; Cloud and others, 1974).

The origin of the tube structures is controversial. Wrucke (1966) suggested that they may be solution phenomena. Hunt and Mabey (1966) noted that they are suggestive of *Scolithus* tubes, but *Scolithu*s is not known to have lived in the Proterozoic and, thus, were discounted by Cloud and others (1974) as the cause. Instead, Cloud and others (1974) suggested that the tubes developed from solutions or gas propagated upward under hydrostatic pressure from within algal mounds. Kennedy and others (2001b) attributed the tubes to rapid and abundant release of methane from gas hydrate destabilized either in permafrost or in deposits along continental rises during rapid warming following a glacial episode; they described and illustrated similar structures in cap-carbonate rocks in Namibia. Marenco and others (2002) proposed that the tubes represent an unusual stromatolite morphology that grew upward at high places on microbial mats. No consensus on the origin in the tube structures is yet at hand.

The age of the Noonday Dolomite is not closely constrained. Miller (1987) concluded that the age of the formation is 700 to 600 Ma, on the basis of her interpretation that the age is close to that of the Kingston Peak Formation. Prave (1999), however, disagreed, believing instead that the Noonday Dolomite rests unconformably on the Kingston Peak Formation. He concluded that the Noonday Dolomite is Varangerian (590 Ma or younger), the same conclusion reached by Abolins and others (2000).

Johnnie Formation

The Johnnie Formation (Ꝑ$_{N}$j), the youngest Proterozoic unit in the Warm Spring Canyon area, is exposed in the extreme northwestern part of the map area. Only the lower part of the formation, probably equivalent to the transitional member of Stewart (1970), is present in the map area; higher beds in the formation are exposed north of the area. A depositional contact with the Noonday Dolomite is well exposed locally, but elsewhere in the Warm Spring Canyon area the base of the Johnnie Formation is a low-angle fault. The conformable basal contact is here placed at base of the lowest quartzite bed that breaks the succession of carbonate rocks of the underlying Noonday Dolomite.

The incomplete section of the Johnnie Formation in the Warm Spring Canyon area is about 130 m thick, only a small fraction of the 430 to 750 m total thickness found to the west and north in the Panamint Range (Johnson, 1957; Hunt and Mabey, 1966; Stewart, 1970). The lower 100 m in the map area consists of interbedded quartzite, sandy dolomite, and dolomitic quartzite; the upper 30 m consists of shale, siltstone, and quartzite. The lower 100 m is about 25 percent quartzite and 75 percent sandy dolomite and dolomitic quartzite; a notable 12-m-thick section of quartzite begins 2.4 m above the base of the formation. Beds commonly are 10 cm to 2.4 m thick and are conspicuously crossbedded. Generally the quartzite is fine to medium grained and light to medium gray on fresh and weathered surfaces, whereas the dolomite commonly has abundant coarse sand grains and weathers in shades of brown. A few conspicuous beds of stromatolitic dolomite, each about 60 cm thick, crop out 55 m above the base. Stewart (1970) reported stromatolites 47 to 83 m above the base of the Johnnie Formation 6 km northeast of the Warm Spring Canyon area.

The shale, siltstone, and quartzite that make up the upper 30 m of the Johnnie Formation in the map area are laminated to thin bedded and generally yellowish brown to medium brown. The highest beds of the Johnnie Formation exposed in the Warm Spring Canyon area are composed of a 10-m-thick quartzite.

Paleozoic Rocks

Paleozoic formations exposed in the Warm Spring Canyon area are the Devonian Lost Burro Formation (Dlb); the Mississippian Tin Mountain Limestone (Mtm), Stone Canyon Limestone (Msc), Santa Rosa Hills Limestone (Msr), and Indian Springs Formation (Mis); the Pennsylvanian and Early Permian Bird Spring Formation (PPbs); and the Early Permian Owens Valley Group (Pov and Povl). These units, which form a southward-younging stratigraphic sequence, are broken internally by one major fault and are separated from the Proterozoic rocks to the north by Jurassic or Cretaceous tonalite (unit KJt) that intrudes the inferred trace of the Butte Valley Thrust Fault.

All units except those of the Owens Valley Group were deposited on a broad carbonate platform, which extended over much of southern Nevada and adjacent parts of California. The Owens Valley Group is composed primarily of siliciclastic rocks, including turbidites, that were deposited in relatively deep water.

Paleozoic rocks in the Warm Spring Canyon area and at Striped Butte, 5 km to the southwest in Butte Valley (fig. 2), originally were assigned to a single heterogeneous unit called the Anvil Spring Formation (Johnson, 1957; Wrucke, 1966); however, more recent stratigraphic and paleontologic studies of these rocks (for example, Stone, 1984; Stone and Stevens, 1984) have shown that the Anvil Spring Formation comprises rocks of Devonian to Permian age identified as separate formations in this report. The name Anvil Spring Formation, therefore, is no longer needed and is herein abandoned.

One goal of this study has been to document the fossil content of the Paleozoic rocks. Table 2 lists the fossils identified at 28 mapped localities in the area (fig. 3); of these localities, 16 were originally described by Wrucke (1966) and six by Stone (1984). During the course of this study, many of these localities were revisited, and additional fossils were collected from some. In particular, Stevens (this report) studied in detail the fusulinids and corals from the Bird Spring Formation, in part by reexamining the previous collections of Stone (1984).

Devonian and Mississippian Rocks

Devonian and Mississippian rocks in the map area have not been examined in detail. All of these units can be recognized and mapped, with only relatively minor variations in lithology and thickness, throughout the greater Death Valley region, reflecting the uniformity and stability of depositional and tectonic environments that characterized the broad carbonate platform on which these units accumulated.

Lost Burro Formation

The oldest Paleozoic rocks exposed in the area, which consist of an incomplete thickness of 50 to 60 m of massive, pale-yellow to medium-gray dolomite, are herein assigned to the Middle and Late Devonian Lost Burro Formation (Dlb). These rocks crop out near the west end of the map area, on the steep south slope of the first canyon north of Warm Spring Canyon. The lowest rocks are intruded by tonalite (unit KJt). Assignment of these rocks to the Lost Burro Formation is based on their general lithologic similarity to rocks included in the Lost Burro Formation in other parts of the Death Valley region, including the type section of McAllister (1952) in the Cottonwood Mountains about 100 km to the northwest (fig. 1), as well as on their stratigraphic position below dark limestone that can be definitively assigned to the Early Mississippian Tin Mountain Limestone (Mtm). The uniform sequence of massive light-colored dolomite present in the map area differs from typical Lost Burro Formation, which generally consists of well-bedded limestone, dolomite, and minor amounts of quartzite and commonly is color banded. Too little of the section is exposed in the map area, however, to make a meaningful comparison with other areas.

Tin Mountain Limestone

The Lost Burro Formation is overlain by about 50 m of dark-gray cherty limestone that we herein assign to the Early Mississippian Tin Mountain Limestone (Mtm). This unit is characterized by limestone in beds 10 to 40 cm thick, as well as by the presence of brown-weathering chert nodules that compose about 25 percent of the rock. Many limestone beds contain crinoid columnals, and some contain syringoporid corals, horn corals, and sparse brachiopods (see, for example, fossil locality 1; table 2). The dark-gray limestone, brown-weathering chert, and fossil content are characteristic of the Tin Mountain Limestone throughout the Death Valley region, including the type section in the Cottonwood Mountains (McAllister, 1952). The 50-m thickness in the map area, however, is less than the general thickness of about 105 to 145 m in the region (Stevens and others, 1979).

Stone Canyon Limestone

The Tin Mountain Limestone is gradationally overlain by about 60 m of dark-gray, very cherty limestone, herein assigned to the Early Mississippian Stone Canyon Limestone (Msc), a unit named by Stevens and others (1996) for rocks previously referred to as the carbonate facies of the Perdido Formation (Stevens and others, 1979). The sequence in the map area is characterized by the presence of abundant black-chert beds 1 to 5 cm thick that are interbedded with limestone beds 5 to 20 cm thick. The abundance of bedded chert is the main distinguishing feature of the Stone Canyon Limestone at its type area in the northern Argus Range (fig. 1) and elsewhere in the Death Valley region (Stevens and others, 1979, 1996).

One somewhat unusual feature of the Stone Canyon Limestone in the map area is the presence of a matrix-supported conglomerate, here interpreted as a debris-flow deposit, near the easternmost exposure of the formation. This bed, which is several meters thick, is composed of siltstone and limestone pebbles in a calcareous matrix. Similar debris-flow deposits are known in the Stone Canyon Limestone 55 to 75 km to the northwest, in the southern Darwin Hills and the Santa Rosa Hills (fig. 1) (Stevens and others, 1995).

Table 2. Fossils identified from Paleozoic units in Warm Spring Canyon area.

Map Locality	Map Unit	Field Locality	Fossils	Age	References
1	Mtm	792	Corals: Syringoporid Brachiopods	Middle to late Paleozoic	Wrucke, 1966; this study[1]
2	PIPbs	276	Corals: Unidentified colonial coral Gastropods: *Omphalotrocus whitneyi*	Early Permian (Wolfcampian)	Wrucke, 1966; this study[1]
3	PIPbs	277	Brachiopods: *Marginiferidae* *Neospirifer*? sp.	Late Paleozoic	Wrucke, 1966
[2]4	PIPbs	S-0382	Fusulinids: *Triticites buttensis* *Triticites* aff. *T. californicus* *Triticites cellamagnus* *Triticites gigantocellus*	Latest Pennsylvanian	Stevens and Stone, 2007; this study[1]
		705	Fusulinids	Early Permian (middle Wolfcampian)	Wrucke, 1966; Stone, 1984; Stevens and Stone, 2007; this study[1]
		S-1930	Fusulinids: *Pseudoschwagerina roeseleri*		
		S-1958 04-WS-2	Fusulinids: *Pseudoschwagerina roeseleri*		
		S79-WS-3	Fusulinids: *Schwagerina aculeata*?		
		SJS-684	Corals: *Protowentzelella noinskyi*		
		SJS-690	Corals: *Tschussovskenia connorsensis*		
		SJS-691	Corals: *Tschussovskenia connorsensis*		
		SJS-698	Corals: *Tschussovskenia connorsensis*		
		SJS-710	Corals: *Protowentzelella kunthi*		
		SJS-776	Corals: *Protowentzelella hyporiphaea*		
		SJS-801	Corals: *Tschussovskenia* cf. *T. captiosa*		
		SJS-819 S-1969	Corals: *Protowentzelella cystosa* Fusulinids: *Pseudoschwagerina uddeni*		
		SJS-819b	Corals: *Protowentzelella cystosa*		
		SJS-819c	Corals: *Fomichevella*? sp.		
		SJS-819d	Corals: *Protowentzelella cystosa*		
5	PIPbs	292 SJS-1316 S-1967	Corals: *Tschussovskenia connorsensis* Fusulinids: *Triticites gigantocellus*	Latest Pennsylvanian	Wrucke, 1966; Stevens and Stone, 2007; this study[1]
6	PIPbs	672	Corals: Syringoporid	Middle to late Paleozoic	Wrucke, 1966
7	PIPbs	676 S-1910 03-WS-1	Fusulinids: *Triticites californicus*?	Latest Pennsylvanian?	Wrucke, 1966; Stevens and Stone, 2007
8	PIPbs	677	Fusulinids	Permian or Pennsylvanian	Wrucke, 1966; this study[1]
9	PIPbs	299 442	Fusulinids	Permian or Pennsylvanian	Wrucke, 1966
10	Msr	706	Corals: Syringoporid	Middle to late Paleozoic	This study[1]
11	PIPbs	507	Fusulinids Corals: Syringoporid	Permian or Pennsylvanian	Wrucke, 1966
12	PIPbs	673	Gastropods: *Omphalotrocus whitneyi*	Early Permian	Wrucke, 1966
13	Povl	678	Gastropods: Euphemitid	Late Paleozoic	Wrucke, 1966

Table 2. Fossils identified from Paleozoic units in Warm Spring Canyon area.—Continued

Map Locality	Map Unit	Field Locality	Fossils	Age	References
14	Povl	674	Brachiopods	Late Paleozoic?	Wrucke, 1966; this study[1]
15	Povu	300	Ammonoids	Early to Middle Permian?	Wrucke, 1966; this study[3]
16	Povu	444	Ammonoids	Early to Middle Permian?	Wrucke, 1966; this study[3]
17	Povu	506a	Ammonoids	Early to Middle Permian?	Wrucke, 1966; this study[3]
18	PIPbs	S79-WS-2 (USGS loc. 28009-PC)	Conodonts: *Declinognathodus* cf. *D. noduliferus* *Idiognathoides* or *Idiognathodus* *Rachistognathus primus*	Early Pennsylvanian (early Morrowan)	Stone, 1984[4]
19	PIPbs	82-WS-4	Small fusulinids	Pennsylvanian?	Stone, 1984
20	Povu	S-1841	Conodonts: *Mesogondolella simulata*	Early Permian (Wolfcampian)	This study[5]
21	Povu	82-WS-13	Fusulinids: *Parafusulina*? sp.	Early Permian (Leonardian?)	Stone, 1984; this study[1]
[2]22	PIPbs	S79-WS-20	Fusulinids: *Leptotriticites californicus* *Leptotriticites warmspringensis*? *Leptotriticites* sp. 1[6] *Leptotriticites* sp. 3[6] *Pseudoschwagerina uddeni* *Pseudoschwagerina* cf. *P. gerontica*	Early Permian (middle Wolfcampian)	Stone, 1984; Stevens and Stone, 2007; this study[1]
		SJS-1311 S-1965	Fusulinids: *Leptotriticites warmspringensis*? *Eoparafusulina linearis* *Pseudoschwagerina arta* *Pseudoschwagerina* sp. 1[6] *Schwagerina* sp. 4[6] Corals: *Protowentzelella* sp.		
23	PIPbs	82-WS-14	Fusulinids: *Leptotriticites californicus* *Leptotriticites warmspringensis*	Early Permian (middle Wolfcampian)	Stone, 1984; Stevens and Stone, 2007; this study[1]
		SJS-1310 S-1966	Fusulinids: *Leptotriticites californicus* *Leptotriticites warmspringensis* *Eoparafusulina linearis* *Pseudoschwagerina*? sp. *Stewartina* sp. Corals: *Paraheritschioides stevensi* Large solitary caninoid		
24	PIPbs	82-WS-2	Fusulinids: *Eoparafusulina linearis* *Schwagerina* sp. 4[6]	Early Permian (middle Wolfcampian)	Stone, 1984; Stevens and Stone, 2007; this study[1]
25	Povu	S-1701	Fusulinids: *Parafusulina*? aff. *P. bakeri*	Early Permian (Leonardian?)	This study[1]
26	PIPbs	S-1959 04-WS-5	Fusulinids: *Triticites cellamagnus* *Leptotriticites* sp. Corals: *Tschussovskenia* sp.	Latest Pennsylvanian	Stevens and Stone, 2007; this study[1]
		S-1960 04-WS-6	Fusulinids: *Triticites cellamagnus* *Triticites* aff. *T. californicus* *Leptotriticites* sp. Corals: *Tschussovskenia* sp.		
27	PIPbs	WSC-18	Fusulinids: *Pseudoschwagerina*? sp.	Early Permian (Wolfcampian)?	This study[7]
28	Povl	WSC-37	Fusulinids: *Eoparafusulina*? sp. Corals: Large solitary caninoids	Early Permian (middle Wolfcampian)	This study[1]

[1]Fossils identified by C.H. Stevens.
[2]See figure 6, which shows stratigraphic columns and positions of field localities.
[3]Ammonoids examined by B.F. Glenister, University of Iowa.
[4]Conodonts identified by J. Repetski, U.S. Geological Survey.
[5]Conodonts identified by S.M. Ritter, Brigham Young University.
[6]Species are of Stevens and Stone (2007).
[7]Fusulinids identified in field by P. Stone.

The 60-m thickness of the Stone Canyon Limestone in the map area is less than elsewhere in the region: the formation is more than 400 m thick in the Santa Rosa Hills and Darwin Hills; 107 m thick in its type area in the northern Argus Range; and about 300 m thick at Tucki Mountain (fig. 1) (Stevens and others, 1979, 1996).

Santa Rosa Hills Limestone

The Stone Canyon Limestone is overlain gradationally by about 100 m of massive light-gray limestone, herein assigned to the Early and Late Mississippian Santa Rosa Hills Limestone (Msr). The limestone is recrystallized, commonly is coarse grained, and has little evidence of bedding. The only fossils noted are large but sparse colonies of robust syringoporid corals (see, for example, fossil locality 10; table 2), as well as locally abundant crinoid columnals. Large chert nodules are present in the lowermost 10 m, but chert is rare higher in the section.

The Santa Rosa Hills Limestone was named by Dunne and others (1981) for exposures in the Santa Rosa Hills (fig. 1) but is recognized throughout a large part of the Death Valley region (Stevens and others, 1996). Rocks assigned to this formation in most areas range in thickness from about 75 to 200 m (Stevens and others, 1979). The approximate 100-m thickness determined in this study supersedes the 235-m thickness reported by Stevens and others (1979).

Indian Springs Formation

The Santa Rosa Hills Limestone is sharply overlain by about 6 m of mostly brown-weathering calcareous siltstone and dark-gray shale that contains minor amounts of dark-gray crinoidal limestone. This thin but continuous unit is herein assigned to the Late Mississippian Indian Springs Formation (Mis), the type section of which is in southern Nevada (Webster and Lane, 1967). Similar sequences of clastic strata, commonly 10 to 30 m thick, have been assigned to the Indian Springs Formation in the Santa Rosa Hills and Darwin Hills (fig. 1) (Stone and others, 1989) and in the Cottonwood Mountains (Snow, 1990).

Regional Framework of the Devonian and Mississippian Rocks

The Devonian and Mississippian dolomite and limestone units at Warm Spring Canyon accumulated on the broad carbonate platform that covered much of eastern California and southern Nevada (Stevens and others, 1995). The platform paralleled the southwest-trending continental margin and generally deepened to the northwest. On the basis of their geographic location, rocks in the Warm Spring Canyon area evidently represent the inner part of the carbonate platform as defined by Stevens and others (1995), although some features of the rocks, particularly the debris-flow deposit in the Stone Canyon Limestone and the absence of dolomite in the upper part of the sequence, are atypical of the inner platform. Carbonate sedimentation ended in the Late Mississippian when the platform was subaerially exposed, probably due to a eustatic drop in sea level (Stevens and others, 1995). Deposition of the thin, shallow-water clastic sediments of the Indian Springs Formation followed the end of carbonate sedimentation.

The Mississippian limestone sequence in the map area is thinner than in areas to the northwest, such as the Argus Range and Cottonwood Mountains (Stevens and others, 1979). The total thickness in the map area (about 210 m), however, is similar to the 230-m thickness of the correlative Monte Cristo Limestone in the Providence Mountains, 180 km to the southeast (fig. 1) (Hazzard, 1954). The 110-m combined thickness of the Dawn and Anchor Members of the Monte Cristo Limestone is the same as that of the correlative Tin Mountain and Stone Canyon Limestones at Warm Spring Canyon. In addition, the 120-m combined thickness of the Bullion and Yellowpine Members of the Monte Cristo Limestone is comparable to the 100-m thickness of the correlative Santa Rosa Hills Limestone. These similarities in thickness evidently reflect the approximately uniform rates of subsidence and sedimentation across a large part of the Mississippian carbonate platform, from the vicinity of Warm Spring Canyon southeastward into the Mojave Desert.

Pennsylvanian and Permian Rocks

Pennsylvanian and Permian rocks form a belt of almost continuous exposure about 5 km long in Warm Spring Canyon. These rocks have been studied previously as part of a regional paleogeographic investigation (Stone, 1984; Stone and Stevens, 1984); in addition, parts of the section, particularly those containing fossils, were critically reexamined during the present study. Most of the discussion presented here is based on observations made and fossils collected in the western part of the outcrop belt. Rocks in the eastern part of the belt are more highly metamorphosed and have not been studied as thoroughly.

Bird Spring Formation

The lower part of the Pennsylvanian-Permian section in Warm Spring Canyon is composed primarily of medium- to dark-gray, thick-bedded, fine-grained limestone that is, in part, cherty and locally fossiliferous. This limestone unit, which conformably overlies the Indian Springs Formation, is herein assigned to the Pennsylvanian and Early Permian Bird Spring Formation (PIPbs), the type section of which is in the Spring Mountains area of southern Nevada (Hewett, 1931) (fig. 1). The rocks in Warm Spring Canyon generally are similar in lithology and age to rocks that have been assigned to the Bird Spring Formation throughout eastern California, from the Nopah Range on the north to the Providence Mountains on the south (see, for example, Hazzard, 1937; Hewett, 1956; Burchfiel and others, 1983; Stevens and Stone, 2007). Regionally, rocks of the Bird Spring Formation in California are here interpreted to have formed as shallow-water carbonate-platform deposits, which contrast with the coeval, deeper water gravity-flow deposits (mostly assigned to the Keeler Canyon Formation) in the Inyo Mountains and other areas 60

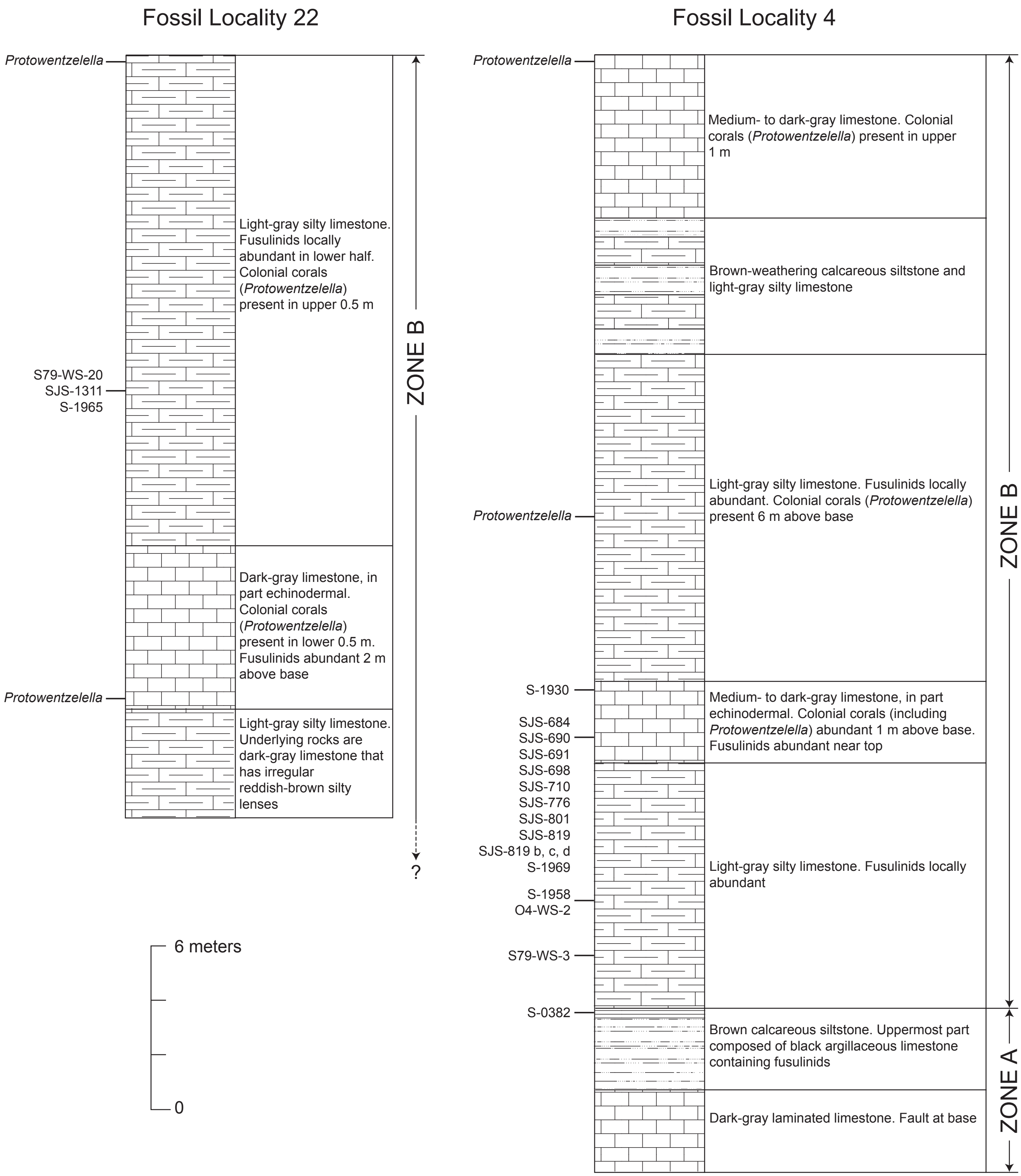

Figure 6. Stratigraphic columns for fossil localities 4 and 22 in uppermost part of Bird Spring Formation, showing stratigraphic positions of fossil samples, as well as of additional occurrences of colonial coral *Protowentzelella*, in relation to faunal zones A and B. See table 2 for fossil names and ages.

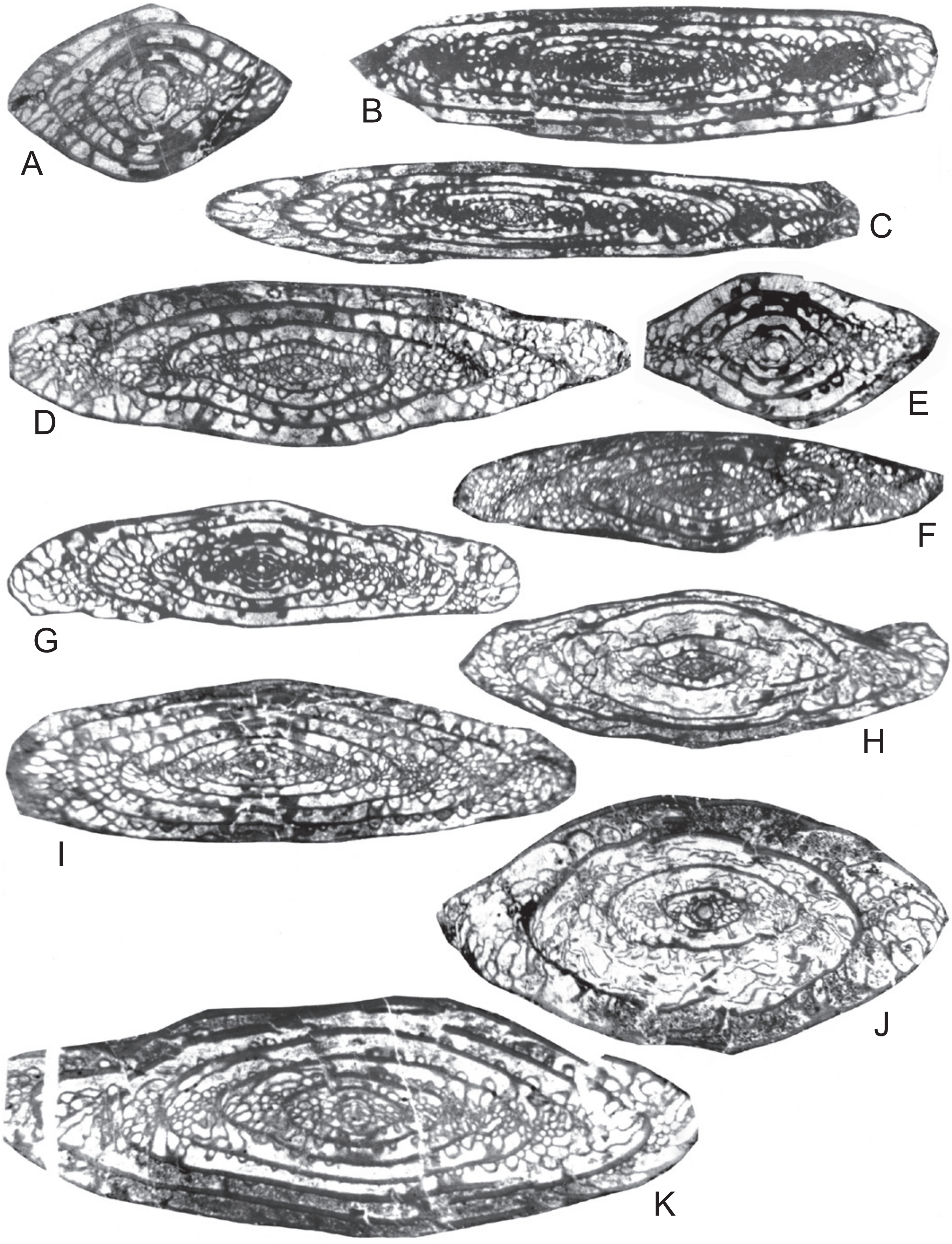

Figure 7

Figure 7. Fusulinids from upper part of the Bird Spring Formation at Warm Spring Canyon (all figures are axial sections, ×10). Specimens labeled USNM are housed in National Museum of Natural History (formerly U.S. National Museum); specimens labeled SJS are housed in San Jose State University Museum of Paleontology, San Jose, California. A, *Triticites gigantocellus* Stevens and Stone; fossil locality 4, sample S-0382, specimen USNM 531295. B, C, *Eoparafusulina linearis* (Dunbar and Skinner); fossil locality 24, sample 82-WS-2, specimens SJS 211f and SJS 212f, respectively. D, *Schwagerina* sp. 4 of Stevens and Stone (2007); fossil locality 24, sample 82-WS-2, specimen SJS 188f. E, *Triticites cellamagnus* Thompson and Bissell; fossil locality 4, sample S-0382, specimen SJS 266f. F, *Schwagerina* sp.; fossil locality 22, sample S-1965, specimen SJS 267f. G, *Leptotriticites californicus* Stevens and Stone; fossil locality 23, sample 82-WS-14, specimen USNM 531312. H, *Pseudoschwagerina* sp. 1 of Stevens and Stone (2007); fossil locality 22, sample S-1965, specimen SJS 184f. I, *Leptotriticites warmspringensis*? Stevens and Stone; fossil locality 22, sample S-1965, specimen SJS 268f. J, *Pseudoschwagerina* cf. *P. gerontica* Dunbar and Skinner; fossil locality 22, sample S79-WS-20, specimen SJS 269f. K, *Stewartina* sp.; fossil locality 23, sample S-1966, specimen SJS 270f.

to 130 km to the north and northwest (fig. 1) (Stone, 1984; Stone and Stevens, 1984; Stevens and others, 2001).

The lower part of the Bird Spring Formation at Warm Spring Canyon primarily consists of dark-gray limestone that contains abundant nodules, lenses, and thin beds of dark-brown-weathering chert. Some of the chert nodules are spherical to ovoid; such chert nodules are common in the "golfball beds" that characterize the lower part of the Pennsylvanian limestone section throughout the region (Stone, 1984). Beds higher in the formation contain less chert, generally are lighter in color, and are more fossiliferous than the lower beds.

The Bird Spring Formation is transected by a steep, east-west-striking fault that is particularly well exposed in the vicinity of cross section *B–B'*. Because of this fault, a complete section of the formation is not exposed. About 225 m of section is exposed north of the fault; 300 m is exposed south of it. Stratigraphic comparison with a thick, continuous section of Bird Spring Formation exposed at Striped Butte, west of the map area (fig. 2), suggests that the fault in Warm Spring Canyon cuts out several hundred meters of strata (Stone, 1984).

Rocks of the Bird Spring Formation at Warm Spring Canyon are here interpreted as moderately shallow carbonate-shelf deposits similar to those that typify this widespread formation throughout its extent (Stone, 1984; Stone and Stevens, 1984; Stevens and Stone, 2007). This interpretation is based on the thick to massive bedding and the fine-grained micritic texture of most of the limestone, as well as the local abundance of fossils, especially colonial corals in the upper part of the formation, and the absence of evidence that suggests significant transport of sediments or fossils prior to deposition.

The Bird Spring Formation at Warm Spring Canyon ranges in age from Early Pennsylvanian (early Morrowan) to Early Permian (middle Wolfcampian), on the basis of conodonts from the lowermost beds at fossil locality 18 (table 2; see also, Stone, 1984) and fusulinids and corals from numerous localities in the upper beds (Stone, 1984; also, this study). Stevens and Stone (2007) described some of the fusulinids as part of a regional investigation of the Bird Spring Formation. Most of the lower part of the formation lacks datable fossils.

We recognize two faunal zones in the upper part of the Bird Spring Formation that contain substantially different fusulinids and corals (figs. 6, 7, 8). The lower faunal zone (Zone A; see fig. 6), which includes beds from about 150 to 35 m below the top of the formation, is represented at fossil localities 4 (table 2; sample S-0382; see fig. 6), 5, and 26 (table 2; see fig. 3 for all locations). Zone A is characterized by species of the fusulinid genera *Triticites* (particularly *T. cellamagnus*) and *Leptotriticites*, as well as by the colonial coral *Tschussovskenia* (figs. 7, 8). The fusulinids of Zone A indicate a latest Pennsylvanian age, as interpreted by Stevens and others (2001).

The higher faunal zone (Zone B; see fig. 6), which includes the upper approximately 35 m of the Bird Spring Formation, also is represented at fossil locality 4 (all samples higher in section than S-0382; see fig. 6; table 2), as well as at fossil locality 22 (table 2; fig. 6; see fig. 3 for location). At locality 4, light-gray limestone at the base of Zone B sharply overlies soft black limestone of sample S-0382, which marks the top of Zone A. Zone B at locality 4 is characterized primarily by the fusulinid *Pseudoschwagerina* and by colonial corals that include *Tschussovskenia* and several species of *Protowentzelella* (figs. 7, 8). The latter coral is present most abundantly in a single bed 25 m below the top of the Bird Spring Formation. At locality 22, which is 3.5 km west of locality 4, Zone B is characterized by locally abundant *Protowentzelella* in association with the fusulinids *Leptotriticites*, *Pseudoschwagerina*, and rare *Eoparafusulina* (fig. 7). Fusulinids from Zone B indicate an Early Permian (middle Wolfcampian) age as defined by Magginetti and others (1988). The boundary between Zones A and B constitutes a major faunal break.

Fusulinids and corals also are abundant at fossil localities 23 and 24 (table 2; see fig. 3 for location), which are located in a narrow fault sliver of limestone herein assigned to the Bird Spring Formation that lies structurally between the lower and upper units of the Owens Valley Group, near locality 22. Fusulinid genera represented at these localities include *Leptotriticites*, *Eoparafusulina*, *Stewartina*, and *Pseudoschwagerina*. In addition, a large caninoid solitary coral is abundant at locality 23, along with the colonial coral *Paraheritschioides stevensi*. The fusulinids generally are comparable to those at locality 22 and are here interpreted to represent Zone B of middle Wolfcampian age, although the abundant *Eoparafusulina* and large caninoid corals that characterize localities 23 and 24 are not present in Zone B at either locality 22 or locality 4.

Faunal Zones A and B of this study correspond to Fusulinid Zones 2 and 3 of Stevens and Stone (2007), respectively, which are recognized throughout the regional extent of the Bird Spring Formation in east-central and southeastern California.

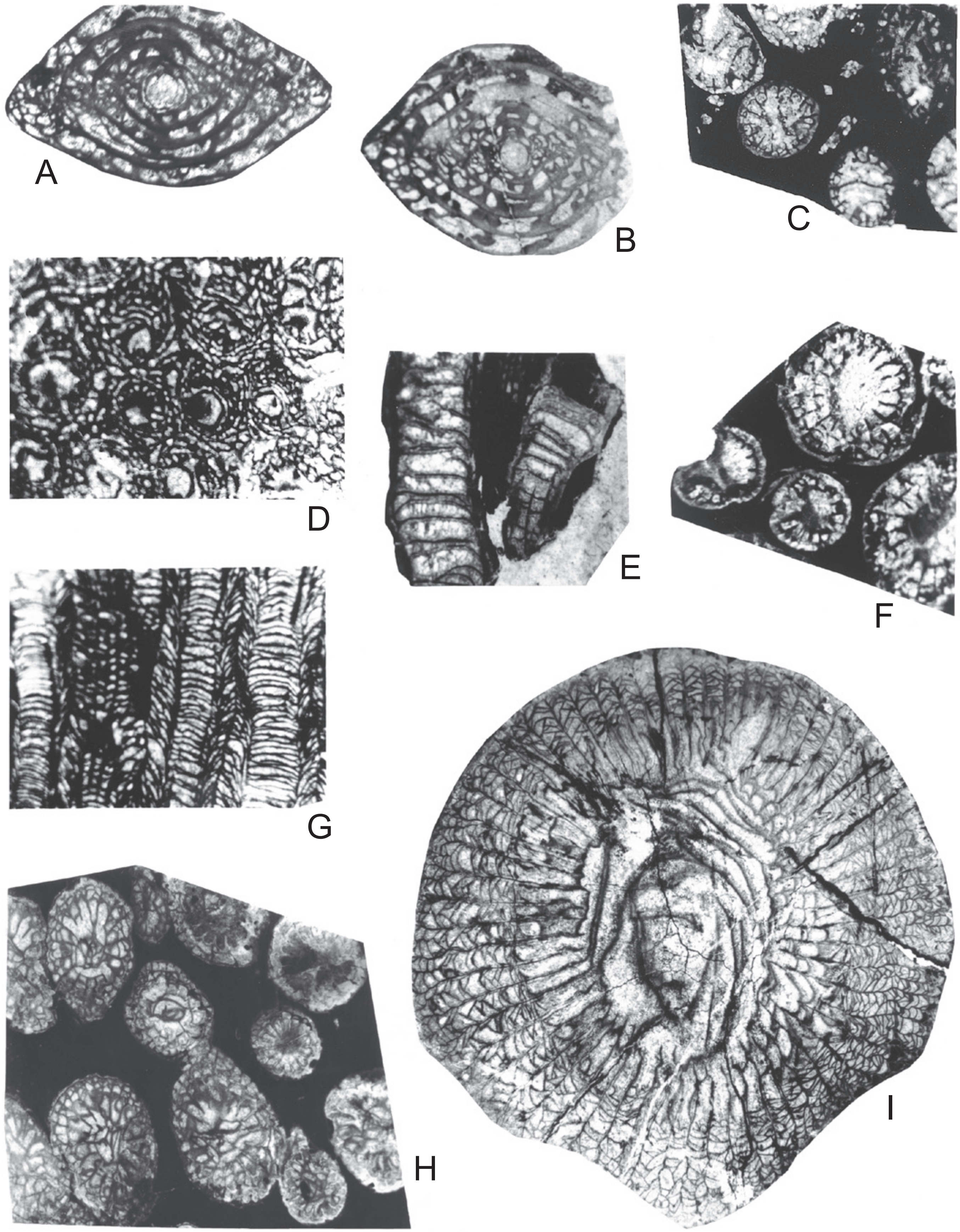

Figure 8

Figure 8. Fusulinids (A, B) and corals (C, D, E, F, G, H, I) from upper part of the Bird Spring Formation at Warm Spring Canyon (figures are ×3, except where noted). Specimen labeled USNM is housed in National Museum of Natural History (formerly U.S. National Museum); specimens labeled SJS are housed in San Jose State University Museum of Paleontology, San Jose, California. A, B, *Triticites gigantocellus* Stevens and Stone (axial sections, ×10); fossil locality 4, sample S-0382, specimens USNM 531296 and SJS 271f, respectively. C, E, F, *Tschussovskenia connorsensis* (Easton)? (C, F, transverse sections; E, longitudinal section); fossil locality 4, sample SJS-698, specimens SJS 221c, 223c, and 224c, respectively. D, G, *Protowentzelella cystosa* (Dobrolyubova) (D, transverse section; G, longitudinal section); fossil locality 4, sample SJS-819b, specimens SJS 222c and SJS 225c, respectively. H, *Paraheritschioides stevensi* (Wilson) (transverse section); fossil locality 23, sample SJS-1310, specimen SJS 226c. I, Unidentified solitary caninoid coral (transverse section, ×2); fossil locality 23, sample SJS-1310, specimen SJS 227c.

Several taxa identified in the carbonate-shelf deposits of the Bird Spring Formation at Warm Spring Canyon also provide a tie to the Early Permian calcareous turbidites and debris-flow deposits of the Owens Valley Group (Stone and others, 1987, 1989), exposed in areas 50 to 80 km to the north and northwest. Probably the strongest tie is with beds at Darwin Canyon (fig. 1) that contain the corals *Protowentzelella*, *Tschussovskenia*, and *Paraheritschioides stevensi*; large caninoid corals; and an advanced form of the fusulinid *Eoparafusulina*. These beds, which include a widespread debris-flow deposit commonly 25 to 30 m thick, are assigned to the Millers Spring Member of the Darwin Canyon Formation (Stone and others, 1987, 1989; Magginetti and others, 1988; Stevens and others, 1989). Similar correlative rocks and fossils have been described at Marble Canyon in the Cottonwood Mountains (Stone, 1984) and near Conglomerate Mesa (Magginetti and others, 1988; Stone and others, 1989) (fig. 1).

Owens Valley Group

The primarily siliciclastic upper part of the Pennsylvanian-Permian section at Warm Spring Canyon is herein assigned to the Owens Valley Group, which was named as a formation by Merriam and Hall (1957) and which has a type section in the southern Inyo Mountains 120 km to the northwest (fig. 1). Later stratigraphic revisions raised the Owens Valley Formation to group rank throughout the Inyo Mountains–Darwin Canyon–Argus Range region and established several new formations to make up the group (Stone and Stevens, 1987; Stone and others, 1987). Rocks herein assigned to the Owens Valley Group at Warm Spring Canyon are lithologically similar to and correlative with some of the rocks at Darwin Canyon and in the Argus Range.

We divide the Owens Valley Group at Warm Spring Canyon into two unnamed units. The lower unit, which averages about 90 m in thickness, consists primarily of light- to dark-gray, thin- to medium-bedded, unfossiliferous to sparsely fossiliferous limestone that contains abundant discontinuous beds and lenses of buff- to brown-weathering siliceous rock 3 to 7 cm thick. The siliceous layers give typical outcrops a distinctive, darkly banded appearance. The relatively thin bedding, siliceous composition, and scarcity of fossils suggest a somewhat deeper water environment than that of the shelf limestones of the underlying Bird Spring Formation. The base of the lower unit is marked by a conspicuous bed of brown calcareous siltstone that averages about 5 m thick. In unfaulted stratigraphic sections, this bed sharply but concordantly overlies the uppermost fossiliferous limestone of the Bird Spring Formation.

The upper unit of the Owens Valley Group consists of about 565 m of buff- to brown-weathering calcareous siltstone, fine-grained sandstone, and argillite. Many of these rocks exhibit low-angle cross-lamination, convolute lamination, and locally graded bedding suggestive of turbidite deposition. Subordinate sequences of medium-gray, thin-bedded, fine-grained limestone and silty limestone, also interpreted as turbidites, are most abundant in the uppermost 200 m of the sequence. Locally, the lower 50 to 75 m of the upper unit contains thin- to medium-bedded siliceous limestone similar to that in the lower unit.

Our mapping indicates that the contact between the lower and upper units of the Owens Valley Group probably is faulted everywhere in the Warm Spring Canyon area, so a continuous depositional sequence is not preserved. The nature of the original depositional contact is, thus, uncertain. However, the similarity in lithology between the siliceous limestone locally present in the lower part of the upper unit to that in the lower unit suggests that the contact may be gradational.

The age of the Owens Valley Group at Warm Spring Canyon is problematic because of a paucity of fossils. Most of the fossils we have observed in the lower unit are small, poorly preserved brachiopods and gastropods that do not provide the basis for a precise age interpretation (fossil localities 13, 14; table 2; see fig. 3 for locations). The only other fossils we have observed in rocks assigned to the lower unit are solitary caninoid corals and poorly preserved fusulinids at fossil locality 28 (table 2; see fig. 3 for location), which is in an area of complex faulting. These fossils are found in silty limestone about 3 m above a faulted contact with the uppermost limestone bed of the Bird Spring Formation. The fossils are similar to those in the fault sliver of Bird Spring Formation at fossil locality 23 and are here interpreted to represent about the same age (middle Wolfcampian). This similarity suggests that rocks in the fault sliver at locality 23 represent the uppermost part of the Bird Spring Formation.

Ammonoid impressions and casts are present at fossil localities 15, 16, and 17, just above the faulted base of the upper unit of the Owens Valley Group (table 2; see fig. 3 for locations). Wrucke (1966) distinguished two genera (fig. 9) suggestive of a Leonardian or Guadalupian (late Early or Middle Permian) age; because of poor preservation, however, neither genus was positively identified. More recently, Wrucke's specimens were examined by B.F. Glenister (written and oral commun., 2002), who could not identify the specimens to genus but suggested that they resembled Wordian (middle

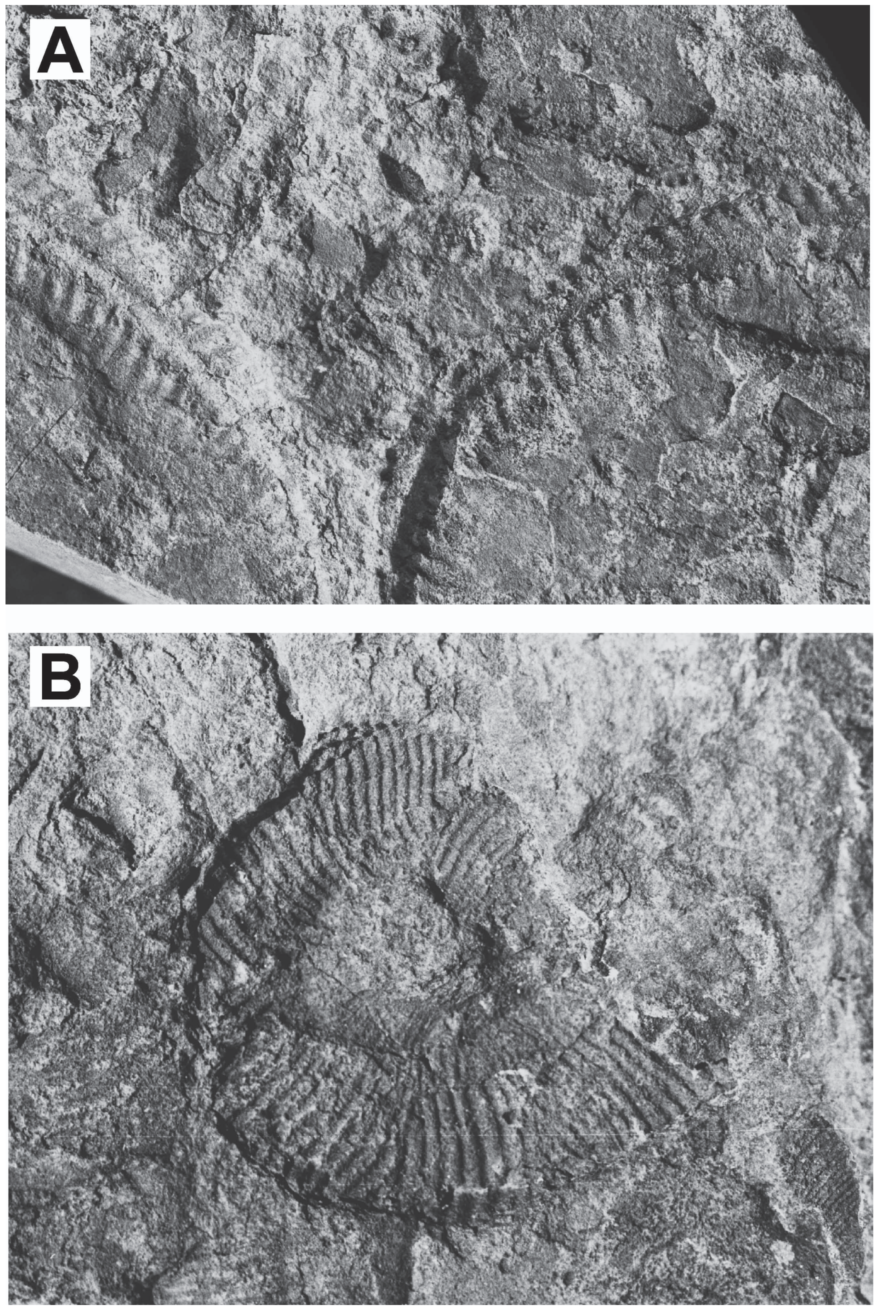

Figure 9. Casts of ammonoids from fossil locality 16 in upper unit of the Owens Valley Group at Warm Spring Canyon. A, x1.4; B, x2.5. Photographs reproduced from figs. 34 and 35 of Wrucke (1966).

Guadalupian) forms. Regional stratigraphic and paleogeographic considerations, however, make it unlikely that any rocks in this deep-water sequence are as young as Wordian (Stone and others, 2000). [Also, see addendum.]

The only other fossils that we have recovered from the upper unit are two fusulinid specimens and a single conodont (fossil localities 20, 21, and 25; table 2; see fig. 3 for locations). The fusulinids probably are species of *Parafusulina* consistent with a Leonardian age, which we consider to be the most probable age of deposition. S.M. Ritter (written commun., 2003) identified the conodont as *Mesogondolella similaris*, a late Asselian to Sakmarian (middle to late Wolfcampian) form that evidently is reworked.

Our assignment of these rocks at Warm Spring Canyon to the Owens Valley Group is based on their general lithologic similarity to some of the rocks that make up Early Permian formations of the Owens Valley Group in Darwin Canyon and the northern Argus Range to the northwest (fig. 1) (Stone and others, 1987). Probably the closest lithostratigraphic match is with the Panamint Springs Member of the Darwin Canyon Formation. That unit, like the sequence at Warm Spring Canyon, primarily is composed of thin-bedded siliciclastic turbidites and largely lacks coarse-grained bioclastic limestone turbidites and debris-flow beds. In addition, as noted above, the underlying Millers Spring Member of the Darwin Canyon Formation apparently is coeval with the upper part of the Bird Spring Formation at Warm Spring Canyon, consistent with a correlation between the Panamint Springs Member and rocks assigned to the Owens Valley Group at Warm Spring Canyon. Given the lack of more specific evidence for correlation, however, we prefer to use the general name Owens Valley Group for the rocks at Warm Spring Canyon, rather than extend formation and member names from Darwin Canyon into this area.

Regional Framework of the Pennsylvanian and Permian Rocks

Pennsylvanian and Early Permian rocks in the Death Valley region record a complex evolution from carbonate-shelf to basinal depositional environments (Stone and Stevens, 1988). From Early to early Middle Pennsylvanian time, a gently northwest-sloping carbonate shelf or ramp covered the entire Death Valley region, including the Warm Spring Canyon area (Stone and Stevens, 1988). This shelf or ramp is represented by dark-gray cherty limestone of the lowermost Bird Spring Formation and its correlative units to the northwest. In the late Middle Pennsylvanian, a shelf margin that persisted into the earliest Permian developed east of the Argus Range and the northern Cottonwood Mountains (fig. 1). Carbonate sediment was transported westward and northwestward downslope across the shelf margin to feed a submarine-fan system, which is represented by the limestone turbidites of the Keeler Canyon Formation that are widely exposed in the Inyo Mountains region (fig. 1) (Stevens and others, 2001). Late Pennsylvanian and Early Permian shelf deposits to the east are represented by thick-bedded to massive limestone of the Bird Spring Formation in the southern Cottonwood Mountains (Panamint Butte) and the southern Panamint Range (Warm Spring Canyon and Striped Butte) (figs. 1, 2).

In the middle Wolfcampian, the outer part of the carbonate shelf and slope (Argus Range and southern Cottonwood Mountains) subsided, and basinal deposition of the Owens Valley Group began (Stone and Stevens, 1988). At Warm Spring Canyon, shallow-water shelf conditions persisted longer, as shown by fusulinids and corals in beds at the top of the Bird Spring Formation, which are correlative with the deeper water deposits to the northwest. The overlying, deeper water sequence of the Owens Valley Group indicates a period of rapid subsidence in late Wolfcampian or Leonardian time and the demise of the carbonate shelf.

Jurassic and Cretaceous Rocks

Warm Spring Formation

The Warm Spring Formation (units Jws, Jwt, Jwa), named by Johnson (1957), forms a narrow belt of metavolcanic rocks and minor amounts of metasedimentary rocks in the southwestern part of the map area. It is a remnant from a thick stratigraphic section that was metamorphosed and penetrated by Late Jurassic or Early Cretaceous monzogranite (unit KJm). The formation lies near the east edge of the Mesozoic arc of volcanic and intrusive rocks that extends across the miogeoclinal platform from the Mojave Desert northwestward into northern California.

A unit composed of fine-grained sandstone and other rocks (unit Jws) forms the lowest part of the Warm Spring Formation. In addition to the sandstone, this unit contains calc-silicate rock of grossular garnet, scapolite, and vesuvianite, which represent original calcareous layers, as well as metamorphosed argillaceous rocks containing sillimanite and andalusite. Local aphanitic lenses may be silicic volcanic rock. The unit also contains locally abundant irregular bodies of fine- to medium-grained granitic rocks that have not been mapped separately from the host metamorphic rocks. The unit, which is as much as 100 m thick, is absent at the west border of the area. Locally overlying the sandstone, especially in the central part of the outcrop belt of the formation, is a conspicuous hornfels, as thick as 30 m, that is interpreted to be a metamorphosed tuff (unit Jwt). It contains light-gray, angular to lens-shaped fragments, commonly about 1 cm across, that resemble partly flattened pumice lapilli, all set in a medium-gray aphanitic groundmass. The majority of the Warm Spring Formation, at least 350 m thick and lying above the sandstone and tuff, consists of dark-gray, almost black, meta-andesite flows (unit Jwa), in which layering is obscure.

The age of the Warm Spring Formation is uncertain. The formation rests in apparent fault contact with the underlying Permian Owens Valley Group in the map area, but only a few kilometers to the west it stratigraphically overlies the Triassic Butte Valley Formation, which is at least 1,200 m thick (Johnson, 1957; Cole, 1986). Fossils of Early Triassic age (Johnson, 1957) in the Butte Valley Formation provide a maximum age for the Warm Spring Formation. The meta-andesite in the formation resembles rocks in the Soda Mountains, 65 km to the south (fig. 1), that Grose (1959) considered to be Triassic or Jurassic. The best indication of the age of the Warm Spring Formation, however, is that the meta-andesite and silicic rocks are similar to volcanic sequences in the Inyo Mountains and Alabama Hills (fig. 1) that have U-Pb zircon ages as old as about 170 Ma (Middle Jurassic) (Dunne and Walker, 1993), as well as to volcanic rocks in the Slate Range (fig. 1) that have U-Pb zircon ages as young as 150 to 148 Ma (Late Jurassic) (Dunne and others, 1994).

Plutonic Rocks

A narrow tonalite pluton (unit KJt) separates Proterozoic and upper Paleozoic rocks everywhere in the Warm Spring Canyon area. It emerges from beneath alluvium in Butte Valley and extends east to Warm Spring, where it is overlain by Tertiary volcanic rocks; because the pluton extends beneath cover at its extremities, it may be considerably longer than its exposed length. Its outcrop averages about 0.5 km in width and is as narrow as 60 m or less at one locality north of Warm Spring Canyon, where the contacts are covered by older gravels (unit QToa). The disproportionate length of the pluton compared to its width, as well as its position between rocks of very different ages, is interpreted as resulting from emplacement along the Butte Valley Thrust Fault, discussed later. Biotite and hornblende from the tonalite at localities 3 and 4 (table 3; see fig. 3 for locations) yielded K-Ar ages of 154.7 ± 5 and 147.1 ± 5 Ma, respectively, which together suggest a Late Jurassic to Early Cretaceous age of emplacement.

The tonalite contains mappable inclusions of Paleoproterozoic monzogranite gneiss (unit Ꮲ$_{P}$mg), Neoproterozoic diamictite of the Surprise Member of the Kingston Peak Formation (Ꮲ$_{N}$ksd), and Paleozoic(?) orthoquartzite (unit oq) and fine-grained, banded, biotite- and calc-silicate-bearing quartzite (unit fq). The inclusions are inferred to represent slivers between different strands of the Butte Valley Thrust Fault.

Monzogranite (unit KJm), exposed in the southwest corner of the area, is part of a large granitic mass that continues west from the Warm Spring Canyon area and includes the Manly Peak pluton to the southwest (Johnson, 1957). The granitic mass intruded strata of Permian and Jurassic age, and it also underlies some of these strata at shallow depths as cupolas, as indicated by areas of recrystallized limestone, by the development of calc-silicate minerals locally in the recrystallized rocks, and by the many small exposures of monzogranite in the Permian section. K-Ar ages at localities 1 and 2 (table 3; see fig. 3 for locations) are 151.2 ± 5 Ma (biotite) and 144.1 ± 5 Ma (hornblende), suggesting a Late Jurassic to Early Cretaceous emplacement age, similar to that of the tonalite.

Miocene Volcanic Rocks

Volcanic rocks consisting principally of andesite and basalt flows (unit Tab) and lesser amounts of silicic tuff (unit Tst) crop out extensively in the southeastern part of the map area and at scattered localities in the western half. These rocks constitute the northernmost remnants of a volcanic section many hundreds of meters thick that extends from the southern Panamint Range south to the Owlshead Mountains (fig. 1) (Wagner, 1998; Luckow and others, 2005). In the Warm Spring Canyon area, these rocks are about 400 m thick.

Although these rocks were not studied in detail, field identification suggests that many of the flows are olivine basalts and others are andesitic. Some andesites contain pyroxene and olivine; others have abundant oxyhornblende. Individual flows in the section commonly are 6 to 15 m thick, but a few are as thick as 30 m. Silicic rocks interbedded with the flows appear to be dacitic to rhyolitic in composition and to be in sections as thick as 100 m. Most of the silicic rocks are tuffs. A latite flow close to the mountain front north of Warm Spring Canyon, in the eastern part of the map area, yielded a ^{40}Ar/^{39}Ar age from hornblende of 13.56 ± 0.82 Ma, and an overlying tuff gave a zircon fission-track age of 10.3 ± 0.5 Ma (both ages from Topping, 1993). Wagner (1998) obtained a K-Ar age of 14.0 ± 0.3 Ma from related basalt in the Wingate Wash area, 23 km south of Warm Spring (fig. 1). Luckow and others (2005) reported several ^{40}Ar/^{39}Ar ages that range from about 14 to 12.5 Ma and a tephrochronologic age of about 12 Ma from related volcanic rocks in the Wingate Wash–Owlshead Mountains area, clearly indicating the middle Miocene age of this volcanic assemblage.

The volcanic rocks were deposited on a middle Miocene landscape that was similar in relief to the present-day topography. In the southeastern part of the area, these rocks were deposited in a paleovalley that trended northeast, crossing what later became the modern Warm Spring Canyon, then turned to the east, where the volcanic rocks now are buried beneath Quaternary fanglomerate at the mountain front. Judged by the elevation difference between the highest and lowest positions of the volcanic rocks on its south side, this paleovalley was at least 250 m deep, cutting through the Noonday Dolomite and deep into the underlying Mesoproterozoic section (cross section *D–D'*).

In the western half of the area, scattered exposures of the volcanic section indicate accumulation of the volcanic unit in an ancestral Warm Spring Canyon, as close as about 100 m from the present-day canyon bottom. Remnants of volcanic rocks on the sides and close to the bottom of Warm Spring Canyon show that this ancestral Warm Spring Canyon trended toward the paleovalley in the southeastern part of the area. Exposures of the volcanic rocks north of the Grantham Mine indicate that the ancestral canyon reached the mountain front north of the present-day mouth of Warm Spring Canyon.

In addition to filling valleys, the volcanic rocks were deposited on a highland developed on granitic rocks south of Warm Spring Canyon, as well as possibly on a plateau of Noonday Dolomite represented by an outcrop of volcanic rock near the north-central border of the area. This Miocene highland was 350 to 450 m above the present-day bottom of Warm Spring Canyon.

Surficial Deposits

Bedrock units in the Warm Spring Canyon map area are unconformably overlain by late Cenozoic surficial deposits that primarily consist of alluvial gravels. These deposits are divided herein into five units, distinguished on the basis of their geomorphic and surficial characteristics. They probably range in age from Miocene or Pliocene to Holocene. Landslide and talus deposits of probable Holocene age also have been mapped.

Older Alluvium

Probably the oldest surficial deposits in the map area are mapped as older alluvium (unit QToa), which consists of weakly consolidated, sandy pebble to boulder gravel that crops out in numerous patches on the flanks of Warm Spring Canyon and on a few ridge crests. The gravel, much of which is subrounded to rounded, comprises various rock types, which include quartzite, carbonate rock, schist, granite, and diabase, most or all of which probably could have been derived from sources in the Panamint Range. Most of the numerous patches of this alluvium extend from the floor of Warm Spring Canyon 50 to 100 m up the canyon walls; some extend as much as 200 m above the canyon floor, as observed 1.5 km northwest and 2 km northeast of Warm Spring. The highest outcrop is an isolated patch at the head of Galena Canyon near the north-central border of the map area, 240 m above the floor of Warm Spring Canyon. This outcrop distribution suggests that the older alluvium was originally at least 200 m thick and represents a major aggradation that took place after Warm Spring Canyon and its tributaries had been eroded to about their present depth. The highest patch, because of its great height above the floor of Warm Spring Canyon, may be remnant from an earlier stream system. Erosion subsequent to the aggrading event has exhumed the canyon, leaving remnants of the older alluvium.

The older alluvium unit is not dated, although it clearly is younger than the unconformably underlying middle Miocene (14 to 10 Ma) volcanic rocks (Tab, Tst). Similar alluvial deposits assigned to the Navadu and Nova Formations in the southern Cottonwood Mountains, about 70 km northwest of the map area (fig. 1), are interbedded with tuffs and basalt flows that have been dated at about 12 to 3 Ma (Snow and Lux, 1999), and fanglomerate in Wingate Wash to the south contains an ash thought to be 12 Ma (Wagner, 1998). Thus, our older alluvium unit could be as old as late Miocene or Pliocene in age. As mapped, however, this unit could include some younger deposits as well, so it is assigned a Quaternary-Tertiary age.

Landslide Deposits and Talus

Landslide deposits in the map area include blocky rubble and brecciated fragments of Tertiary andesite and basalt, Mesoproterozoic diabase, and the stromatolite and cherty dolomite members of the Crystal Spring Formation. These deposits, which are mapped as units Qlab, Qld, Qls, and Qlc, respectively, flank the north slope of Warm Spring Canyon 1 to 2 km northwest of the Grantham Mine (fig. 3) and descend steeply to the canyon floor. Another landslide mass (also mapped as unit Qld), composed of blocks of diabase, is located on the south side of Warm Spring Canyon above the Grantham Mine. Finally, a large landslide deposit composed of blocks of Noonday Dolomite is present at the Queen of Sheba Mine; this deposit is mapped as unit Qln. All of the landslide deposits are of presumed Holocene age.

Talus of presumed Holocene age, composed of rubbly debris from various upslope sources, is mapped at three places in the western part of the map area. Smaller, unmapped areas of talus and colluvium are common.

Alluvial-fan and Wash Deposits

Most surficial deposits in the map area are parts of alluvial fans. Fan deposits that flank the east side of the Panamint Range in the map area are divided into two older dissected, heavily varnished, pavement-forming units (QTa_1 and Qa_2), as well as a younger unit (Qa_4) of poorly sorted gravels that cuts into the older units and grades from lightly varnished, inactive deposits to very recent deposits of modern channels. Deposits of unit Qa_4 cover the floors of Warm Spring Canyon and its tributaries. Units QTa_1 and Qa_2 compare closely with deposits that form the regionally distinct geomorphic surfaces Ql and Q2 of Bull

Table 3. Potassium-argon (K-Ar) ages of plutonic rocks in Warm Spring Canyon area.

[Data from E.H. McKee (written commun., 1973, 2005). K_2O analysis by lithium metaborate flux fusion-flame photometric technique that has lithium acting as internal standard. Argon determinations made in USGS laboratory in Menlo Park, Calif., using standard isotope-dilution techniques in 60°-sector, 15.2-cm-radius, Neir-type mass spectrometer. Ages recalculated using constants of Dalrymple (1979): $^{40}K/K = 1.167 \times 10^{-4}$; $\lambda_\varepsilon + \lambda_\varepsilon' = 0.581 \times 10^{-10}$/yr; $\lambda_\beta = 4.962 \times 10^{-10}$/yr]

Locality No.	Sample No.	Map Unit	Mineral Dated	K_2O%	$^{40}Ar_{rad}$%	$^{40}Ar_{rad}$ mole g^{-1}	Age (Ma)
1	W703	Monzogranite (KJm)	Biotite Hornblende	9.12 1.383	96.2 86.5	2.068×10^{-9} 2.981×10^{-10}	151.2 ± 5 144.1 ± 5
2	W713	Monzogranite (KJm)	Hornblende	1.115 1.153	62.2	2.443×10^{-10}	144.1 ± 5
3	W700	Tonalite (KJt)	Biotite	8.56	86.2	1.9955×10^{-9}	154.7 ± 5
4	W707	Tonalite (KJt)	Hornblende	1.073 1.074	66.9	2.3685×10^{-10}	147.1 ± 5

(1991), the ages of which are considered to be early Pleistocene (and older) and middle to late Pleistocene, respectively. The oldest unit (QTa$_1$) could be equivalent to some deposits included in the older alluvium unit (QToa). Unit Qa$_4$ is equivalent to Holocene deposits that form geomorphic surfaces Q3 and Q4 of Bull (1991). Dissected but unvarnished fan deposits near the western and southern margins of the area are mapped as a separate unit (Qa$_3$) that we provisionally interpret to be intermediate in age between units Qa$_2$ and Qa$_4$.

Recent studies of alluvial-fan deposits in Death Valley, summarized by Knott and others (2005), suggest the following subunits and numerical age constraints: Q1 (our unit QTa$_1$), 3.35 Ma to 300 ka; Q2 (our unit Qa$_2$), 318 ka to 30 ka; Q3 (older part of our unit Qa$_4$), younger than 125 ka; and Q4 (younger part of our Qa$_4$), younger than 1.2 ka. Knott and others (2005) stressed that numerical age control on the alluvial-fan deposits is sparse and that additional research clearly is needed.

Structural Geology

Rocks in the Warm Spring Canyon area are deformed by a complex array of structural features that range in age from Paleoproterozoic to Cenozoic. The dominant structural feature in the map area is a major discontinuity, intruded and largely obscured by Late Jurassic to Early Cretaceous tonalite (unit KJt), that strikes east-southeast across the area and juxtaposes Proterozoic rocks on the north against Devonian to Jurassic rocks on the south. This discontinuity, the Butte Valley Thrust Fault of Wrucke and others (1995), has a stratigraphic throw of more than 10 km and an anomalous east-west strike in a region characterized by northerly structural trends. Another prominent structure in the area is a zone of low-angle faults at or near the base of the Noonday Dolomite (Ᵽ$_N$n), probably part of the late Cenozoic Tucki Mountain detachment system (Hodges and others, 1990).

Paleoproterozoic Structural Features

Deformation of Paleoproterozoic age produced gneissic fabric in the monzogranite gneiss (Ᵽ$_P$mg), as well as schistosity, axial-plane cleavage, mineral alignments, and folds in the mica schist (Ᵽ$_P$ms) (Wrucke, 1966). The steeply dipping schistosity in the mica schist and the gneissic structure in the monzogranite gneiss developed concomitantly and produced a strong northwest-trending fabric.

Two sets of folds are evident in the mica schist. The older set, abundant throughout the area, has amplitudes equal to about half the distance between crests, commonly a few centimeters to more than 10 m. These folds generally trend N. 20° W. to N. 40° W. and plunge 15° to 50°. Many folds display axial-plane cleavage. Crinkles in the schistosity and lineations from mineral streaks run parallel to the axes of the larger folds. The younger set of folds is found mostly within about 1 km of the monzogranite gneiss. These folds are more open than those of the older set and have wavelengths that commonly are 5 to 30 cm, three to six times their amplitude. They commonly plunge 5° to 15°, generally to the northwest, though some plunge gently to the southeast. Mineral lineations of the older fold set wrap across folds of the younger set at moderate angles. Possibly the younger folds developed soon after the older ones, on the basis of the small divergence in trends and of the location of the younger set, mainly near the monzogranite gneiss. The gneiss may have acted as a bulwark near which stresses were reoriented slightly during the waning phase of the deformation.

Mesoproterozoic Structural Features

Diabase (unit Ᵽ$_M$d) intruded the stratigraphic section in the Warm Spring Canyon area late in the Mesoproterozoic, producing numerous faults but not significantly deforming the rocks internally. Diabase invaded the section as tabular masses along bedding planes in the dolomitic members of the Crystal Spring Formation, lifting overlying strata. As a result, diabase bodies added at least 300 m to the thickness of the Mesoproterozoic section.

Faults developed locally in the host rocks where diabase broke the section and thickened it abruptly, lifting parts of the section higher than adjacent parts. Vertical contacts between diabase and the broken section at the ends of sills, however, are not faults, as proved by chilled diabase against the vertical contacts. Many faults near the north border of the map area are in strata that are surrounded by diabase into which the faults do not extend.

Mesozoic Structural Features

Butte Valley Thrust Fault

The fault that was later called the Butte Valley Thrust Fault was first recognized by Johnson (1957), who considered it to have normal movement; it was extended into the Warm Spring Canyon area as a north-dipping thrust fault of Mesozoic age by Wrucke (1966). The fault also was shown as a thrust fault by Poole and others (1967), but its detailed thrust-fault geometry was not described until later by Stevens and others (1974). Subsequent studies have enhanced the thrust-fault interpretation (Wrucke and others, 1995, 1997; Wrucke and Stevens, 1998), although, alternatively, the feature also has been interpreted as a segment of a circular, caldera-bounding structure (Davis and Burchfiel, 1997). Several tectonic reconstructions, however, have related the Butte Valley Thrust Fault to regional thrust-fault systems (Wernicke and others, 1988; Serpa and Pavlis, 1996; Snow and Wernicke, 2000). In this report, we interpret the Butte Valley Thrust Fault as a major thrust fault that carried Proterozoic rocks present in the southern part of the Panamint Range over Devonian to Jurassic rocks.

The overall inferred trace of the Butte Valley Thrust Fault is arcuate (Wrucke and others, 1995). West of the map area, it trends northeastward and then eastward beneath alluvium through Butte Valley (fig. 2). In the map area, the inferred trace of the Butte Valley Thrust Fault follows the elongate tonalite pluton (unit KJt) southeastward to Warm Spring before passing under Tertiary volcanic rocks. Near the pluton, Proterozoic strata on the north steepen anticlinally to the south from subhorizontal to vertical and locally are overturned (see cross section *B–B*'). Devonian to Jurassic strata south of the pluton dip away from the fault, steeply in places close to the pluton and more gently farther south. In addition, the Bird Spring Formation (Pℙbs) is deformed locally by minor folding. These structural relations are consistent with the interpretation that the Proterozoic rocks are separated from the Devonian to Jurassic rocks by a pre-intrusive, north-dipping thrust fault. Stretched pebbles in the Early Mississippian Stone Canyon Limestone just below the inferred trace of the thrust fault plunge 55° to the north; these pebbles may parallel the slip direction of thrusting.

Map-scale inclusions of gneiss, diamictite of the Kingston Peak Formation (Ꞓ$_{N}$ksd), fine-grained biotite-bearing quartzite (unit fq), and orthoquartzite (unit oq) in the tonalite pluton are inferred to lie between strands of the Butte Valley Thrust Fault. We interpret the gneiss to be a fragment of the Paleoproterozoic monzogranite gneiss (Ꞓ$_{P}$mg) from the upper plate of the thrust fault and the quartzite inclusions to be Paleoproterozoic and lower Paleozoic units, possibly the Wood Canyon Formation, the Zabriskie Quartzite, the Eureka Quartzite, or quartzite of the Lost Burro Formation (Dlb) derived from the lower plate. The westernmost inclusion of orthoquartzite appears to be faulted against the Early Mississippian Tin Mountain Limestone (Mtm), although the contact is not well exposed.

The age of the Butte Valley Thrust Fault is bracketed between deposition of the Middle to Late Jurassic(?) Warm Spring Formation (units Jws, Jwt, and Jws) and emplacement of the Late Jurassic to Early Cretaceous tonalite (unit KJt) along the fault zone (Wrucke and others, 1995). This age of thrusting overlaps with that of the East Sierran thrust-fault system to the west (Dunne, 1986; Dunne and Walker, 2004). The Layton Well Thrust Fault in the Slate Range, 35 km southwest of Warm Spring Canyon (fig. 1), is considered to be part of the East Sierran system and is thought to lie in the upper plate of the Butte Valley Thrust Fault (Wrucke and others, 1995).

The Butte Valley Thrust Fault may be related to other thrust faults to the east. One of these is the Chicago Pass Thrust Fault in the Nopah Range, 80 km east of Warm Spring (Serpa and Pavlis, 1996) (fig. 1). It resembles the Butte Valley Thrust Fault in having Paleozoic rocks in the lower plate and Neoproterozoic rocks in the upper plate, but its age is bracketed only between Middle Pennsylvanian and Tertiary (Burchfiel and others, 1983). Other thrust faults that resemble the Butte Valley Thrust Fault are the Winters Pass Thrust Fault in the Clark Mountain Range, 120 km southeast of Warm Spring, and the Wheeler Pass Thrust Fault in the northern Spring Mountains, Nevada, to the north of the Winters Pass Thrust Fault (fig. 1). The Winters Pass Thrust Fault, like the Butte Valley Thrust Fault, carries Proterozoic rocks over Paleozoic strata and is considered to be Early Cretaceous in age (about 143 Ma) (Walker and others, 1995). The Wheeler Pass Thrust Fault ramps Proterozoic rocks above Pennsylvanian and Permian strata. Its age is poorly constrained but may be Middle Cretaceous (Snow and Wernicke, 2000).

The Butte Valley Thrust Fault has been correlated with the Winters Pass Thrust Fault by Snow and Wernicke (2000) and with the Wheeler Pass Thrust Fault by Wernicke and others (1988). In these reconstructions, a Winters Pass–Wheeler Pass upper plate underlies all intervening mountain blocks west to the Panamint Range, apparently as a regionally extensive ramp-flat allochthon. According to our interpretations, however, the Butte Valley Thrust Fault could not underlie a "thin-skin" allochthon in the Panamint Range, as required in these models, because it dips steeply to the north and cuts deep into the Proterozoic crystalline basement. Thus, the interpretation of a fault cutting deeply into the basement casts doubt on the concept of a thin allochthon extending from the Clark and Spring Mountains as far west as the Panamint Range and suggests that the Butte Valley Thrust Fault is not the same as either the Winters Pass Thrust Fault or the Wheeler Pass Thrust Fault.

Other Mesozoic Faults

Several additional faults of Mesozoic age cut Paleozoic or Mesozoic rocks in the area south of the inferred trace of the Butte Valley Thrust Fault. Probably all of these faults are genetically related to the Butte Valley Thrust Fault.

A minor thrust fault near the eastern margin of Butte Valley separates the Mississippian Santa Rosa Hills Limestone and the underlying strata from the Pennsylvanian and Permian Bird Spring Formation. This fault dips 30° N. and apparently is truncated by the tonalite pluton (unit KJt) that is inferred to intrude the Butte Valley Thrust Fault, of which it probably is a splay.

A major east-southeast-striking fault of probable Mesozoic age transects the Bird Spring Formation and Owens Valley Group throughout the western part of the map area. In the hills just east of Butte Valley, the fault juxtaposes Pennsylvanian rocks of the Bird Spring Formation against the upper unit of the Owens Valley Group and dips 80° N., indicating reverse displacement. Traced eastward, the fault cuts obliquely across the lower unit of the Owens Valley Group and the upper part of the Bird Spring Formation before being truncated by tonalite (unit KJt), just east of fossil locality 5 (see fig. 3 for location). The reverse displacement and the truncation by tonalite suggest that this fault is genetically related to the Butte Valley Thrust Fault. The same fault apparently emerges from the other side of the tonalite and continues through the Bird Spring Formation eastward past fossil locality 27 (see fig. 3 for location), where fusulinid-bearing limestones of probable Permian age are juxtaposed against rocks of probable Pennsylvanian age on the north.

Another east-southeast-striking reverse fault forms the boundary between the lower and upper units of the Owens Valley Group. This fault terminates against the major fault described above 1 km west of fossil locality 4 (see fig. 3 for location). Just east of this termination, a thin structural sliver of Bird Spring Formation that contains the Early Permian colonial coral *Protowentzelella* is present along the fault. The presence of this sliver implies a complex fault history. Further complications are seen 1.5 km farther east, where an anomalous sliver of the upper unit of the Owens Valley Group is present in the fault zone, as well as just east of that locality, where the lower unit of the Owens Valley Group has been duplicated by faulting.

A fault between the lower and upper units of the Owens Valley Group, probably a western extension of the fault described in the previous paragraph, also is exposed in the isolated hills south of the Warm Spring Canyon road, near the west boundary of the map area. Here the fault zone dips steeply northward and includes a prominent sliver of fossiliferous limestone, assigned to the Bird Spring Formation, that contains abundant Early Permian fusulinids and solitary corals (fossil localities 23 and 24), again implying a complex history of fault movement.

The Bird Spring Formation and the Owens Valley Group also are cut by at least seven exposed or inferred faults that strike northeast, offsetting parts or all of the stratigraphic sequence, mostly with left-lateral separation. Five of these faults are near the west boundary of the map area; another is opposite Tram Canyon, and one more is near the east end of the Pennsylvanian-Permian outcrop belt. These faults cut the east-southeast-striking reverse fault that separates the lower and upper units of the Owens Valley Group, but they do not appear to cut the major reverse fault that juxtaposes the lower part of the Bird Spring Formation against younger rocks on the south. They also do not cut either the basal contact of the Warm Spring Formation or the monzogranite pluton that intrudes that contact. The fault opposite Tram Canyon is intruded by an elongate protrusion of the monzogranite.

The observed relations indicate a sequential fault history in which the fault between the lower and upper units of the Owens Valley Group is the oldest, followed by the crosscutting, northeast-striking fault set, and finally by the east-southeast-striking reverse fault that has the Pennsylvanian part of the Bird Spring Formation in the hanging wall. This entire fault sequence probably was part of a major contractional deformation that culminated in the Butte Valley Thrust Fault. The northeast-striking faults may have originated as tear faults in the reverse-fault system.

The basal contact of the Jurassic Warm Spring Formation probably is a fault. This contact clearly truncates the northeast-striking fault set described above, indicating that it is either a postdeformational unconformity or a fault. Because deposition of the Warm Spring Formation almost certainly predated deformation related to the Butte Valley Thrust Fault (Wrucke and others, 1995), which evidently included movement on the northeast-striking faults, our interpretation is that the basal contact of this unit is much more likely to be a fault than an unconformity.

Cenozoic Structural Features

Numerous faults of known and inferred late Cenozoic age cut rocks of the Warm Spring Canyon area. Possibly the oldest, here named the Queen of Sheba Fault, is at the base of the Noonday Dolomite. This fault bears a strong resemblance to faults of the late Cenozoic Tucki Mountain detachment system farther north in the Panamint Range (Hodges and others, 1990) and probably is a southern representation of that system. Hodges and others (1990, fig. 3) showed the Harrisburg Fault at the base of that system as extending south from Harrisburg Flats (fig. 1) to the Warm Spring Canyon area. The Tucki Mountain detachment system is interpreted to have originated as a complex system of west-dipping low-angle normal faults, movements on which led to unroofing of the metamorphic core of the Panamint Range during a period of large-scale crustal extension (Hodges and others, 1990). Similar detachment masses involving the Kingston Peak Formation, the Noonday Dolomite, and the Johnnie Formation dip westward on the west side of the Panamint Range (Labotka and Albee, 1990). Subsequent folding of the Panamint Range into a north-northwest-trending anticline rotated faults of the Tucki Mountain system such that the Harrisburg Fault now dips generally eastward.

Except in the northwestern part of the map area, the Queen of Sheba Fault separates the Noonday Dolomite from the underlying Kingston Peak Formation, Crystal Spring Formation, and diabase. The fault surface dips eastward at a low to moderate angle. In the northwestern exposures, a zone of low-angle faults that has the Queen of Sheba Fault at the base involves the Noonday Dolomite, the South Park Member of the Kingston Peak Formation, and the Johnnie Formation. The highest fault in this zone separates the Johnnie Formation from the underlying Noonday Dolomite. Locally within the fault zone, the Noonday Dolomite and the South Park Member are tectonically interleaved, and a wedge of the South Park Member is ramped to the east from the Queen of Sheba Fault into the Noonday Dolomite.

In contrast to the Harrisburg Fault, which moved to the north-northwest relative to the lower plate (Hodges and others, 1990), the eastward ramping of the South Park Member over part of the Noonday Dolomite on the Queen of Sheba Fault, in the northwestern part of the map area, suggests that movement of the upper plate of the fault at that location was to the east.

Most of the Noonday Dolomite in the Warm Spring Canyon area is highly disturbed. For example, the dolomite centered along the northeastern border of the map area, 2.5 km west of the Queen of Sheba Mine, consists of disoriented breccia blocks as large as room size that dip in a variety of directions. Beds in the lower few meters of this dolomite and other beds low in the formation elsewhere in the map area are not highly deformed but locally are buckled and brecciated. In many parts of the area, the Noonday Dolomite is fractured into blocks that appear to be slightly disoriented one to another and locally are separated by thin zones of dolomite breccia. In the southeasternmost outcrops of the Noonday Dolomite, well-indurated, horizontally laminated, dolomitic calcarenite between breccia fragments apparently was deposited by water flowing through the fragmented rock.

Miocene volcanic rocks (units Tab, Tst) depositionally overlap the Queen of Sheba Fault near the eastern margin of the map area, restricting the minimum age of the fault to about 13.6 Ma. This age constraint further suggests a possible correlation between the Queen of Sheba Fault and the Harrisburg Fault, which predates the 10.6-Ma Little Chief stock (Hodges and others, 1990). The Harrisburg Fault is similar to the Queen of Sheba Fault at the base of the Noonday Dolomite in the Warm Spring Canyon area, except that the Harrisburg Fault is located mostly at the base of the Johnnie Formation. The Queen of Sheba Fault possibly is a sub-Noonday splay of the

Harrisburg Fault, similar to that illustrated by Hodges and others (1990, fig. 4) at Harrisburg Flats.

Numerous high-angle faults in a variety of orientations, which are present mainly along the topographically low eastern part of the map area, are a southward continuation of a complex of faults mapped by Hunt and Mabey (1966). Several normal faults cut Miocene volcanic rocks and, thus, clearly are younger than the Queen of Sheba Fault. Faulted alluvial fans on the west side of Death Valley, north of the map area (Hunt and Mabey, 1966), show that crustal adjustments have continued on the east side of the Panamint Range into the Quaternary.

DESCRIPTION OF MAP UNITS

Alluvial-fan and wash deposits (Holocene to Miocene?)—Unconsolidated to weakly consolidated gravel deposits. Not studied in detail; mapped primarily from inspection of aerial photographs and by comparison with units mapped in areas to north by Hunt and Mabey (1966). Subunits distinguished on basis of geomorphic and surficial characteristics, following principles of Bull (1991). Divided into following subunits:

Qa_4 **Unit 4 (Holocene)**—Deposits of modern washes and very young deposits of alluvial fans. Unconsolidated and unsorted gravel, sand, and silt. Present in channels of modern streams in Warm Spring Canyon and its tributaries, as well as in narrow, incised ephemeral streams and adjacent alluvial surfaces in northeastern part of map area. Clasts are unvarnished; surfaces commonly are very rough and appear light on aerial photographs. Equivalent to no. 4 and no. 3 gravels of Hunt and Mabey (1966) and alluvium-forming geomorphic surfaces Q4b and Q4a (latest Holocene) and Q3c to Q3a (late to early Holocene) of Bull (1991)

Qa_3 **Unit 3 (Holocene)**—Young deposits of alluvial fans. Present near western and southern margins of map area. Unconsolidated and unsorted gravel, sand, and silt forming rough to locally smooth surfaces that have abundant interfluves. Gravels are unvarnished; surfaces appear light on aerial photographs. Incised by channels of ephemeral streams

Qa_2 **Unit 2 (Pleistocene)**—Old deposits of alluvial fans. Weakly consolidated, poorly sorted gravel, sand, and silt forming smooth, planar surfaces that have desert pavements. Gravels are darkly varnished; surfaces appear dark on aerial photographs. Abundant interfluves commonly are 5 to 10 m deep. Partly equivalent to no. 2 gravel of Hunt and Mabey (1966) and alluvium-forming geomorphic surfaces Q2c to Q2a (late to middle Pleistocene) of Bull (1991)

QTa_1 **Unit 1 (Pleistocene to Miocene?)**—Very old deposits of alluvial fans. Weakly consolidated, poorly sorted gravel, sand, and silt forming elongate, rounded ridges between closely spaced interfluves that commonly are as deep as 15 to 20 m and, locally, 25 m. Some ridges preserve narrow, elongate, planar areas of desert pavement composed of darkly varnished gravel; other ridges are sharper, more deeply eroded, and do not preserve such pavements. Surfaces lie above those of unit Qa_2. Unit appears dark on aerial photographs. Partly equivalent to no. 2 gravel of Hunt and Mabey (1966) and also alluvium-forming geomorphic surface Q1 (early Pleistocene and older) of Bull (1991)

Qt **Talus (Holocene)**—Rubbly debris derived mainly from overlying rock units. Found in three areas near west border of map area

Landslide deposits (Holocene)—Blocky rubble and brecciated masses of dislocated rock, present in eastern part of map area. Divided according to bedrock source unit into following subunits:

Qlab **Deposits derived from the Miocene andesite and basalt [unit Tab]**—Dominant component of several large landslide masses and one small landslide in area of Miocene volcanic rocks on north side of Warm Spring Canyon

Qln **Deposits derived from the Neoproterozoic Noonday Dolomite [unit $\text{Ꝑ}_N\text{n}$]**—Present in large, monolithic landslide mass at Queen of Sheba Mine (Morton, 1965) and also in part of smaller landslide mass on south side of Warm Spring Canyon, south of Grantham Mine

Qld **Deposits derived from the Mesoproterozoic diabase [unit $\text{Ꝑ}_M\text{d}$]**—Minor component of large landslide mass on north side of Warm Spring Canyon; primary component of smaller landslide mass on south side of Warm Spring Canyon, south of Grantham Mine

Qls **Deposits derived from the stromatolite member of the Mesoproterozoic Crystal Spring Formation [unit $\text{Ꝑ}_M\text{cst}$]**—Minor component of large landslide mass on north side of Warm Spring Canyon

Qlc **Deposits derived from the cherty dolomite member of the Mesoproterozoic Crystal Spring Formation [unit Ꝑ$_{M}$cc]**—Very minor component of large landslide mass on north side of Warm Spring Canyon, present adjacent to wash

QToa **Older alluvium (Pleistocene? to Miocene?)**—Unconsolidated to weakly consolidated gravel deposits. Found mainly along lower slopes of Warm Spring Canyon and locally on ridge crests. Deposits along Warm Spring Canyon slope toward canyon bottom; surfaces are variably dissected. Also found on crest of ridge south of Tram Canyon and at head of Galena Canyon in north-central part of map area, 130 m and 240 m above Warm Spring Canyon, respectively. Possibly correlative with similar high-level alluvial deposits of Miocene and Pliocene age 70 km to north in Cottonwood Mountains (Snow and Lux, 1999); may also include deposits of Pleistocene age

Volcanic rocks (middle Miocene)—Divided into following units:

Tab **Andesite and basalt**—Gray to brown flows of olivine-pyroxene and oxyhornblende andesite and lesser amounts of red vesicular andesite, black olivine basalt, and volcanic rocks of other compositions; locally interspersed with prominent light-colored silicic tuff (unit Tst). Flows south of Warm Spring commonly are 6 to 15 m thick, in sections as thick as 425 m; flows in southeastern part of map area filled deep northeast-trending paleovalley that records deeply dissected topography of Miocene age. Latite flow in eastern part of map area north of Warm Spring Canyon yielded ^{40}Ar/^{39}Ar age of 13.56 ± 0.82 Ma (Topping, 1993)

Tst **Silicic tuff**—Tan to white silicic tuff in distinctive layers within andesite-basalt sequence (unit Tab), in southeastern part of map area. Consists of poorly stratified, unwelded lapilli-rich tuffs, in layers commonly as thick as 30 m but locally as thick as 120 m. Tuff above dated latite flow in unit Tab gave zircon fission-track age of 10.3 ± 0.5 Ma (Topping, 1993)

KJm **Monzogranite (Early Cretaceous or Late Jurassic)**—Generally medium-grained plutonic rock composed of about 30 percent quartz, 35 percent plagioclase, 27 percent orthoclase, 6 percent biotite, and minor amounts of hornblende. Locally contains orthoclase phenocrysts as long as 2.5 cm. Fine-grained phase abundant along border adjacent to and within rocks of the Warm Spring Formation. K-Ar ages include one from biotite of 151.2 ± 5 Ma and two from hornblende, both 144.1 ± 5 Ma (table 3, localities 1 and 2)

KJt **Tonalite (Early Cretaceous or Late Jurassic)**—Medium-gray, medium-grained hypidiomorphic-granular tonalite; emplaced as narrow pluton for at least 7 km along inferred former trace of Butte Valley Thrust Fault, from west border to south-central part of map area near Warm Spring, where the pluton passes beneath cover of Tertiary volcanic rocks. Consists of, on average, about 50 to 55 percent plagioclase, 25 percent quartz, 3 percent orthoclase, 10 percent biotite, and accessory hornblende, sphene, allanite, epidote, prehnite, magnetite, and zircon. Plagioclase strongly zoned from labradorite in cores to oligoclase at rims. Biotite and hornblende K-Ar ages of 154.7 ± 5 Ma and 147.1 ± 5 Ma, respectively (table 3, localities 3 and 4)

Warm Spring Formation (Late to Middle Jurassic?)—Consists of volcanic rocks and minor amounts of sandstone; adjacent to type locality of the Warm Spring Formation of Johnson (1957) to west. Not dated, but probably approximately coeval with similar volcanic sequences to northwest, in Inyo Mountains and Alabama Hills, and southwest, in Slate Range (Dunne and Walker, 1993), as well as to south, in Soda Mountains (Grose, 1959) (fig. 1). U-Pb zircon ages of these sequences range from 170 to 148 Ma (Dunne and others, 1994). In map area, rests in fault contact on upper unit of the Owens Valley Group (Povu), but few kilometers to southwest, lies on Triassic sedimentary rocks (Johnson, 1957; Cole, 1986). Divided into following subunits:

Jwa **Andesite**—Dark-gray, almost black, fine-grained meta-andesite flow rocks. Forms section 300 m thick in which individual flows are obscure. Represents uppermost of four units in the Warm Spring Formation as distinguished by Cole (1986)

Jwt **Ash-flow tuff**—Light-gray, angular, lens-shaped fragments commonly 1 cm across that resemble partly flattened pumice lapilli, in medium-gray aphanitic matrix. Metamorphosed to fine-grained hornfels. Makes up lens about 30 m thick near base of andesite (unit Jwa)

Jws **Sandstone**—Fine-grained sandstone. Locally contains calcium-silicate minerals in remnants of metamorphosed calcareous arenite; sillimanite and andalusite in former shaly layers; and minor amounts of metatuff. In easternmost exposures, thin metalimestone conglomerate containing metatuff matrix locally is present at base. Eastern outcrops include unmapped fine-grained quartz monzogranite that intruded metasedimentary rocks of this unit. Thickness, as much as 100 m

Owens Valley Group (Early Permian)—Consists primarily of brown-weathering siltstone, sandstone, and argillite, and lower unit of siliceous limestone. Divided into following subunits:

Povu **Upper unit**—Buff- to brown-weathering calcareous siltstone to fine-grained sandstone, brown-weathering argillite, and subordinate amounts of medium-gray, fine-grained limestone and silty limestone. Calcareous siltstone to fine-grained sandstone form even, planar beds, 5 to 25 cm thick, that exhibit low-angle cross-lamination, convolute lamination, and, locally, graded bedding; these beds are here interpreted as turbidites. Interbedded argillite forms layers averaging about 5 cm thick. Limestone beds, also here interpreted as turbidites, are planar and 20 to 50 cm thick; they are present both individually and in zones as thick as 25 m. Bases of selected limestone zones are indicated on map by blue line symbol. Ammonoids are present at three localities (table 2, localities 15–17; see also, fig. 9); fusulinids and conodonts are rare (table 2, localities 20, 21, and 25). These fossils suggest late Early Permian age. Total exposed thickness, about 565 m. Base and top of unit are faulted

Povl **Lower unit**—Medium- to dark-gray, medium-bedded, fine-grained limestone containing irregular, discontinuous beds and lenses of buff- to brown-weathering siltstone, argillite, and possibly chert, 3 to 7 cm thick, that impart distinctive striped appearance. Buff- to brown-weathering layers are dark gray on fresh surface; some are finely laminated. Conformably overlies the Bird Spring Formation (PIPbs); base of unit is drawn at base of prominent bed of brown calcareous siltstone about 5 m thick. Average thickness of unit, about 90 m. Top of unit is faulted

PIPbs **Bird Spring Formation (Early Permian and Pennsylvanian)**—Light- to dark-gray, thin-bedded to massive limestone, silty to argillaceous limestone, cherty limestone, and rare siltstone. Upper part, which is south of major east-west-striking fault that transects formation at low angle to bedding, is composed of light- to medium-gray, thick-bedded to massive limestone, silty to argillaceous limestone, minor amounts of cherty limestone, and rare siltstone; limestone commonly is recrystallized to marble. This part of formation locally contains abundant fusulinids and colonial corals (table 2) that indicate Late Pennsylvanian to Early Permian age. Lower part of formation, north of major fault, primarily consists of medium- to dark-gray, thin- to thick-bedded micritic limestone containing abundant nodules, lenses, and thin beds of dark-brown-weathering chert. Many chert nodules are spherical to ovoid. Some limestone contains thin beds and lenses of brown-weathering argillite or siltstone. Toward top of lower part, thick-bedded to massive limestone, like that south of fault, is present. Fossils are rare in lower part of formation. Lowermost 3 m of formation consists of medium- to coarse-grained limestone containing abundant crinoid and shell fragments. Conodonts from this basal zone (table 2, locality 11) indicate Early Pennsylvanian age (Stone, 1984). Maximum exposed thickness of formation, about 525 m

Mis **Indian Springs Formation (Late Mississippian)**—Brown-weathering calcareous siltstone and dark-gray shale; minor amounts of dark-gray crinoidal limestone. Basal contact sharp. Uniformly about 6 m thick

Msr **Santa Rosa Hills Limestone (Late and Early Mississippian)**—Mostly light-gray, coarse-grained, massive limestone; upper 10 m, medium to dark gray. Beds generally are not evident but locally are 0.3 to 3 m thick. Chert nodules 10 to 20 cm thick and as much as 60 cm long are common in lower 10 to 15 m of formation. Locally contains abundant crinoid columnals and sparse large syringoporid corals (table 2, locality 10). Contact with the underlying Stone Canyon Limestone (Msc) is gradational. Thickness, about 100 m

Msc **Stone Canyon Limestone (Early Mississippian)**—Dark-gray limestone, in beds commonly 5 to 20 cm thick; interbedded with variable amounts of black chert, in beds 1 to 5 cm thick. Matrix-supported conglomerate bed, interpreted as submarine debris-flow deposit, locally present near top. Grades into the underlying Tin Mountain Limestone (Mtm) over interval of several meters. Thickness, about 60 m

Mtm **Tin Mountain Limestone (Early Mississippian)**—Mostly dark-gray cherty limestone, in beds 10 to 50 cm thick. Upper and lower parts contain about 25 percent chert, in thick black nodules that weather brown and black; middle part is chert free. Many limestone beds are rich in fossil material that includes crinoid columnals, syringoporid corals, horn corals, and sparse brachiopods (table 3, locality 1). Thickness, about 50 m

Dlb **Lost Burro Formation (Late and Middle Devonian)**—Pale-yellow to medium-gray dolomite, at base of Paleozoic sequence near west border of area. Bedding not evident. Weathered surface very rough. Base intruded by tonalite (unit KJt). Exposed thickness, 50 to 60 m

oq **Orthoquartzite (Paleozoic?)**—Light-gray, fine- to medium-grained quartzite forming large inclusions in tonalite (unit KJt). Exposed in western part of map area between inferred strands of Butte Valley Thrust Fault. Inclusion in exposed thrust-fault contact with the Tin Mountain Limestone (Mtm) has vague crossbedding. Stratigraphic position is unknown, but unit resembles the Eureka Quartzite of Ordovician age; also may be the Zabriskie Quartzite of Cambrian age or quartzite in the Lost Burro Formation of Devonian age

fq **Fine-grained quartzite (Paleozoic?)**—Light-gray and very dark gray, fine-grained, layered quartzite forming inclusions in tonalite (unit KJt) between inferred strands of Butte Valley Thrust Fault near west border of area, southeast of section *B–B*'. Light-colored layers are composed of quartz; dark layers are composed of white mica, biotite, quartz, idocrase, and scapolite(?). Layers are 1 to 10 mm thick, variable in thickness, and distorted. Stratigraphic identity is uncertain but possibly is the Wood Canyon Formation of Cambrian and Neoproterozoic age

Ꝑ$_{N}$j **Johnnie Formation (Neoproterozoic)**—Shale, siltstone, and quartzite in upper part (about 30 m thick); interbedded quartzite, sandy dolomite, and dolomitic quartzite in lower part (about 100 m thick). Exposed in northwest corner of map area. Beds in upper part are yellowish brown to medium brown, laminated to thin bedded; uppermost 10 m consists of quartzite that contains abundant well-rounded, black granules. Beds in lower part commonly are 10 cm to 2.4 m thick and conspicuously crossbedded. Quartzite is light to medium gray; dolomitic rocks are brown to yellowish brown and contain coarse quartz grains that are patinated dark brown. Stromatolitic dolomite is present 55 m above base. Basal contact with the Noonday Dolomite (Ꝑ$_{N}$n) is placed at base of lowest quartzite bed. Exposures in map area make up only lower 130 m of formation, probably equivalent to transitional member of Stewart (1970). Formation is 700 m thick farther north in Panamint Range (Stewart, 1970)

Ꝑ$_{N}$n **Noonday Dolomite (Neoproterozoic)**—Light-colored, cliff-forming dolomite; widespread across northern part of map area and on ridge crest at southeast corner. Upper part of formation, about 60 m thick, consists of light-gray, generally massive dolomite that locally has discontinuous beds as thick as 4 m. Uppermost 3 m in northwest corner of map area consists of light-brown sandy dolomite resembling lower parts of the overlying Johnnie Formation. Lower part, about 100 m thick, consists of pale-yellow to light-gray, fine-grained, laminated to very thin bedded, microbial-mat dolomite. This dolomite locally contains abundant conspicuous, cylindrical, tube-like structures, 20 cm to more than l m long, oriented normal to bedding. Rests on low-angle Queen of Sheba Fault of Tertiary age throughout map area; rocks at or near base commonly are shattered and brecciated and locally are broken into rotated blocks as large as rooms. Basal dolomite breccia blocks present on ridge south of Grantham Mine near southeast corner of map area are cemented by dolomitic calcarenite laminated parallel to basal contact. Age, probably 590 Ma or younger (see discussion of unit in section entitled, “Mesoproterozoic and Neoproterozoic Rocks”)

Ꝑ$_{N}$cs **Conglomerate and sandstone (Neoproterozoic)**—Exposed at three small localities in map area. First locality, consisting of both conglomerate and sandstone, is 2.8 km west-southwest of the Queen of Sheba Mine: pebbles and cobbles of sandstone, subordinate amounts of carbonate rock, and single boulder of diabase, all supported in more abundant matrix of brown-weathering arkosic sandstone and siltstone; iron oxide abundant in upper 1 m; thickness, 6 m. Other two localities, consisting of only sandstone, are 1.2 km west-southwest and 1.5 km south-southeast of mine: yellowish-brown-weathering, highly indurated sandstone; locality south-southeast of mine also has bed that may be altered tuff; thickness, maximum of 3.5 m

Pahrump Group (Neoproterozoic and Mesoproterozoic)

Kingston Peak Formation (Neoproterozoic)—Divided into following members:

Ꝑ$_{N}$ksp **South Park Member**—Black to medium-gray, very thin bedded limestone, argillite, and silty limestone resting unconformably on diamictite (unit Ꝑ$_{N}$ksd), in northwestern part of map area. Well exposed in Tram Canyon. Beds typically 1 to 5 cm thick; ratio of limestone to argillite and calcareous siltstone, about 60:40. Locally contains sections 2 to 25 m thick composed primarily of laminated, black to purple-black argillite. Contains few conspicuous brecciated beds of orange limestone, 0.5 to 2 m thick; rare beds of light-gray, flat-pebble limestone conglomerate, 0.5 to 1 m thick; and, in lower 20 m, yellowish-gray to pale-brown, laminated dolomite. Bedding locally distorted by soft-sediment deformation. Age, probably 750 Ma or younger (Prave, 1999). Thickness, at least 100 m

Surprise Member—Diamictite and subordinate amounts of siltstone; exposed in Tram Canyon and, as thin remnants beneath the Noonday Dolomite (Ꝑ$_N$nd), east of Galena Canyon. Divided into following subunits:

Ꝑ$_N$ksd **Diamictite**—Matrix-supported conglomerate composed of randomly distributed pebbles, cobbles, and boulders, in considerably more abundant fine-grained, greenish-black matrix. Clasts larger than 1 cm make up about 5 percent of rock. Clasts of granitic gneiss, granite, and quartzite are dominant, but limestone and diabase clasts also are present. Most clasts greater than sand size are subangular to rounded pebbles; cobbles and boulders are rare. One prominent limestone boulder 5 by 5 by 9 m is exposed high in formation in Tram Canyon. Matrix, originally silty mudstone, is hornfels composed of green biotite, subordinate amounts of actinolite, sparse calcite, and 10 percent detrital silt and sand. Most exposures are massive. Rests unconformably on diabase (unit Ꝑ$_M$d) and locally on the cherty dolomite and stromatolite members of the Crystal Spring Formation (units Ꝑ$_M$cc and Ꝑ$_M$cst, respectively). Complete section not present. Exposed thickness, about 90 m

Ꝑ$_N$kss **Siltstone**—Dark-purplish-gray and less abundant greenish-gray, extremely well indurated siltstone. Present as lens in diamictite (unit Ꝑ$_N$ksd) in Tram Canyon, and as separate body on diabase (unit Ꝑ$_N$d) adjacent to diamictite on east side of Galena Canyon, near north-central border of area. Bedding mostly not evident, but, locally in Tram Canyon, wispy, irregular lenses 2 cm thick and 10 to 20 cm long are present, reflecting slight variations in composition. Contains few conspicuous lenses of light-gray to blue-gray limestone and yellow dolomite, 0.5 to 3 m thick. Thickness, as much as about 30 m

Crystal Spring Formation (Mesoproterozoic)—Divided into following members:

Ꝑ$_M$cch **Chert member**—Purplish-black and less abundant brownish-red chert; exposed high on ridge south of Grantham Mine and on north side of Warm Spring Canyon east of mine, as well as at scattered localities in north-central parts of map area. Mostly massive to vaguely bedded but locally finely laminated. Contains abundant very fine grained iron oxide; minor amounts of biotite, chlorite, and white mica; and rare local concentrations of plagioclase. Interpreted to have originated by replacement of carbonate strata previously present in upper part of the stromatolite member (Ꝑ$_M$cst). Fills former irregular solution cavity, several tens of meters across, developed in upper part of stromatolite member northeast of Grantham Mine. Contains few thin lenses of dolomite near base. Top either faulted, intruded by diabase, or unconformably overlain by Tertiary volcanic rocks or the Noonday Dolomite (Ꝑ$_N$n). Maximum exposed thickness, about 90 m

Ꝑ$_M$cst **Stromatolite membe**r—Medium-gray to purplish-gray, light-brown to brownish-red-weathering, cliff-forming dolomite that locally contains abundant stromatolites. Excellent exposures in vicinity of Grantham Mine. Sequences of stromatolitic dolomite alternate with sequences of massive to laminated dolomite that contains small amounts of chert. Where best preserved, stromatolites are laterally linked stacked hemispheres, 5 to 10 cm in diameter, in beds 0.2 to 3 m thick. Top transected by either disconformity, low-angle fault, or diabase sill. Thickness, 110 m

Ꝑ$_M$cc **Cherty dolomite member**—Predominantly pale-grayish-orange dolomite and brown to black chert; in beds and nodules 1 to 3 cm thick and spaced 1 to 8 cm apart, locally cross-bedded. Forms sequences 10 to 15 m thick that alternate with sequences 4 to 6 m thick of brown-weathering dolomite or dolomitic limestone that contains much fine-quartz silt and microscopic chert, in fine meshwork that forms crepe-like appearance on weathered surfaces. Unit contains penecontemporaneous breccia, suggestive of deposition in shallow waters. Rests gradationally on the argillite member (Ꝑ$_M$ca) with which it intergrades over interval of several meters. Locally altered to talc, which was produced from several mines in map area (Wright, 1963; Chidester and others, 1964; Franklin, 1965). Thickness, about 90 to 115 m

Ꝑ$_M$ca **Argillite member**—Reddish-purple, firmly indurated argillite and subordinate amounts of siltite, quartzite, and dolomite; rests on the sandstone member (Ꝑ$_M$cs). Upper one-third of member is composed of thin-bedded, commonly cross-laminated siltite, silty argillite, and dolomite; lower two-thirds of member is composed of argillite and silty argillite characterized by light-gray to greenish-gray reduction spots that average 2 to 3 cm in diameter. Bedding in lower two-thirds of member is indistinct, but closely spaced partings give appearance of thin bedding. Some argillite may be tuffaceous, as vaguely suggested in thin sections. Thickness, 100 m

Ꝑ$_{M}$cs **Sandstone member**—Brownish-gray to pale-pinkish-gray arkose, pale-red to light-gray feldspathic sandstone, and minor amounts of conglomerate. Rests on Paleoproterozoic schist and gneiss (units Ꝑ$_{P}$ms and Ꝑ$_{P}$mg, respectively) in east-central part of map area. Well bedded and commonly cross-laminated. Upper one-fourth of member is composed of fine- to medium-grained, feldspathic sandstone, in beds 5 to 40 cm thick; lower three-fourths of member is composed of medium- to coarse-grained arkose, in beds 15 cm to 3 m thick. Well-rounded pebbles of quartz, quartzite, schist, and gneiss are scattered in member and form thin conglomerate beds and lenses, 2 to 60 cm long, in lower 20 m. Basal conglomerate present locally. Thickness, 140 to 190 m

Ꝑ$_{M}$d **Diabase (Mesoproterozoic)**—Dark-greenish-black, highly altered sills and sheets, 15 to 110 m thick. Present principally in the Crystal Spring Formation in southeastern, north-central, and northeastern parts of map area; also includes dike in Paleoproterozoic mica schist (unit Ꝑ$_{P}$ms) north of Warm Spring. Consists primarily of albite, actinolite, and chlorite derived from ophitic, mainly plagioclase-augite alkalic olivine basaltic rock. In southeast part of map area, comprises six separate sills, including one sill (mapped separately as unit Ꝑ$_{M}$ds) within another south and east of Grantham Mine. Adjacent carbonate-rock (stromatolite and cherty dolomite) members of the Crystal Spring Formation (units Ꝑ$_{M}$cst and Ꝑ$_{M}$cc, respectively) are metamorphosed to calc-silicate–bearing limestone only within few meters of contact. Talc deposits and rare seams of chrysotile asbestos locally are present in the cherty dolomite member of the Crystal Spring Formation (Ꝑ$_{M}$cc) adjacent to sills. Age of probably about 1,070 to 1,090 Ma is based on baddeleyite from diabase in Kingston Range east of Death Valley (Heaman and Grotzinger, 1992) (table 1)

Ꝑ$_{P}$pa **Pegmatite and aplite (Paleoproterozoic)**—Pegmatite and aplite dikes that intrude mica schist (unit Ꝑ$_{P}$ms) in central part of map area. Pegmatite principally consists of quartz, perthitic microcline, and albite, in crystals 2 to 8 cm in diameter. Books of white mica, 1 to 8 cm across, make up 10 percent of rock. Aplite dikes, 4 to 30 cm wide and 5 m or more long, are found only on east side of north-trending canyon 2.8 km northwest of Warm Spring. Wasserburg and others (1959) reported Rb-Sr ages of 1,700 Ma and 1,660 Ma for white mica and 1,570 Ma for perthite, all from pegmatite 2.6 km north-northeast of Warm Spring

Ꝑ$_{P}$mg **Monzogranite gneiss (Paleoproterozoic)**—Reddish-brown monzogranite gneiss intrusive in mica schist (unit Ꝑ$_{P}$ms); overlain by the sandstone member of the Crystal Spring Formation (Ꝑ$_{M}$cs) north of Warm Spring. Main body is 0.5 by 1.8 km in size. Several dikes of gneiss, as much as 3 m wide and 0.5 km long, are exposed north of main body. Composed of closely spaced augen of quartz and microcline, commonly 6 to 30 mm wide, separated by anastamosing foliation planes and matrix of quartz, albite, and white mica. Contact locally is parallel to lithologic layers in adjacent mica schist. Northwest-trending foliation, approximately parallel to schistosity in schist, crosses contact with schist in several places, indicating contemporaneity of planar fabric in both units. Elongate patches of biotite and long dimension of augen form weak to strong lineation. Zircons from unit gave minimum ^{207}Pb/^{206}Pb age of 1,780 ± 20 Ma (Silver and others, 1961)

Ꝑ$_{P}$ms **Mica schist (Paleoproterozoic)**—About 80 percent of unit consists largely of dull-green to very dark brown, biotite–white mica–quartz schist, exposed in central part of map area. Other rock types include dark-brown to dark-brownish-green, fine-grained metasandstone; metavolcanic rocks; andalusite(?)-bearing white-mica schist; tourmaline-bearing white-mica schist; white mica–quartz schist; and greenstone containing blue-green actinolite. Metarhyolite or metarhyodacite is present 2.6 km north of Warm Spring (Silver and others, 1961). Most minor constituents form layers 2 to 60 cm thick. Metasandstone locally exhibits graded bedding and sparse cross-lamination. Unit locally is folded on scale of outcrop or smaller. Linear mineral alignments and crinkles are present in nearly every exposure. Zircon from metarhyolite or metarhyodacite gave minimum ^{207}Pb/^{206}Pb age of 1,720 ± 20 Ma (Silver and others, 1961)

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Addendum

Since the original publication of this report, additional information has come to light on ammonoids from the upper unit of the Owens Valley Group in Warm Spring Canyon. This information is based on ammonoids collected by one of the authors (Stevens) in 1972 at or near fossil locality 17 of this report (see table 2). Shortly after they were collected, these ammonoids were sent to the University of Iowa where they were examined as part of a Ph.D. dissertation by Lee (1975). In that work (Lee, 1975, p. 116) the following taxa were identified from the Warm Spring Canyon locality of Stevens: *Bransonoceras bakeri* Miller and Parizek, *Akmilleria electraensis* (Plummer and Scott), *Marathonites* (*Almites*) *sellardsi* (Plummer and Scott)?, and *Crimetes elkoensis* Miller, Furnish, and Clark?. According to Lee (1975), this ammonoid assemblage clearly indicates an Early Permian age and suggests correlation with the Clyde Formation of Leonardian age in Texas, which has a comparable ammonoid fauna. This correlation corroborates the probable Leonardian age interpretation given for the upper unit of the Owens Valley Formation in the main part of this report (see p. 23).

Figure 10 shows the Warm Spring Canyon ammonoid specimens illustrated by Lee (1975). The specimens of *Akmilleria electraensis* (fig. 10A,B) and *Bransonoceras bakeri* (fig. 10C,D) bear some resemblance to the specimens illustrated in Figures 9A and 9B, respectively, of this report, which were originally reported on and illustrated by Wrucke (1966). Better preservation of the specimens examined by Lee (1975), including partial preservation of sutures in *Akmilleria electraensis*, allowed more precise identification than was possible for Wrucke's specimens, which could not be identified to genus (see p. 21).

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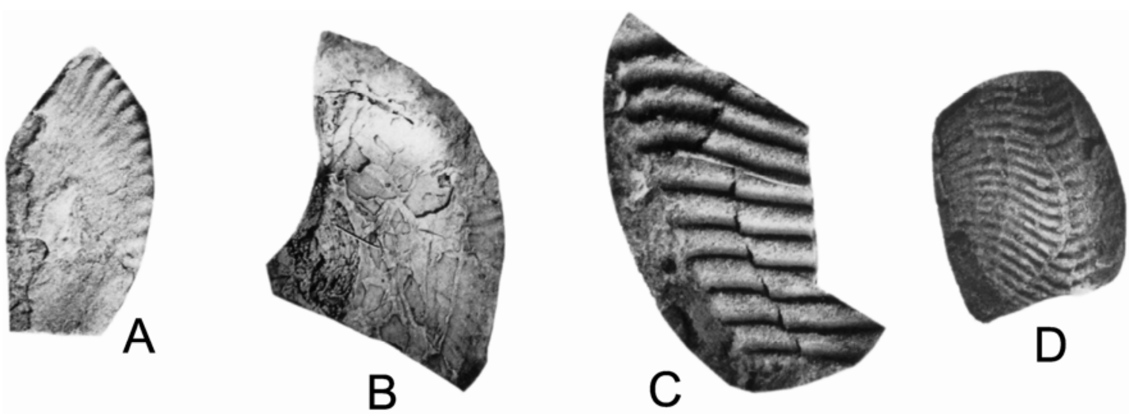

Figure 10. Ammonoid fragments from upper unit of the Owens Valley Formation in Warm Spring Canyon, at or near fossil locality 17 (see table 2). Photographs (all x1) are reproduced from Lee (1975). A, B, *Akmilleria electraensis* (Plummer and Scott); plate 9, figs. 10 and 11, respectively, of Lee (1975). C, D, *Bransonoceras bakeri* Miller and Parizek, 1948; plate 3, figs. 5 and 6, respectively, of Lee (1975).