Geology of the Connors Pass Quadrangle Schell Creek Range East-Central Nevada

By HARALD DREWES

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Stratigraphy and structure of a complexly deformed area in the Basin and Range province, and an evaluation of the tectonic environment in which it was developed

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By Harald Drewes

ABSTRACT

The Connors Pass quadrangle lies southeast of Ely and covers a moderately rugged area in the central part of the Schell Creek Range, largely north of Connors Pass.

The stratigraphic column in the area is about 5 miles thick and comprises rocks of late Precambrian to Recent age. Only a few hundred feet of upper Precambrian rock consisting of slightly metamorphosed quartzite and slate is present. Paleozoic rocks, about 23,000 feet thick and reflective of three major environmental changes, make up most of the section. The Paleozoic sequence begins with the clastic rocks of the Prospect Mountain Quartzite and Pioche Shale, whose grains are fine to moderately coarse and only moderately rounded and clean, and it continues with the dominantly calcareous rocks of the Pole Canyon Limestone and the Lincoln Peak Formation. Above this sequence is a unit of undifferentiated Cambrian and Ordovician age that is identical to Hague's (1883) Pogonip Limestone. Locally some of these formations are slightly metamorphosed.

The overlying middle Paleozoic rocks consist mainly of dolomite, but also include a little quartz sandstone composed of well-rounded clean grains. They include the Eureka Quartzite and Fish Haven Dolomite, of Ordovician age, the Laketown Dolomite, of Silurian age, and the Sevy Dolomite, Simonson Dolomite, and the Guilmette Formation, all of Devonian age. Disconformities probably separate the Silurian rocks from both older and younger rocks.

The upper Paleozoic rocks consist mainly of limestone but include much impure siltstone and sandstone containing poorly rounded grains. The impure sandstone first appears in the upper part of the Guilmette Formation, which is successively overlain by the Pilot Shale, of Devonian and Mississippian age, the Joana Limestone and Chainman Shale, of Mississippian age, the Ely Limestone, of Mississippian, Pennsylvanian, and Permian age, and the Rib Hill Sandstone and Arcturus Formation, both of Permian age. A little gypsum occurs in the Chainman Shale and possibly also in the upper part of the Arcturus Formation, which contains a thick wedge of chert-pebble conglomerate. The continuity of sedimentation was broken by disconformities probably near the end of the Devonian and certainly in Late Pennsylvanian time. The upper Paleozoic rocks were derived from areas at least partly exposed by orogenic uplift; such areas, probably 100–200 miles distant, lay northeast, west, and possibly southeast of the Connors Pass area.

The Mesozoic and Tertiary rocks are poorly represented. Between Middle Jurassic time and early Tertiary time the area was strongly uplifted, greatly faulted, and intruded by granophyric porphyritic rhyolite dikes, which are probably related to monzonite stocks in the adjacent ranges. At some time during the Cretaceous or Paleocene, black shale, tuffaceous sandstone, and conglomerate were deposited in a terrestrial environment.

During Eocene and Oligocene(?) time, as much as 6,000 feet of volcanic rock and terrestrial red conglomerate, originating in the Schell Creek Range and in the Snake Range immediately to the east, was deposited on a piedmont surface of moderate relief. Volcanism began with a relatively small extrusion of latite lava and tuff, which was followed by widespread extrusions of dacitic lava and tuff. Conglomerate of very local origin was deposited unconformably on the volcanics and the underlying red conglomerate, perhaps during Pliocene time. During Quaternary time, pediment and alluvial gravels were spread along the flanks of the Schell Creek Range and far into the adjacent valleys. During late Pleistocene time a pluvial lake occupied Spring Valley, east of the Schell Creek Range, and its lacustrine and beach deposits are probably interbedded with the youngest part of a gravel formation of Pleistocene age.

The rocks are much faulted and locally folded. Most of the structural features are relatively simple, but the outcrop patterns along the Schell Creek Range thrust fault are moderately complex because faults different in age and direction of movement share fault surfaces. The oldest faults, probably of Mesozoic age, are chiefly low-angle faults almost parallel to bedding planes. Beds adjacent to these faults dip into the fault planes in many places, and along the faults beds are cut out rather than repeated. Along the Schell Creek Range thrust fault, rocks as young as the Arcturus Formation are thrust onto rocks of Late Cambrian and Early Ordovician age, and in some places about 2.5 miles of strata is missing; in other places faults do not commonly cut out more than a few hundred to a few thousand feet of strata. Some faults seem to die out along their courses; some merge into adjacent thrust faults or are truncated by thrust faults higher in the section; and others abut against or merge into tear faults. Rocks within the thrust plates are broken by normal and reverse faults, many of which are restricted to a particular plate or group of plates. Northeastward or southwestward movement of the upper plates relative to the lower plates is indicated by the alinement of axes of drag folds, by tear faults, and by other minor structural features, but the direction of absolute movement is less certain; the upper plates may have been displaced northeastward with respect to the lower ones. All the thrust faults of this region have been dated as Late Jurassic or Early Cretaceous. The origin of the low-angle faults is not entirely clear, but regional relations indicate that the lower ones are thrust faults and that the upper ones may be either thrust faults or glide faults (faults directly of gravity origin in which horizontal stresses are absent or minor).

Some normal faults of middle Tertiary age may be as old as, or older than, some of the volcanic vents which are alined along

these faults. Other faults of this general age cut the Eocene conglomerate, and Eocene and Oligocene(?) volcanic rock and produced typical Basin and Range topography. One of the larger normal faults was locally deflected along a rather steeply inclined part of the Schell Creek Range thrust fault.

During middle or late Tertiary time, two plates of Tertiary rocks, the larger several square miles in extent, slid westward as much as 3,500 feet from an upfaulted part of the range onto the adjacent downdropped block. The plates moved partly along a surface close to the Schell Creek Range thrust fault and partly on the Chainman Shale of the upper thrust plate. The parts of the glide plates nearest their source were tilted with a reverse rotation, such as commonly occurs in slump blocks, and were warped into open synclines during the process of gliding. A few normal faults of late Tertiary and Quaternary age cut the glide plates and also the gravel older than the late Pleistocene lacustrine deposits. The only range-front fault recognized trends southwest and extends into the northeast corner of the area.

The mines in the quadrangle have yielded $1–$2 million worth of silver and a small amount of lead, copper, zinc, gold, antimony, and tungsten. In the principal mining area, the Taylor district, silver ore is scattered in silicified bodies that replace parts of the Guilmette Formation and the Joana Limestone near northeast-trending faults. The mineralization may be genetically related to the granophyric porphyritic rhyolite dikes, which are concentrated in the district. Some mineralization occurs along Cleve Creek, near Majors Place, and in Tamberlaine Canyon. A cursory spectrographic study reveals anomalously high amounts (as much as 0.1 percent) of several metals, including silver, copper, and zinc, in some parts of the Chainman Shale, but these anomalies are so localized that they seem to be related to sedimentation or diagenetic processes rather than to later mineralization. Reconnaissance geochemical prospecting indicates that average background values for copper, lead, and zinc in alluvium are 20 ppm (parts per million), 25 ppm, and 25 ppm, respectively. Slightly greater amounts of zinc are in alluvium derived from Chainman Shale.

INTRODUCTION

The Connors Pass quadrangle (fig. 1), an area of 207 square miles, lies 7–25 miles southeast of Ely, in east-central Nevada. It straddles the south-central part of the Schell Creek Range. Spurr (1903, p. 38–47) briefly described parts of the range. Young (1960a, b) mapped the northern part of the range, and Tschanz and Pampeyan (1961) mapped the extreme southern part of the range. Brief descriptions of several of the mining districts in the range have been published (Hill, 1916; Roberts, 1942; Couch and Carpenter, 1943). Slightly more geologic work had been done in the Snake Range to the east and considerably more in the Egan Range to the west than had been done in the Schell Creek Range, but less than a fifth of the total area of the three ranges had been mapped when this study began.

OBJECTIVES AND METHODS

This area was selected for geologic study chiefly because it is representative of a part of the Great Basin that contains a variety of complex and unusual structural features. Few geologic studies have been made in this general region, and these have been largely of reconnaissance type. The area was especially attractive because of its location between two other areas in which detailed geologic studies are currently in progress; one of these is in the southern Snake Range, to the southeast, and the other in the Egan Range, to the west. The combined knowledge obtained by the three studies should lead to a fairly comprehensive understanding of the stratigraphy, structure, and mineral deposits of a large part of eastern Nevada.

This geologic study was designed to obtain a representative stratigraphic section, to determine the relations between numerous low-angle faults, steep normal faults, intrusive bodies, and mineral deposits, and particularly to gain a better understanding of the origin of the low-angle faults. A geochemical reconnaissance was also made for guidance in prospecting.

The report is based on 190 days of fieldwork from the fall of 1958 through the fall of 1961. The geology was mapped (pl. 1) on a topographic base at a scale of 1:48,000. Geologic mapping was facilitated by the use of a sketchboard and on open-sight alidade. High-altitude aerial photographs were used chiefly in mapping valleys, for elsewhere the geologic features are too complex to be mapped on the photographs. About 750 samples were collected; these consisted of three groups roughly equal in number—petrographic, geochemical, and paleontologic.

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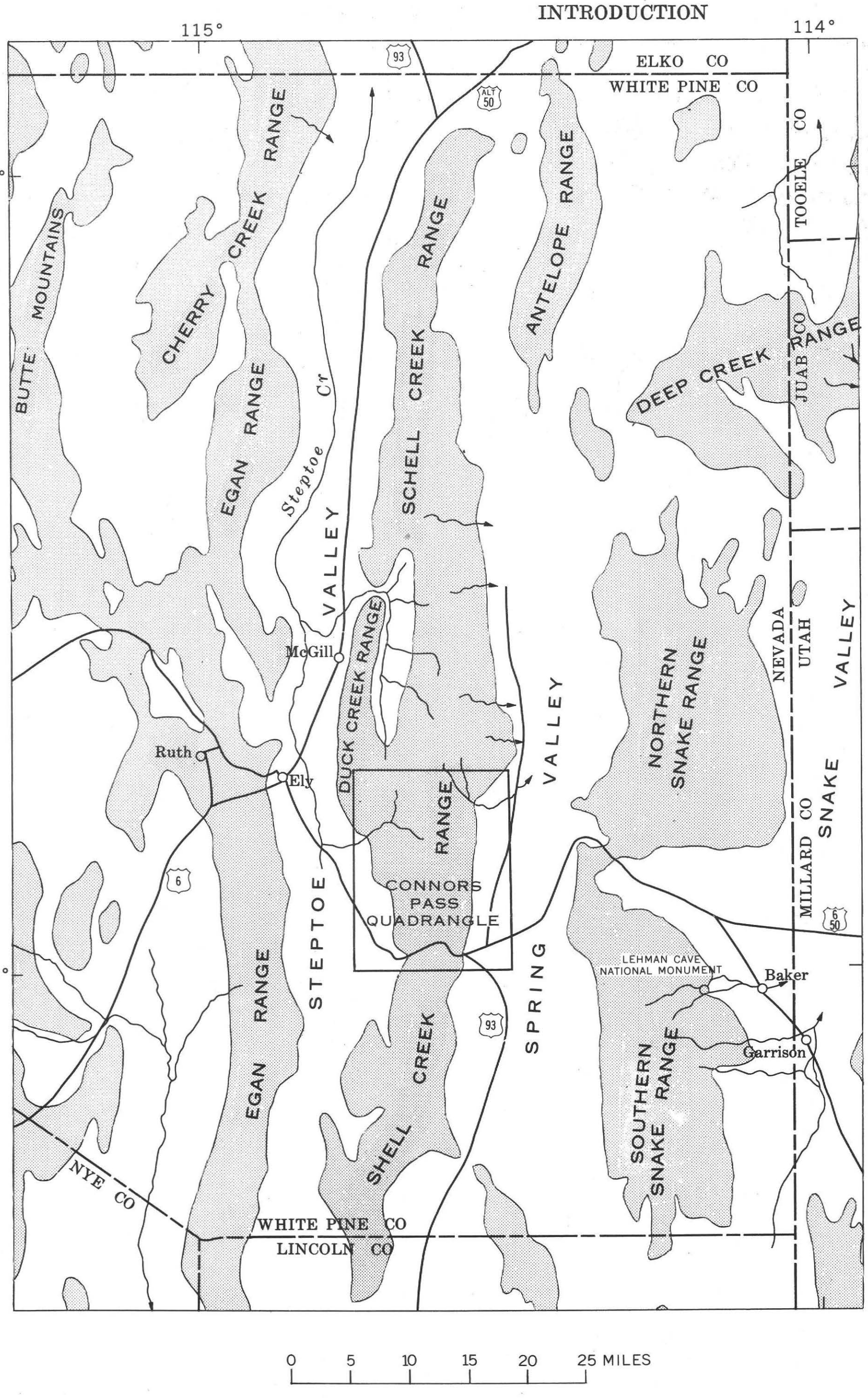
I was assisted in the field by J. O. Gallagher in 1959, B. M. Irvine in 1960, and E. J. Young in 1961. I also benefited from conversations in the field with W. H. Easton, Carew McFall, A. R. Palmer, R. K. Hose, and H. R. Christner. Mackenzie Gordon, Jr., furnished the unpublished results of his studies of a suite of fossils collected in the area by him, Hose, and C. A. Repenning.

GEOGRAPHY

Linear mountain ranges, separated by broad, flat valleys, are typical physiographic features of the central part of eastern Nevada. The Connors Pass quadrangle is in a part of the Schell Creek Range that is transitional between a more rugged topography to the north and a less rugged topography to the south. Spring Valley, which lies east of this quadrangle, forms a closed drainage basin having a floor less than 6,000 feet above sea level. Steptoe Valley, west of the range, is largely at an altitude of less than 7,000 feet and drains northward into two closed depressions.



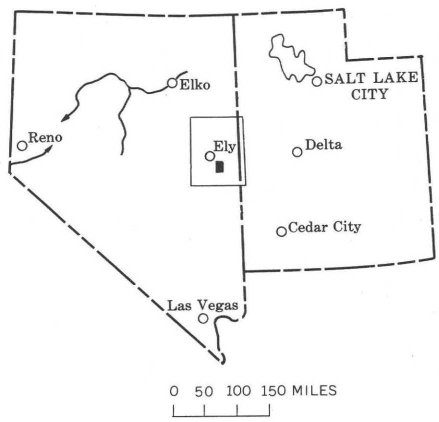


FIGURE 1.—Index map showing location of the Connors Pass quadrangle.

Many roads and tracks give relatively easy access to most of the area, although only one ranch and Majors Place, are inhabited. Paved highways cross Connors Pass and lie close to most of the east and west edges of the quadrangle. Several graded access roads and many smaller roads are shown on the topographic map. There are also several unmapped tracks that are passable to vehicles having four-wheel drive: to Summit Spring from the southeast; eastward nearly to East Canyon Spring; to Crethers Springs from the north; from the nearest part of the Nevada National Forest boundary to a treeless area about 1.75 miles southwest of Bastian Spring; along the fence on the western boundary of the national forest.

Eastern Nevada has a semiarid temperate climate with a strongly seasonal orographic influence. Climatic data have been recorded by the U.S. Weather Bureau (1957) at several stations near the quadrangle for at least 10 years, and a few years ago a station at which only general data are collected was set up at Connors Pass.

There are some streams and springs in the mountains of the northern half of the quadrangle, but observations made during years of average to below-average rainfall show that water also flows perennially in a half-mile segment of the canyon northwest of the Kolcheck mine in the Cleve Creek area, in a 1-mile segment of the canyon midway between Kolcheck Basin and Cottonwood Spring, in all the valley extending from Clear Spring to Steptoe Creek, and in a half-mile segment below the spring in an unnamed valley tributary to Steptoe Creek from the west. There are also moderately large unmapped springs a quarter of a mile southwest of East Canyon Spring, and in the mouth of a cave in the narrow canyon midway between the Kolcheck Basin and Cottonwood Spring. The spring mapped half a mile north of the abandoned Taylor townsite has been dry for years. The largest springs in the quadrangle are Cave Springs, Clear Spring, and the unnamed spring a third of a mile northwest of Clear Spring; all are near the junction of Cave Creek and Steptoe Creek.

GEOLOGIC SETTING

Upper Precambrian and Paleozoic sedimentary rocks underlie most of the mountainous part of the quadrangle, but Tertiary volcanic and sedimentary rocks and Quaternary deposits (table 1) underlie small

TABLE 1.—*Generalized section of rocks exposed in the Connors Pass quadrangle*

Age		Formation	Thickness, estimated (feet)	Lithology
Quaternary	Recent	Alluvial sand and gravel.	0–100+	Unconsolidated sand and gravel in lowest stream terraces.
			Unconformity	
	Pleistocene	Lacustrine silt, sand, and gravel.	0–100+	Unconsolidated silt and sand, and bars of rounded gravel.
			Unconformity(?)	
		Younger alluvial and fan gravel.	0–100+	Very weakly indurated sand and gravel in low stream terraces and fans that have a slightly mature morphology.
			Unconformity(?)	
		Older alluvial and fan gravel.	0–200+	Very weakly indurated sand and gravel in high stream terraces and fans that have a mature morphology.
			Unconformity	
Tertiary	Pliocene(?)	Fanglomerate	0–600+	Weakly indurated fanglomerate of locally derived coarse subangular fragments.
			Unconformity	
	Oligocene(?) and Eocene	Dacite vitrophyre	0–1, 200+	Lava flows, tuff, and intrusive bodies.
	Eocene	Latite tuff	0–2, 000	Light-gray massive tuff and local lenses of welded tuff.
		Quartz latite vitrophyre.	0–300	Lava flow underlying conspicuous brownish-gray cliff.
		Conglomerate	0–100+	Reddish-gray to yellowish-gray conglomerate of coarse subangular fragments derived locally and from adjacent ranges; contains a lens of fresh-water limestone near base and lenses of tuff near top.
			Unconformity	
		Basaltic andesite vitrophyre.	----	One small dike.
			Unconformity(?)	
Tertiary or Cretaceous		Shale and tuffaceous sandstone.	50±	Black shale and mudstone and light-gray tuffaceous sandstone in one small downfaulted block.
			Unconformity	
Tertiary(?), Cretaceous, and Jurassic		Porphyritic rhyolite	----	Altered granophyric dikes, probably associated with silicification and mineralization and possibly associated with an unexposed plutonic body.

TABLE 1.—*Generalized section of rocks exposed in the Connors Pass quadrangle*—Continued

Age			Formation	Thickness, estimated (feet)	Lithology
				—Unconformity—	
Permian		Early Permian	Arcturus Formation	2, 000–3, 000+	Pale-yellow-brown siltstone and sandstone and interbedded medium-gray shaly limestone; possibly contains a little gypsum; 500–1,000 ft; probably as young as Leonard age. Reddish-brown conglomerate of rounded chert and quartzite pebbles derived from a distant source; 0–500 ft. Medium-gray shaly limestone, reef limestone, and pale-yellow-brown limy siltstone; 1,500–2,000 ft; of Wolfcamp age.
			Rib Hill Sandstone	1, 000+	Pale-yellow-brown fine-grained sandstone.
			Ely Limestone	2, 500±	Light-gray limestone and pale-yellow-brown siltstone; contains thin local basal conglomerate; 500–800 ft; of Wolfcamp age. Light-gray shaly limestone and pale-yellow-brown siltstone. *Chaetetes*-bearing beds at base; 500–600 ft.
Carboniferous	Pennsylvanian and Mississippian	Middle and Early Pennsylvanian		—Unconformity—	Light-gray to medium-gray medium-to-thick-bedded cherty limestone; 1,400 ft; locally includes some rocks of Mississippian age at base.
	Mississippian	Late Mississippian	Chainman Shale	1,100±	Dark-gray clay-shale containing a little sandstone, conglomerate, and limestone, and traces of gypsum.
		Early Mississippian	Joana Limestone	300–500	Medium-gray bioclastic cherty massive to thick-bedded limestone.
	Mississippian and Devonian	Early Mississippian and Late Devonian	Pilot Shale	320–480	Dark-gray shale and siltstone that weather to yellowish-brown and reddish-brown plates; contains a little quartzite, argillite, and limestone.
				–Unconformity (?)	
Devonian		Late and Middle Devonian	Guilmette Formation	2, 000±	Sandstone and conglomerate overlain by shaly limestone and reef limestone; 600± ft; only very locally preserved. Bluish-gray cherty limestone and shaly limestone; local lenses of dark-brown coarse-grained dolomite; 400–600 ft. Dark-brown coarse-grained dolomite, interbedded shaly limestone, and a sandstone unit; 300–700 ft. Blue-gray thick-bedded cliff-making limestone above a basal silty dolomitic limestone; 500–600 ft.
		Middle Devonian	Simonson Dolomite	600–700	Light-brown and dark-brown laminated coarse-grained dolomite; 150+ ft. Dark-brown coarse-grained cliff-forming dolomite; 100+ ft. Light-brown and dark-brown laminated coarse-grained dolomite; 150+ ft. Pale-brown coarse-grained dolomite; 200+ ft.
		Middle(?) and Early Devonian	Sevy Dolomite	900+	Medium-gray fine-grained dolomite that weathers light gray.
				—Unconformity —	
Silurian		Middle and Late Silurian	Laketown Dolomite	600–700	Light-gray coarse-grained thick-bedded dolomite.
				—Unconformity(?)—	
Ordovician		Late Ordovician	Fish Haven Dolomite	400–500	Dark-brown coarse-grained dolomite; contains interbedded gray dolomite near the top.
		Middle Ordovician	Eureka Quartzite	300–400	Very light gray to pinkish-gray thick-bedded quartzite.

TABLE 1.—*Generalized section of rocks exposed in the Connors Pass quadrangle*—Continued

Age		Formation	Thickness, estimated (feet)	Lithology
Ordovician and Cambrian	Middle and Early Ordovician and Late Cambrian	Limestone	3,000±	Medium-gray fine-grained bioclastic, highly fossiliferous limestone; 300–400 ft. Dark-olive-gray fissile shale, interbedded fossiliferous limestone, and a little sandstone and conglomerate; 400± ft. Medium-gray limestone and siltstone, interbedded with shale and a little intraformational conglomerate; 1,100–2,000 ft. Light-medium-gray thick-bedded cherty limestone; 1,000–1,100 ft. Light-medium-gray thin-bedded limestone, silty partings; 200–800 ft. This formation is equivalent to the Pogonip Limestone of Hague (1883) and approximately equivalent to the Ordovician Pogonip Group and the Cambrian Windfall Formation (Nolan and others, 1956).
Cambrian	Late and Middle Cambrian	Lincoln Peak Formation	1,600–1,800	Shale and dark-gray thin-bedded highly fossiliferous limestone. Medium-dark-gray medium-thick-bedded limestone; pinches out to south; 0–250 ft. Medium-gray to pale-olive-gray slightly limy shale.
	Middle Cambrian	Pole Canyon Limestone	1,500–2,000	Medium-gray fine- to medium-coarse grained limestone; 100–300 ft. White to very light gray saccharoidal limestone or low-grade marble; contains thin gray limestone and local lenses of dolomite; 1,200± ft. Medium-dark-gray fine- to medium-coarse grained limestone; contains a few local quartzite beds near base; 100–200 ft.
	Middle and Early Cambrian	Pioche Shale	250–315	Olive-black to greenish-gray quartzitic shale and siltstone; contains some interbedded quartzite and limestone.
	Early Cambrian	Prospect Mountain Quartzite	4,000–5,000	Light-gray to dark-purplish-gray thick-bedded crossbedded quartzite and metaquartzite; contains some shale near top and conglomeratic quartzite near base; locally contains some bodies of altered metadiabase; as mapped may include some rocks of Precambrian age.
Late Precambrian		Metasedimentary rocks	700–800	Phyllitic shale, metaquartzite, and slaty argillite.

areas. The major valleys are almost entirely underlain by Quaternary deposits. Within the quadrangle the Precambrian and Paleozoic rocks form a marine sequence almost 5 miles thick that consists dominantly of limestone and dolomite but that also contains considerable quartzite and shale. The lithology and thickness of the sequence is fairly uniform throughout much of western Utah and eastern Nevada. The continuity of sedimentation was broken, however, by disconformities in Middle and Late Pennsylvanian time, probably by one in the Early Silurian and one in the Late Silurian, and possibly by one in the Late Devonian. Rocks of Late Permian through early Tertiary age are, with few exceptions, absent. Middle Tertiary through Quaternary time is represented by a sequence of continental conglomerate, lava flows, intrusive bodies, and tuffs having a combined thickness of more than a mile. Angular unconformities are numerous in the younger sequence but are poorly dated. The surficial deposits in the valleys are largely a continuation of Tertiary sedimentation. They are poorly consolidated, are derived from nearby sources, and are interbedded with fine lacustrine sediments deposited in a pluvial lake.

The sedimentary rocks are cut by many low-angle faults and by many steep faults. The low-angle faults are largely thrust faults parallel or almost parallel to bedding planes, and they generally reduce the thickness of the sequence of Paleozoic rocks, in some places by as much as 2.5 miles; only rarely are beds repeated along these faults. Because of this fact, the thicknesses given for the older formations can only be estimated and are commonly the minimum amounts. Other low-angle faults are glide faults; "glide fault" is a term here used to describe rootless faults that were formed near the surface under the direct influence of gravity. Many normal faults and some glide faults cut the thrust faults or are nearly parallel to them. Complex outcrop patterns were formed near segments of faults along which movement has been recurrent and varied in direction. Movement along some normal faults during late Tertiary and Quaternary times caused linear physiographic features, but not all such features are related to faults. During the Tertiary when normal faulting was frequent, magmas ranging in composition from rhyolite to dacite were intruded into the older strata and were extruded as lava and tuff. Siliceous material that partially replaced limestone and that locally contains ores of base metals and silver is closely associated with the oldest intrusive rock.

STRATIGRAPHY

PRECAMBRIAN SYSTEM

METASEDIMENTARY ROCKS

Less than 800 feet of metasedimentary rocks of late Precambrian age underlies a few square miles in the northeastern corner of the quadrangle, and 15 miles north of the quadrangle similar rocks form the upper part of the Piermont Group of Young (1960b, p. 158). Pending a more extensive study of the Precambrian rock in the intervening area, no formal names are applied to it in this quadrangle. I have probably assigned less rock to the Precambrian and more to the Cambrian than Young has done, for his uppermost shale unit seems to be lenticular and to be absent in this quadrangle, and the quartzite beds above and below it are lithologically similar and cannot be separated where the shale unit is absent.

The lowest 500–600 feet of the Precambrian rock exposed in the quadrangle consists of impure quartzite, conglometratic quartzite, a little conglomeratic shist, and argillite, which are grayish brown [1] and underlie moderately steep slopes. The pebbles in the congomeratic rocks are moderately well rounded, are less than 1 inch in diameter, and consist of white quartz and quartzite. They are scattered or poorly sorted and lie in a matrix that ranges in composition from relatively pure quartz sand to silt. The quartzitic beds are a few inches to a few feet thick, but the schist and argillite beds are thinner.

The quartzite, schist, and argillite unit is overlain by about 200 feet of slaty argillite and quartzitic phyllite, which are generally also grayish brown, but which form a bench and gentle slopes above the more resistant quartzite, schist, and argillite. A few slaty beds are more colorful—dusky blue to purplish gray, pale bluish green, and pale yellowish orange.

The Precambrian rock is slightly metamorphosed. Phyllitic foliation and slaty cleavage are conspicuous in many of the fine-grained rocks. The cleavage commonly parallels the bedding, but in some rocks diverges as much as 30° from bedding. Chlorite is abundant in the groundmass of this rock and also forms some larger pods, as seen in thin section. The grains of the quartzitic rock interlock, and in the phyllitic rock biotite lies between the quartz grains.

The metasedimentary rocks begin about 5,000 feet below *Olenellus*-bearing beds and are the youngest fairly thick argillaceous rocks beneath the Prospect Mountain Quartzite. They are considered late rather than early Precambrian because of their apparent conformity to overlying Paleozoic rocks and their contrast to another older sequence of moderately metamorphosed rocks in the southern end of the Deep Creek Range (Nelson, 1959; Bick, 1958) about 30 miles to the northeast.

Rocks similar to, and possibly correlative with, the metasedimentary sequence in the Schell Creek Range were named the Goshute Canyon Formation and Horse Canyon Formation in the Deep Creek Range (Bick, 1959), and were the basal quartzite and the overlying slate or shale member of the lower part of the Prospect Mountain Quartzite in the Gold Hill district (Nolan, 1935, p. 6). Correlative rocks also appear in the southern Snake Range (Drewes, 1954; 1958, p. 224–225) and in the Egan Range (Fritz, 1960; Woodward, 1962). Individual units of upper Precambrian rock in these four ranges cannot be definitely correlated, because the units apparently wedge out or change facies within 10–30 miles.

CAMBRIAN SYSTEM

PROSPECT MOUNTAIN QUARTZITE

Quartzite, 4,000–5,000 feet thick, that lies with apparent conformity on the phyllite at the top of the metasedimentary Precambrian rocks, constitutes the Prospect Mountain Quartzite. The name was first applied by Hague (1883, p. 253–254) near Eureka, Nev., and has since been widely used in Nevada and Utah. Hague, however, was unable to define the base of the formation, because only its upper 1,500 feet is

[1] All colors used in this report are from Rock-Color Chart (Goddard and others, 1948).

exposed in the type locality. The lower contact of the Prospect Mountain was selected at six places in eastern Nevada and western Utah; though at each place the contact is marked by similar lithologic changes, each selection actually may represent different horizons. At each of the six places, as well as in the Connors Pass quadrangle, the base of the Prospect Mountain Quartzite is placed at the top of the highest mappable unit of shale, schist, or argillite beneath the main body of quartzite. At the base of the formation the change from phyllitic rock upward to quartzite is either sharp or transitional over a few feet. The thickness assigned to the formation on this basis ranges from 2,000 feet in the Sheeprock Range, Utah (Cohenour, 1957, 1959), to at least 5,000 feet in the Promontory Range, Utah (Olson, 1956). This wide variation in thickness suggests that the argillaceous units are lenticular and that a given lens may be present and well exposed in one district but not in others.

In the northeast part of the Connors Pass quadrangle and for several miles north of it, much of the middle part of the Prospect Mountain Quartzite is poorly exposed on a large dip slope. A thin phyllitic shale unit 2,000–3,000 feet beneath the top of the formation that appears 15 miles to the north (Young, 1960b) is either covered or absent in this quadrangle.

The Prospect Mountain Quartzite underlies about 10 square miles of the quadrangle, mostly along Cleve Creek in the northeast corner of the area; it also crops out topographically low along the east flank of the range between Majors Place and the mouth of Cooper Canyon, and in Mosier Canyon in the northwest corner of the quadrangle. The quartzite generally forms extensive uniform slopes that are interrupted by a few castellated cliffs. Many of these slopes are covered by rubble fields whose fragments have conspicuous grayish-red centers and light-brownish-gray rims. Less typically, as north of Majors Place, the Prospect Mountain weathers to very light gray slopes that are not covered with rubble and are broken only by scattered small ledges.

The Prospect Mountain consists mainly of fairly uniform quartzite or metaquartzite alternating with thin beds or partings of shale or argillite. The fresh quartzite is mostly very light gray to pinkish gray but less commonly greenish gray or dark purplish gray, and the interbedded shale is dark brownish gray to brownish black. Upon weathering, the very light gray to pinkish-gray quartzite does not change color, but the shale and argillite become pale red, pale yellowish brown, or brownish black. The quartzite beds are commonly 3–4 feet thick but range in thickness from 1 to 6 feet, and generally are crossbedded. North of Majors Place most of the foreset beds dip westward as they do in the Wheeler Peak area, 15 miles to the southeast (Drewes, 1954, p. 17), but in the Cleve Creek area no direction appeared to be dominant.

Most grains are subangular to subrounded moderately coarse to very coarse sand-sized quartz, but some scattered grains are granule and pebble sized; a few thin beds contain abundant pebbles. The shale and some beds of quartzite contain biotite, chlorite, sericite, iron oxides, sphene, and zircon; apatite crystals are included in some grains of quartz. The quartz gains are slightly strained, are interlocking, but rarely are interpenetrating. The larger clastic grains are commonly separated by smaller secondary grains. The mica is mostly intergranular, but some flakes of sericite penetrate the secondary quartz grains. The mineralogy and texture of the quartzite indicate a slight metamorphism, and these features will be compared with the characteristics of younger quartzites and sandstones.

Such lithologic variations in the Prospect Mountain are both lateral and vertical. In the Cleve Creek area, pebbles are most common in the lower half of the formation and shaly beds are most common near its top, about 20 percent of the uppermost 300 feet being quartzitic shale. North of Majors Place, however, all but the uppermost few tens of feet of the formation consists of relatively clean very light gray quartzite. The degree of cementing or induration also is varied, and locally the prevailing rock is not a quartzite but a sandstone that weathers to friable slightly rounded fragments.

Altered greenish-gray diabase is scattered in some of the colluvium low on the south wall of Mosier Canyon. The local distribution of the blocks indicates that they have been weathered from a small body within the quartzite. This isolated occurrence is noteworthy because none of the Tertiary igneous rock in the area is similar to the diabase, and because similar rock was described by Morris (1957, p. 4) and by Morris and Lovering (1961, p. 15–16) from the Tintic Quartzite in the East Tintic Hills in central Utah and by Kellogg (1960, p. 190) from the Patterson Pass area in the southern Schell Creek Range. The rock has a moderately coarse ophitic texture and consists dominantly of laths of strongly kaolinized, probably albitized, plagioclase and of interstitial clinopyroxene that has largely been altered to actinolite. It also contains small amounts of apatite and magnetite, xenocrysts of quartz, and interstitial secondary calcite, quartz, chlorite(?), and another amphibole(?). These occurrences of diabasic rock in the Prospect Mountain Quartzite suggest that magmatic activity occurred at widely scattered places in the region during Early Cambrian time, perhaps simultaneously.

In eastern Nevada the Prospect Mountain Quartzite is of Early Cambrian age but may possibly include Precambrian rocks near its base. In my opinion the entire formation should be designated Early Cambrian(?).

The Prospect Mountain Quartzite was recognized in the Egan Range by Fritz (1960), in the Deep Creek Range by Nolan (1935), Bick (1958), and Nelson (1959), in the southern Snake Range by Drewes and Palmer (1957), and elsewhere in the Schell Creek Range by Young (1960b) and Kellogg (1960).

The formation was deposited in a sea which seems, because correlative beds of quartzite to the east are younger, to have transgressed eastward. The type of sedimentation was continuous from late Precambrian time to Cambrian time, and vast quantities of quartz sand and a few pebbles were deposited under conditions that remained constant for a long time. The material in the thicker shaly and schistose beds may mark the occasional influx of mud and silt from a generally subsiding land during minor regressions of the sea, but that in the thinner shaly beds and partings was probably winnowed from the sand by generally westward flowing currents. The Prospect Mountain sediments apparently came from the east. The relative purity of the sand indicates that the source was distant, or perhaps that the sand was recycled. Upper Precambrian quartzite in the Wasatch Mountains may have been the immediate source of the quartz sand before it was submerged.

PIOCHE SHALE

The Pioche Shale conformably overlies the Prospect Mountain Quartzite. It is 250–315 feet thick, and is transitional between a thick quartzitic sequence and a thick carbonate sequence. It was defined by Walcott (1908) and was redefined by Westgate and Knopf (1932).

The Pioche Shale underlies a narrow belt on the east side of the Schell Creek Range between the mouth of Cooper Canyon and Majors Place, and several small areas along Cleve Creek; it commonly forms gently sloping benches on which outcrops are scarce. The shaly lithology of the Pioche is partly masked by quartzite and limestone chips weathered from resistant layers within the formation and from the overlying Pole Canyon Limestone. North of Majors Place the weathered shale is brownish gray and contrasts strongly with the almost white quartzite and gray limestone, but in the Cleve Creek area the color of the weathered shale blends into that of the weathered quartzite.

The Pioche consists mainly of silty shale but contains quartzite and limestone. Its contact with the Prospect Mountain Quartzite is gradational and is placed at the top of the uppermost thick bed of quartzite. In a transitional zone extending a few tens of feet above this contact, there is a larger proportion of quartzite, but shale gradually becomes dominant upward. The general lithologies are described in order of decreasing abundance. The distribution of these rock types in the four stratigraphic sections (fig. 2) shows that the thinner units are discontinuous.

More than half the formation consists of olive-gray, greenish-gray, or olive-black micaceous shale or siltstone that weathers light brown to pale yellow brown. South of Cooper Canyon, and locally to the north, the shale is phyllitic and contains small clusters of metamorphic minerals. The beds are commonly $\frac{1}{16}$–$\frac{1}{2}$ inch thick. The quartzite interbedded in the lower part of the Pioche resembles that of the underlying Prospect Mountain except that it is largely thin bedded and impure. The quartz grains are seen in thin section to be generally less than 0.1 mm (millimeter) in diameter and slightly flattened or elongate parallel to the bedding; these grains either interlock or are separated from one another by sericite, chlorite, and iron oxide. Schistose texture is common in some areas, and some shale also has a cleavage inclined at a moderate angle to the schistosity.

In the upper half of the Pioche there are a few layers of yellowish-brown impure limestone 5–15 feet thick. Some of these contain small bioclastic lenses rich in small fragments of trilobites. Near the top of the formation there are also a few thin beds or lenses of gray platy limestone similar to that in the overlying Pole Canyon.

Sections *B* and *C* of figure 2 were measured with a Jacob staff and hand level; the thicknesses of units in the other sections were estimated in the field and scaled from the map.

Part of the difference in thickness of these sections may be due to minor faults (fig. 2) or to minor unmapped bedding-plane faults, but some of it may be due to variations in original deposition. The consequent difficulty in obtaining a complete and representative stratigraphic column persists in all the Paleozoic formations; some units in these formations show so much lateral variation that a single section of a unit cannot be regarded as representative for the whole area.

Fossils were collected from the Pioche Shale at only four places and were all identified by A. R. Palmer

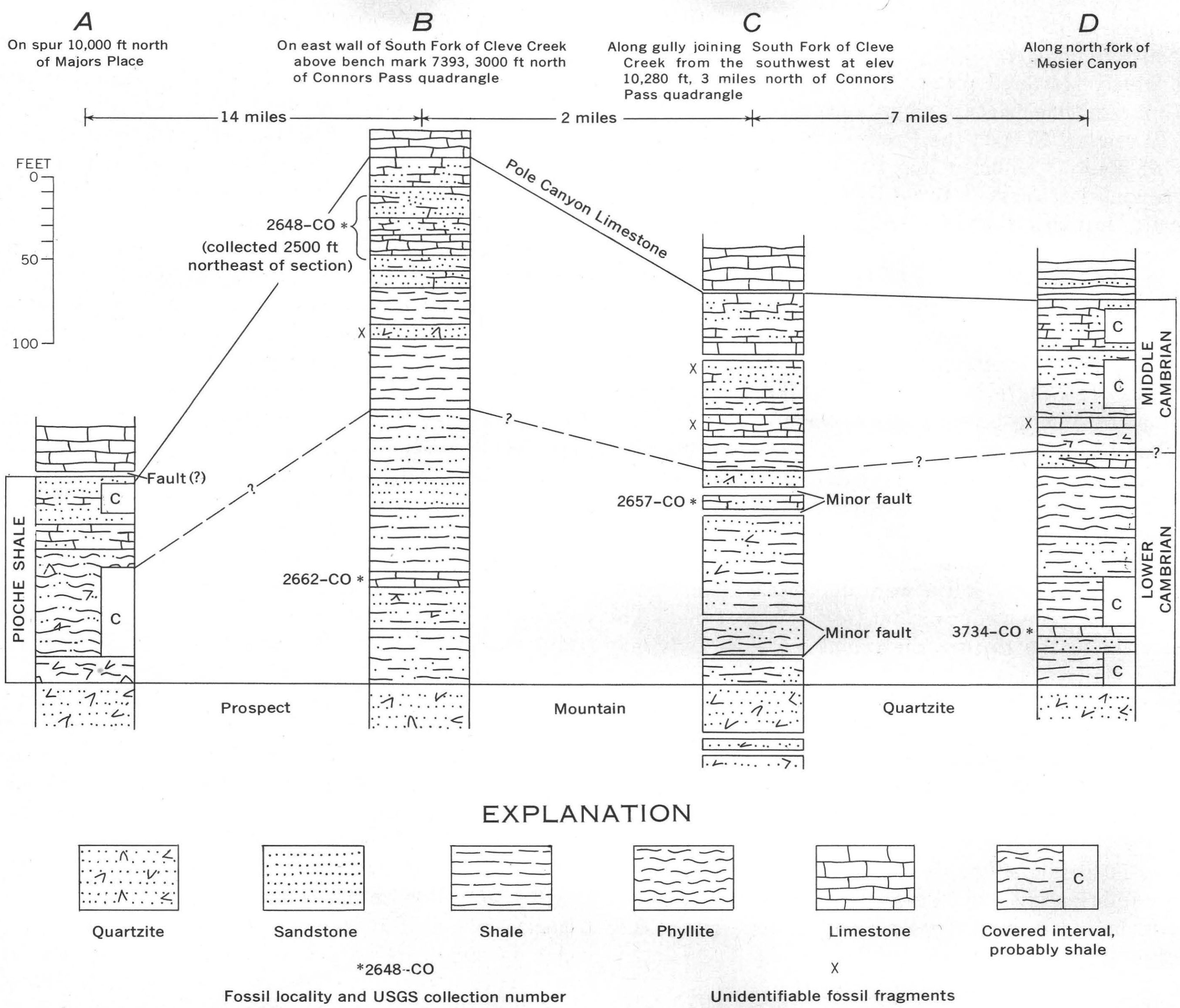


FIGURE 2.—Stratigraphic sections of Pioche Shale.

(written commun., 1962). Two collections, 2657–CO and 2662–CO (fig. 2), contain fragments of olenellid trilobites that indicate an Early Cambrian age; one collection also contains *Bonnia*. Collection 3734–CO contains *Olenellus* sp. and a ptychoparoid trilobite referable to *Crassifimbra*, also of Early Cambrian age. Collection 2648–CO contains fragments of several ptychoparoid trilobites without olenellid material and is probably of very early Middle Cambrian age.

The Pioche Shale has been recognized over a wide area around the quadrangle; its lithology is generally the same everywhere, but the formation is probably slightly younger to the east than to the west.

The silt and clay that formed most of the Pioche Shale are presumed to have been reworked from the underlying and adjacent quartz sands and to have been deposited in a shallow eastward-transgressing sea.

POLE CANYON LIMESTONE

The Pole Canyon Limestone, a white-to-gray saccharoidal limestone of Middle Cambrian age as much as 2,000 feet thick, conformably overlies the Pioche Shale. It was defined by Drewes and Palmer (1957) in the southern Snake Range. It is the oldest formation of the thick dominantly limestone sequence that was deposited in the region almost continuously from Middle Cambrian to Middle Ordovician time.

The limestone underlies about 6 square miles of the area; it crops out chiefly in the lower slopes along the east flank of the Schell Creek Range between Majors

Place and Cleve Creek, but also in smaller areas near the Kolcheck mine and high in the wall of Mosier Canyon. The formation commonly forms large rounded bosses or low gray cliffs and benches, but in Mosier Canyon it forms prominent white to medium-gray cliffs. These cliffs are conspicuous more because of their white color than because of their prominence; at a distance they can readily be distinguished from the higher and more uniformly gray cliffs of the Cambrian and Ordovician limestone above the overlying Lincoln Peak Formation.

Although the Pole Canyon Limestone consists mainly of massive limestone, its lower contact is gradational; thin beds of limestone are intercalated in the uppermost few tens of feet of the underlying Pioche Shale, and a small amount of quartzitic shale is locally interbedded near the base of the Pole Canyon. The base of the Pole Canyon is placed at the bottom of the lowest thick unit of limestone, a position that is never ambiguous.

None of the sections of the Pole Canyon Limestone is complete; in every section one or both contacts are faulted, generally by minor thrust faults that are almost parallel to bedding. The lithologic sequence within the formation is therefore uncertain, but along Cleve Creek, where the formation is thickest, it comprises a basal medium-gray unit 100–200 feet thick, a middle white to light-gray massive unit 400–500 feet thick that contains a few beds of darker limestone, and an upper medium-gray unit 100–300 feet thick. These three units and some of the subdivisions of the middle unit are locally mappable, but none seems to extend continuously over large areas. In the Mosier Canyon area, about 150 feet of limy shale resembling the upper half of the Pioche Shale, but containing younger fossils, lies 300–400 feet above the base of the Pole Canyon Limestone.

The limestone consists almost wholly of calcite, but some of it is dolomitic; the darker rocks contain very small amounts of black opaque material, which is probably organic, and some grains of detrital quartz and secondary sericite and limonite. The few quartzitic shale beds near the base of the Pole Canyon resemble the Pioche Shale in that sericite and chlorite are abundant, are alined, and penetrate through the quartz grains.

Much of the limestone is massive or thick bedded, but the beds in the darker units are commonly only 2–12 inches thick. Primary sedimentary features are scarce; for example, fine siliceous or silty laminae appear in some parts of the Pole Canyon but do not seem to be of great horizontal extent. In a few places there are small dark spheroidal blebs resembling remnants of *Girvanella*. Irregular to lenticular pods of limestone varied in color or texture may be relicts of sedimentary features, but they are difficult to distinguish from sheared pods of structural origin that occur in some places. Intense diagenetic change—or more likely very low grade metamorphism—has obliterated other sedimentary features.

The most distinctive feature of the limestone in the Pole Canyon is its saccharoidal texture. Much of the rock is coarsely crystalline, but the darker limestone is generally fine to medium grained. The calcite is seen in thin section to form interlocking anhedral to subhedral grains, some of which are alined, sheared, or veined by other calcite crystals that are commonly elongated normal to the general foliation. In the exposures south of Cooper Canyon, coarse-grained yellowish-brown dolomite makes up 1–15 percent of the formation. Dolomite also occurs in smaller amounts in some parts of the Cleve Creek area and is abundant north of Mosier Canyon. Some of the dolomite occurs in small pockets 1–2 inches across, but some forms lenses a few tens of feet thick and many tens to hundreds of feet long. One lens, which extends across the mouth of Cooper Canyon, is more than 100 feet thick and more than half a mile long. Most contacts between limestone and dolomite parallel bedding planes, but some cut across the bedding. Some dolomite lenses enclose small angular blocks of limestone, and in places the contacts between limestone and dolomite are intricately embayed. The distribution of the dolomite shows no relation to the thrust faults, and in several places there are fault breccias containing fragments of both limestone and dolomite—these features indicating that the dolomite is older than the fault.

The Pole Canyon Limestone is probably 1,500–2,000 feet thick, but because it is everywhere faulted, its apparent thickness is commonly 800–1,500 feet. A rough estimate of thickness in a few sections has been made by computing from the map. Between Majors Place and the mouth of Cooper Canyon, the apparent thickness is 1,000–1,360 feet; just north of Cooper Canyon it is 1,080 feet; north of the Kolcheck mine it is 1,200 feet; and just south of Cleve Creek, it is about 1,680 feet. Even this last figure, however, is probably somewhat less than the actual total thickness.

Two collections of fossils were obtained from the Pole Canyon Limestone. A. R. Palmer (written commun., 1962) stated that one contained only a single cranidium of *Alokistocare?* and the other (USGS colln. 3404–CO) contained:

Ptarmiganoides cf. *P. poulseni* Resser
Kochaspis dispar Resser

Alokistocare sp.
Pagetia sp.
Albertella sp.

Palmer further mentioned that the trilobites in 3404–CO are all assignable to the *Albertella* zone.

Palmer also stated (written commun., 1962) that the "presence of an *Albertella* zone assemblage in the assemblage in the Pole Canyon Limestone provides the first direct evidence that the lower part of the Pole Canyon Limestone is correlative with the upper part of the Pioche Shale in its type area, and * * * the development of relatively clean carbonate sediments that characterize the lower half of the Middle Cambrian series in the eastern Great Basin began earlier in this area than it did in areas to the south and east. Further north, in the Duck Creek Range, rocks equivalent to the Pole Canyon Limestone were mapped as Eldorado Limestone by Young (1960b)."

Fossils of early Middle Cambrian age that were collected from near the top of the Pioche Shale help to bracket the age of the Pole Canyon Limestone as probably early and middle Cambrian; only the earliest part of early Middle Cambrian and possibly the latest part of middle Cambrian time are not represented by Pole Canyon rocks.

During Pole Canyon time, the Connors Pass area lay so far from the shore of the eastward-transgressing sea that little of the clastic material derived from the land was carried into the area. The fairly abrupt transition from the underlying quartzite and shale to the limestone suggests that the shoreline shifted eastward somewhat rapidly in early Middle Cambrian time. Generally stable conditions, resulting in only very minor fluctuations of the shoreline, prevailed during most of Middle Cambrian time.

LINCOLN PEAK FORMATION

The Lincoln Peak Formation, a sequence of shaly rocks commonly 1,600–1,800 feet thick, overlies the Pole Canyon Limestone and was defined by Drewes and Palmer (1957) in the nearby southern Snake Range. During the time when shaly rocks were being deposited in the vicinity of Connors Pass, more limy rocks were being deposited in other areas, this contrast in deposition producing a fairly marked facies change through western Utah and eastern Nevada. In addition, at least the upper part of the formation is time transgressive to a small, but measurable, extent; it is younger in the Cleve Creek area than in the type locality.

The formation is exposed over more than 10 square miles of the quadrangle, largely along the east flank of the Schell Creek Range, but also on the west flank of Cave Mountain, and in the northwest corner of the quadrangle. The Lincoln Peak commonly forms gentle slopes having few outcrops, though nearly halfway up most of these slopes a resistant bed crops out as a small cliff or a row of knobs. Along U.S. Highway 6–50–93 near Majors Place, the slopes underlain by the formation are moderately steep and are broken by scattered ledges and rocky gullies; but less than a mile north of the highway, they have more of their usual smoothness.

In the northern part of the quadrangle, the Lincoln Peak Formation is divisible into two shale members and a middle limestone member. The limestone member thins abruptly from Grasshopper Canyon eastward to Cleve Creek, and between Cleve Creek and Cooper Canyon it is present only here and there. On the geologic map (pl. 1) the thin edge of the limestone is shown schematically as a marker bed. The upper shale member can generally be distinguished from the lower one by its lithologic character, even where it is not underlain by the middle limestone member, but south of Cooper Canyon, where the formation is slightly metamorphosed, these lithologic distinctions, as well as diagnostic fossils, are obliterated.

The Lincoln Peak Formation is conformable with the Pole Canyon Limestone, and in some places the typical rocks of the two formations are separated by a transitional zone a few score feet thick. The lowest shale beds assigned to the Lincoln Peak contain a few beds of limestone closely resembling that of the Pole Canyon. However, at most outcrops the basal contact relations are obscured by faulting. At such outcrops the beds on one or both sides of the contact are gradually truncated; the shearing, brecciation, minor folds, and crinkles commonly increase downward in the lowest 50 feet of the Lincoln Peak Formation; the contact is probably a minor thrust fault nearly parallel to the bedding. A fault is inferred to extend along the contact wherever beds are missing, for, although stratigraphy itself could account for their absence, no evidence of an unconformity has been found in any place where there is not also some evidence of faulting.

The lower shale member of the Lincoln Peak Formation consists of fissile to thin platy shale, and scattered limestone beds 1–2 inches thick which make up less than 10 percent of its volume. Much of the shale is silty and limy, and some of it is micaceous. On fresh fractures it is medium gray to pale olive gray, but weathered chips are pale yellow brown, pale purple, grayish orange pink, or grayish orange. The upper shale member is rarely as colorful. Along many miles of its base the lowest few tens of feet of the lower shale member weathers to a pale-red or moderate-red rubble, which is probably a fault breccia inasmuch as it appears only where there are other signs of structural disturbance; whatever its origin, it is a useful marker for the base of the formation.

The middle, limestone member of the Lincoln Peak Formation consists of medium-dark-gray thin platy limestone that contains some silty partings, *Girvanella*, fucoidal markings, and toward the top, agnostid tribolites. In upper Cleve Creek the limestone is bioclastic and oolitic. The calcareous beds are commonly ½–4 inches thick, but the shaly interbeds are less than one-fourth of an inch thick. Just south of Cleve Creek, some of the clastic interbeds are highly siliceous and resemble chert. Just north of Cooper Canyon, the limestone member is thin, lenticular, and in places deformed by minor disharmonic folds and thrust faults. It commonly has agnostid tribolites in its uppermost part, but along Cooper Canyon and south of it, the limestone member contains no fossils.

The upper shale member of the Lincoln Peak Formation is more fossiliferous and more varied than the lower shale member. The proportion of limestone beds gradually increases upward from about 30 to 75 percent; most of the beds are 1–2 inches thick, but toward the top some are as much as 9 inches thick. Low in the member, most of the limestone is finely crystalline but some is bioclastic or even coquinoid. Small limy plates having moderate-brown to grayish-brown silty surfaces are common in the lower part of this member. The silty shale generally weathers to larger less colorful fragments than the lower shale member. Nodular beds, crinkly bedding surfaces, beds of trilobite hash, and other bioclastic materials become common as the amount of limestone increases upward in the member. A section of the upper few hundred feet of this member was measured between altitudes of about 9,000 and 9,940 feet on the ridge south of Cave Mountain (fig. 3).

From Cooper Canyon south to Majors Place, the Lincoln Peak Formation is slightly metamorphosed to phyllitic or slaty shale. The intensity of the alteration increases gradually southward, and with one exception, just south of the mouth of Cooper Canyon, it does not change appreciably toward the contacts of the formation and the minor thrust faults. The first change in the rocks appears east of Bastian Spring, where rocks of the lower shale member have a pencil fracture and are sufficiently hard to tinkle under foot. A little north of Cooper Canyon the increasing hardness affects the entire formation. Just south of the canyon some beds about 50 feet above the base of the

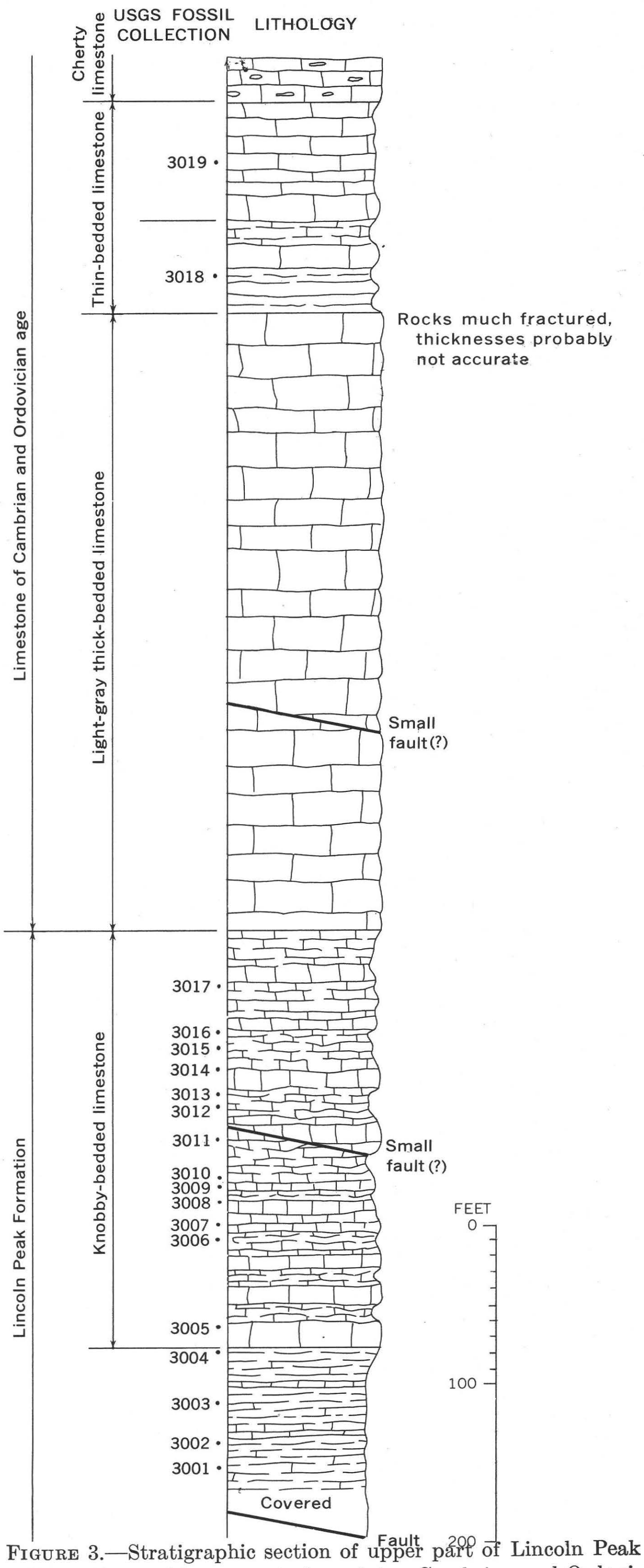


FIGURE 3.—Stratigraphic section of upper part of Lincoln Peak Formation and lower part of overlying Cambrian and Ordovician limestone, measured by A. R. Palmer and Harald Drewes on ridge south of Cave Mountain.

formation are phyllitic, have a weak slaty cleavage inclined about 30° to the bedding, and have bedding planes sprinkled with needles of a secondary mineral, probably actinolite. Quartz rubble from veins appears in areas of slightly metamorphosed rocks but is almost completely absent elsewhere. Near U.S. Highway 6–50–93, slaty cleavage is widespread and cleavage surfaces are spangled with fine-grained mica.

The clastic grains range from 0.01 to 0.25 mm in diameter. Quartz and calcite make up most of the rock, but their proportions are greatly varied. The very slightly metamorphosed rock contains considerable sericite and some spindle- or disk-shaped larger crystals of calcite among smaller equidimensional grains. More distinctly metamorphosed rock contains about 5–50 percent of sericite or muscovite and a small amount of chlorite among the clastic grains of quartz, calcite, black opaque material, and limonite pseudomorphs after pyrite. Even in the more distinctly metamorphosed rock, however, the metamorphism clearly was of very low grade.

Because of inadequate exposures and structural complications it is impossible to measure the thickness of the formation, and estimates (about 1,600 ft and not more than 1,800 ft) obtained from structure sections probably are within 10 percent of the actual thickness.

The Lincoln Peak Formation is the most fossiliferous Cambrian formation in the Connors Pass quadrangle. According to A. R. Palmer (written commun., 1962), it includes beds ranging in age from about middle Middle Cambrian to Franconian, in the Late Cambrian. The Middle Cambrian part of the formation is less fossiliferous than the Upper Cambrian part. The trilobites are predominantly agnostids. Near the top of the Middle Cambrian part of the formation in the northern part of the quadrangle, the limestone member yielded agnostids referable to *Lejopyge*. Limestones containing *Lejopyge* have been found at a comparable stratigraphic position at the top of the Marjum Limestone in the House Range 60 miles to the east, in the Lincoln Peak Formation on the east side of the Snake Range, and in the middle of the Patterson Pass Shale (Kellogg, 1960) near Patterson Pass in the southern Schell Creek Range. A few small collections from beds below the *Lejopyge*-bearing limestone contain species of *Ptychagnostus*, *Hypagnostus*, *Peronopsis*, and *Doryagnostus*, indicative of ages within the upper half of the Middle Cambrian.

Palmer (written commun., 1962) commented:

> The Upper Cambrian part of the Lincoln Peak Formation is generally fossiliferous, and faunas of at least six different Late Cambrian ages have been collected from it. These faunas are largely undescribed, although studies of the trilobites are now underway.
>
> Fossils from the upper part of the formation indicate that the contact with overlying Upper Cambrian and Lower Ordovician limestone varies in age from perhaps as young as the *Conaspis* zone, of Franconia age, to as old as the *Dunderbergia* zone, of Dresbach age. In general, the contact becomes younger to the north and northwest within the quadrangle.
>
> In the area south and east of the peak with triangulation marker Bastian, the top of the Lincoln Peak Formation is within the *Dunderbergia* zone and is of approximately the same age as it is in its type area in the Snake Range.
>
> In the vicinity of Cleve Creek and in the northwestern corner of the quadrangle, the top of the Lincoln Peak Formation is distinctly younger and the upper beds contain trilobites of the lower part of the *Elvinia* zone, of Franconia age. The top of the formation here is approximately equal to the top of the Dunderberg Formation in the vicinity of McGill, about 14 miles to the north. Apparently this quadrangle is in an area of striking facies changes within the early Late Cambrian. The massive nearly lithographic light-gray limestone of the upper part of the Raiff Formation of Young (1960b), which underlies the Dunderberg Formation at McGill, has changed to thin-bedded limestone in the vicinity of Mosier Canyon, and near Cleve Creek and in areas south to Bastian Spring its stratigraphic position seems to be completely occupied by thin-bedded dark silty limestone. The silty limestones of the upper part of the Lincoln Peak Formation in Mosier Canyon and near Cleve Creek also change facies southward and are temporally equivalent to the massive limestones of the lower part of the undifferentiated limestones of Cambrian and Ordovician age in the vicinity of Cave Mountain.
>
> In mapping, the lithographic identity of the map units has been retained at the expense of some inconsistency in regard to age.

Palmer (written commun., 1962) further said that his regional study of these facies changes indicated that the local variations "* * * result from a complex lateral shifting of the contact between a belt of generally silty sediments, represented by the Dunderberg Shale and Lincoln Peak Formation, and a belt of generally clean carbonate sediments lying to the north and east, in response to a significant marine regression during late Dresbach time."

The change from a shaly facies in eastern Nevada to a limestone facies in western Utah, during early and middle Late Cambrian time, indicates a western source for the Lincoln Peak Formation. But late in this period a tongue of clastic sediments derived from a landmass in eastern Utah was spread out to form a part of the upper shale member of the Lincoln Peak Formation in the Cleve Creek area (Palmer, 1960).

METAMORPHISM OF PRECAMBRIAN AND CAMBRIAN ROCKS

Rocks as young as the Late Cambrian, or perhaps even slightly younger, are sufficiently metamorphosed to be either moderately recrystallized or moderately sheared and contorted and to contain a small amount of newly constituted minerals. But not all the rocks older than the Middle Cambrian are metamorphosed; the type and intensity of the changes differ with the composition of the rock, as well as with general geographic distribution, and they are affected to some extent by the proximity of major low-angle faults.

In outcrops the metaquartzite does not seem to differ significantly from other quartzite, but thin sections reveal some textural differences. In the Precambrian and Cambrian quartzites, the quartz grains generally interlock or even interpenetrate; these rocks contain comparatively few remnants of rounded quartz grains having overgrowths of secondary silica, and there are numerous flakes of muscovite or sericite within the quartz grains as well as between them. In the younger quartzite, however, some of the grains interlock but do not interpenetrate; there are also many remnants of rounded quartz grains, and the grains enclose little or no sericite.

Phyllitic or slaty rock occurs in all the Precambrian rocks; there is some in all the Cambrian Pioche Shale, and it is abundant in some of the Cambrian Lincoln Peak Formation. The Precambrian rocks include slaty shale, in which the cleavage is inclined as much as 30° to the bedding, and phyllitic metaquartzite. Although not widely exposed within the quadrangle, the Precambrian metamorphic rocks appear outside the quadrangle in the northern part of the Schell Creek Range (Young, 1960b), where no low-angle faults have been recognized.

The Pioche Shale also contains much metaquartzitic shale, a little phyllitic shale, and some highly micaceous shale; these different types of shale are perhaps examples of Dapples' (1962) phyllomorphic stage of diagenesis, which overlaps the zeolite and chlorite grades of metamorphism. The Lincoln Peak Formation consists of shale to the north and west of Bastian Creek; to the south it changes gradually along the strike but generally does not change across the strike. Along Bastian Creek the shale is so indurated as to tinkle underfoot, and it has a well-formed pencil fracture. South of Cleve Creek all fossils and many sedimentary features are obliterated; here the rock is very hard, has a weak slaty cleavage, and contains many small quartz veins. Near Majors Place these rocks are plicated, have a moderately well formed slaty cleavage and a dull phyllitic luster, and contain some chlorite.

The distribution of the weakly metamorphosed rocks in the Lincoln Peak Formation is largely independent of any main low-angle fault. Near Majors Place the formation lies close beneath the largest low-angle fault in the quadrangle, but along Bastian Creek it is several miles from the trace of the fault, and presumably far beneath a projection of the fault. At the head of Cave Creek and northwest of Grasshopper Canyon, however, where the formation is exposed close to the fault, it is unmetamorphosed, even though the thin sheet of limestone overlying it is sheared and slightly recrystallized as it is everywhere along the fault.

The Pole Canyon Limestone consists chiefly of saccharoidal limestone or low-grade marble, in which the calcite grains are completely and coarsely recrystallized, all sedimentary details are obliterated, and muscovite or sericite, where present, is intergrown with the calcite grains as well as molded around them. Mortar texture or granoblastic texture and shear planes are present on a microscopic scale even at some distance from the major fault. The wide extent of this metamorphism of one formation might suggest that the metamorphism was due to stresses along an unrecognized fault within that formation, but this explanation is not favored, because the Pole Canyon Limestone and the adjacent formations are far less deformed into tectonic lenses than are the formations along the largest recognized thrust fault and because the Pole Canyon contains no relict drag features.

The origin of the metamorphism is difficult to explain, because inferences drawn from field relations require a broader regional base than is now available, and those drawn from microscopic relations require a more systematic study of the physical properties of the rocks than it has been possible to make. At present I believe that some of the metamorphic features are the result of a mild dynamic deformation, caused by movement along the nearby major low-angle faults. Similar views have been expressed and more adequately supported by Misch and Hazzard (1962). However, the occurrence of metamorphic features far from the major low-angle fault and the variations in the intensity and the extent of metamorphism along that fault probably indicate that much of the metamorphism was due to causes independent of this dynamic deformation. Contact metamorphism above a hypothetical buried stock might explain the association of quartz veins with the more intensely altered rock, but inasmuch as none of the exposed stocks of the region has a wide aureole, such a hypothesis fails to explain the widespread mild alteration. This wide alteration most likely was a static metamorphism produced by the tremendous load of overlying sediments, about 6.5 miles thick during Late Jurassic or Early Cretaceous time. This static metamorphism may have been reinforced by some poorly understood mechanism involving diffuse and irregular movements of pore fluids under high hydrostatic pressure, a condition possibly critical in the structural development of the area.

The static metamorphism presumably was in progress during middle Mesozoic time, when the rocks that were metamorphosed were most deeply buried (see p. 77). This process may have begun at an earlier time, when a threshold of depth of burial, of unknown amount, was reached, and it is even more likely that the process continued until another threshold of depth of burial, of unknown amount, was passed as the rocks were exhumed. Throughout the region, the time and rate at which the rocks were exhumed is poorly known, but very likely exhumation began with the Late Jurassic or Cretaceous epeirogenic uplift that is indicated by the withdrawal of the sea. The rate of exhumation may have been greatest during the middle Tertiary time of block faulting; thus it is unlikely that metamorphism continued any later. Armstrong (1963) presented a dozen radiogenic age determinations that placed the end of the metamorphism as late as Miocene.

At least some shearing, a common feature of the dynamically metamorphosed rocks close to the Schell Creek Range thrust fault, has been superposed on some of the static metamorphism.

CAMBRIAN AND ORDOVICIAN SYSTEMS

LIMESTONE

A sequence of limestone, about 3,000 feet thick, overlies the Lincoln Peak Formation. Hague (1883, p. 260) named this sequence the Pogonip Limestone, after Pogonip Ridge near Hamilton, Nev., and believed it to be Lower Ordovician and possibly Upper Cambrian. Since then these rocks have been extensively studied in Nevada and Utah, and sections have been described at Ibex, Utah, by Hintze (1951a, b) and near Eureka, Nev., by Nolan, Merriam, and Williams (1956). As a result, the stratigraphy of the Cambrian and Ordovician rock has been revised in many places. The limestone that is mainly Ordovician age has been raised to the rank of a group, and the Upper Cambrian limestone has been made a separate formation.

In the Connors Pass quadrangle, about midway between the areas studied by Hintze (1951a, b) and by Nolan, Merriam, and Williams (1956), lithologic variations are more subtle than at Eureka, and fossils are less numerous than at Ibex. Consequently, the Pogonip Limestone of Hague in the Connors Pass quadrangle is difficult to divide and to correlate with the Eureka and Ibex sections; it is also impractical to separate the limestone of Ordovician age from that of Cambrian age. As structural complexities add to the difficulties, not only in this quadrangle but throughout much of eastern Nevada and perhaps elsewhere, the usefulness of the Eureka and Ibex sections may be restricted to the vicinity of those sections. These rocks are called simply Cambrian and Ordovician limestone in this report, although they are equivalent to the eminently practical unit, the Pogonip Limestone of Hague. The formation is divided informally into a lower limestone, a shale member, and an upper limestone; where the shale member is not mapped, the limestone members may not always have been differentiated.

The stratigraphic thickness of the Cambrian and Ordovician limestone in this area is estimated to be about 3,000 feet; however, in no one place is there more than about 2,000 feet of strata between the bounding contacts because, although the formation as a whole generally is a nearly flat-lying sheet, its beds are generally inclined at moderate angles and are broken by high-angle faults. Inasmuch as many of these faults remain unmapped because of poor stratigraphic control, estimates of the overall thickness from structural sections are unreliable, and the thicknesses assigned to the constituent units are based on very rough estimates.

The undifferentiated Cambrian and Ordovician limestone underlies a broad area near the center of the quadrangle and smaller areas in all parts of the quadrangle except the southwest corner. The rocks in the lower few hundred feet form small ridges and spurs. The middle part is exposed in high, steep light-gray cliffs. The upper part generally forms low cliffs that alternate with discontinuous benches.

The lower limestone is divisible into three unmapped units: a basal ledgy limestone, a cliff-forming limestone, and a silty limestone. The basal contact of the Cambrian and Ordovician limestone is vertically gradational and is laterally time-transgressive. In most places the base is placed at the base of the lowest small cliff of moderately thick bedded limestone, containing thin silty partings, that overlies the shalier rocks of the Lincoln Peak Formation. In the Cooper Canyon area, however, and southward toward Majors Place, where most of this limestone is cut out by an overlying fault that is almost parallel to the bedding planes and where the underlying Lincoln Peak Formation is slightly metamorphosed, the lowest cliff-forming unit is older and seems to be several hundred feet lower stratigraphically than it is to the north. In some places the lateral changes are entirely stratigraphic and involve minor lithologic differences—in the proportions of silt to carbonate for example—elsewhere they are partly or wholly structural, or are metamorphic. Some apparent differences are only differences in weathering habit. Minor bedding-plane faults have occurred at or near the basal contact in many places, probably because this contact separates incompetent rocks above from competent ones below. In some places, such as the Kolcheck Basin, where the contact is not broadly gradational, the stratigraphic section may be slightly telescoped along a minor bedding-plane fault. In other places, such as northwest of Bastian Spring, weak thin-bedded rock has been brecciated and then thoroughly healed, either with or without the addition of calcite cement. The cemented breccia is relatively resistant to erosion, owing to its alteration, and forms cliffs.

Because of these complications in structure, it is well to apply other criteria for determination of the contact; one useful criterion is the distribution of chert. The Cambrian and Ordovician limestone is generally cherty, whereas the Lincoln Peak Formation is not, except east of Cottonwood Spring where chertlike pods are in the centers of small siliceous siltstone lenses, in a group of ledge-forming limestone beds many hundreds of feet below the top of the Lincoln Peak. However, in most places the lowest chert is in the basal ledgy unit of the lower limestone, or roughly within the lowest 300 feet of the Cambrian and Ordovician limestone and is possibly equivalent to the Windfall Formation; in a few places the lowest chert does not appear below the base of the cliff-forming limestone that lies above the basal ledgy unit.

The basal ledgy limestone unit is commonly 200–300 feet thick, but in some places it is as much as 800 feet thick. It is light-gray to medium dark-gray thin-bedded shaly limestone to a platy limestone containing pale-yellowish-brown, pinkish-gray, and pale-reddish-brown silty partings not constituting more than 15 percent of the rock. Most of the beds are only a few inches thick, but some are as much as 4 feet thick. Crinkly and knobby bedding characterize much of the unit, and in many places nodular beds appear near the top. The texture of the limestone ranges from very fine to coarsely crystalline or bioclastic. Highly fossiliferous beds are common, and some beds contain abundant small spherical bodies that resemble *Girvanella*. Chert forms small dark-gray irregular nodules and pods and a few long lenses. A few irregular nodules of chert cut across the bedding. Generally where the unit is relatively thick a little fissile olive-gray shale is interbedded with nodular limestone near the top. South of Cave Mountain, where the basal ledgy limestone unit (fig. 3) is thickest, it is divisible into four parts, in ascending order: a dark-gray knobby-bedded limestone, a light-gray thick-bedded limestone forming small cliffs, a thin-bedded limestone, and a cherty limestone. South of Cooper Canyon these rocks are slightly recrystallized, and much sedimentary detail and many fossils are obliterated. Here, too, each of the four parts of the sequence becomes thinner, perhaps because it has been partly cut out by faults or thinned by pressure.

The cliff-forming limestone unit is about 1,000–1,100 feet thick, contains relatively little silt and clay, and is thick bedded. It is medium gray and fine to medium-coarse grained. The rock does not part readily on bedding planes; therefore it breaks down into more massive blocks than it would if the bedding were more generally marked by silty layers. Chert nodules are generally present in small amounts, and high in the unit there are several zones, each a few tens of feet thick, containing as much as 20 percent thin-bedded lenticular chert.

The silty limestone unit, roughly estimated to be 1,100–2,000 feet thick, caps the flat-topped ridges at triangulation marker Bastian and between upper Steptoe Creek and Cleve Creek. It consists of thin layers of cliff-forming or ledgy limestone that alternate with weaker, more largely detrital, rocks. This limestone resembles that in the lower units; it is a medium-gray fine-grained to coarsely crystalline bioclastic moderately thick bedded rock having silty partings and some chert nodules. Intraformational conglomerate, which is common, in places contains conspicuously flat limestone pebbles several inches in diameter. The detrital rock also includes much limy siltstone and shale and some coquina. Some of the shale is very fissile and weathers olive gray; other shale and siltstone weather reddish gray. The more detrital clastic rock beds seem to form lenses a few tens to a few hundred feet thick, but inasmuch as bedding-plane thrust faults are common here, the lenses may have been formed by the shearing out of extensive layers.

SHALE MEMBER

The silty limestone is overlain by a shale member about 400 feet thick. An accessible and thick section of the shale member is exposed on the north slope of a large unnamed canyon 2 miles east-northeast of Cave Creek Reservior. This member consists chiefly of a dark-olive-gray highly fissile clayey shale, but it also contains thin beds of siltstone, muddy sandstone, flat-pebble limestone conglomerate, and bioclastic and coquinoid limestone. Small bodies of these rocks may be difficult to distinguish from the clastic material interbedded with the underlying silty limestone unit, but they generally contain a larger proportion of fissile shale and a distinctive fauna consisting of ostracodes, gastropods, and sponges.

A limestone 300–400 feet thick is the highest unit of the Cambrian and Ordovician limestone. It is moderately well exposed above the shale member in the section northeast of the Cave Creek Reservoir, but in some places it is not present above the shale, probably because it is faulted out. The limestone is medium gray on fresh fractures but weathers light gray, is finely crystalline to coarsely crystalline and in part bioclastic, contains a little intraformational conglomerate, and near the top includes some siltstone and sandstone. The limestone beds are 2–10 inches thick, and their bedding surfaces are slightly knobby. Chert occurs in small amounts as nodules and rarely as tubular bodies that may have replaced fossils. The upper limestone thus resembles much of the lower limestone from which it can be distinguished only by its stratigraphic position and by its fauna, which includes abundant ostracodes, gastropods, and brachiopods.

FAUNA, AGE, AND CORRELATION

Much of the Cambrian and Ordovician limestone is moderately fossiliferous, but the basal ledgy limestone unit of the lower limestone and the shale member are highly fossiliferous, and the cliff-forming limestone unit of the lower limestone is only sparsely fossiliferous. Of the Ordovician fossils in table 2, the gastropods were identified by E. L. Yochelsen (written commun., 1960) and all others by R. J. Ross, Jr. (written commun., 1959, 1960); Cambrian fossils not in the table were identified by A. R. Palmer.

A. R. Palmer (written commun., 1962) had the following to say about the Cambrian fossils:

> Locally, in the vicinity of triangulation marker Bastian [Cave Mountain], a rubbly zone in the lower part of this sequence [the base of the thin-bedded limestone unit of fig. 3 of present paper] yields trilobites of the *Elvinia* zone and probably correlates with the Corset Spring Shale of the Snake Range. The sequence of massive carbonate rocks between this and the Lincoln Peak Formation thus correlates with the Johns Wash Limestone of the Snake Range. In the northern part of the quadrangle, a cherty unit correlative with the Catlin Member of the Windfall Formation characteristically yields trilobites probably representing the *Conaspis* zone, of Franconia age, which weather out on the surfaces of thin-bedded dark-gray fine-grained limestones. This unit is usually separated from the underlying beds of the Lincoln Peak Formation by noncherty limestones correlative with the Barton Canyon Limestone Member of Young (1960b), in the Windfall Formation near McGill. No fossiliferous younger Cambrian rocks were found within the quadrangle, but a collection of Trempealeau-age trilobites about 4 miles north of the quadrangle indicates their probable presence also within the area described in this report.

Ross also identified a trilobite from the cliffy limestone unit as *Eurekia* sp. and regarded it as probably correlative with forms found in the basal part of the Goodwin Limestone in the Eureka district.

The Ordovician fossils are listed in table 2.

The fossils thus indicate that the basal ledgy limestone unit and most or all of the cliff-forming limestone unit are of Late Cambrian age. The silty limestone unit, however, is mainly Early Ordovician but may include a few beds of Middle Ordovician near the top; the shale member and the upper limestone are Middle Ordovician.

The sedimentation pattern for the Cambrian and Ordovician limestone in the region resembles the pattern of the earlier Cambrian rocks in that fine clastics alternated with limestone. During the middle Early Ordovician, approximately represented by zones D–I of Ross (1951) and of Hintze (1951a, b, 1952), a sheet of silt and clay containing much calcite was deposited in eastern Nevada. During late Early Ordovician time, a similar sheet, the Ninemile Formaton, was deposited in central Nevada, but this too contained considerable calcite as far east as the Egan Range, as shown in the sections of Kellogg (1960) and Fritz (1960). Together these deposits are equivalent to the silty limestone unit of the lower limestone of the Cambrian and Ordovician limestone in the Connors Pass quadrangle. Inasmuch as clastic material was largely absent in western Utah while the Ninemile Formation was being deposited, the clastic material may have been derived from the west. Strong currents were moving over the sea floor during late Early Ordovician time, for intraformational conglomerate is common throughout the region. During part of early Middle Ordovician time, sediment rich in silt and clay was deposited to form the shale member, which is correlative with the Kanosh Shale of Hintze (1951a) about 25 miles to the east. West of the Egan Range this member becomes more limy, and at Eureka, about 65 miles to the west, silt and clay were not deposited at this horizon. The shale member was probably derived from the south or east, as the Corset Spring Shale, of Late Cambrian age, is inferred to have been.

ORDOVICIAN SYSTEM

EUREKA QUARTZITE

Eureka Quartzite, a nearly white quartzite 300–400 feet thick, overlies the Cambrian and Ordovician limestone. It was named by Hague (1883, p. 262) for its exposures near Eureka, Nev. Since then the formation has been studied by Kirk (1933), Hintze (1951a, b), Webb (1956), and Nolan, Merriam, and Williams

TABLE 2.—*Ordovician fauna from unnamed limestone*

[Zones are lettered in ascending order according to the system of Hintze (1951a, b; 1952). Identified by R. J. Ross, Jr. (written commun., 1959, 1960), and E. L. Yochelson (written commun., 1960)]

Unit	Silty limestone unit of the lower limestone								Shale member							Upper limestone			
Faunal zone	A?	B–A	G_2	J	J–H	—	J–H	L?	M–K	M?	M	M	M	N–M	M	N or O	N? or O	N?	N
Field No.	59D292	59D159	60D545	60D555	59D819	60D531	60D519	60D530	58D54	59D312	60D514	59D331	58D29	59D189	58D56	58D47	60D532	60D515	60D546
U. S. Geol. Survey loc.	D563-€O	D559-€O	D777-€O	D779-€O	D561-€O	D776-€O	D771-€O	D775-€O	D527-€O	D564-€O	D773-€O	D562-€O	D525-€O	D560-€O	D528-€O	D526-€O	D774-€O	D772-€O	D778-€O
Brachiopods:																			
Anomalorthis cf. *A. utahensis* Ulrich and Cooper														×					
Anomalorthis sp.											×	×	×		×?	×			×
Diparelasma sp.				×?															
Orthambonites michaelis Clark											×								
Orthambonites sp.								×				×	×	×	×		×?		×
Orthoid brachiopod																		×	
Gastropods:																			
Eotomaria sp.																	×		
Moderately high spired pleurotomariacean																		×	
Unidentified gastropod									×						×	×	×	×	
Small low-spired gastropod																	×		
Hormotoma? sp.										×									
Ostracodes:																			
Leperditia cf. *L. bivia*											×								×
Large leperditiids										×				×?		×	×?		
Macronotella? sp.											×								×
Nonleperdetiid ostracodes																	×	×?	
Small ostracodes												×	×	×?	×				
Trilobites:																			
Uromystrum pogonipensis Hintze											×	×							
Bathyurus sp.																×	×		
Bathyurid trilobite																		×	
Cybelopsis sp.											×?							×?	×
Eleutherocentrus petersoni Clark												×?							
Hystricurus cf. *M. millardensis* Hintze	×																		
Illaenus cf. *I. utahensis* Hintze																			×
Kawina cf. *K. sexapugia* Ross								×											
Kanoshia cf. *K. kanoshensis* (Hintze)												×							
Kirkella sp.				×?															
Kirkina? sp.																	×		
Hintzeia cf. *H. celsaora* (Ross)			×																
Paranileus? sp.					×														
Pseudomera cf. *P. insolita* Poulsen							×												
Pseudomera? sp.					×					×							×		
Pseudoolenoides acicaudus Hintze																			×
Pseudoolenoides dilectus Hintze											×		×						
Simphysurina cleora (Walcott)		×																	
Simphysurina cf. *S. cleora* (Walcott)	×																		
Trigonocerca sp.					×														
Huge pliomerid trilobite						×													
Others:																			
Unidentified bryozoan										×			×			×	×		
Eofletcheris sp.									×										
Receptaculitea? sp.							×								×				

(1956). It differs from the Prospect Mountain Quartzite in being much thinner and of lighter color. Its upper contact is easily recognized where exposed, because it underlies the darkest dolomite in the section.

The Eureka Quartzite forms lenses commonly less than 1,000 feet long, most of which are high on the flanks of the ridge between Cave Creek and Cleve Creek. The most accessible exposure of a thick lens of the quartzite is about 2 miles east-northeast of Cave Creek Reservoir, along the bottom of the large canyon that is tributary to Cave Creek from the north. The upper contact is well exposed there, but the basal contact is covered, as it is elsewhere, with quartzite debris.

The Eureka Quartzite commonly forms small cliffs that disintegrate into large rounded boulders (fig. 4). At some places on the east flank of the Schell Creek Range, the quartzite weathers to small blocky debris and forms slopes as gentle as those of adjacent formations. Some of the Eureka in small fault blocks resembles the intensely silicified but nonclastic rock that is common along some faults. Both types of rock are believed to occur along the north slopes of Cooper Canyon, where the Eureka Quartzite is white or pinkish gray and the silicified rock, perhaps originally a

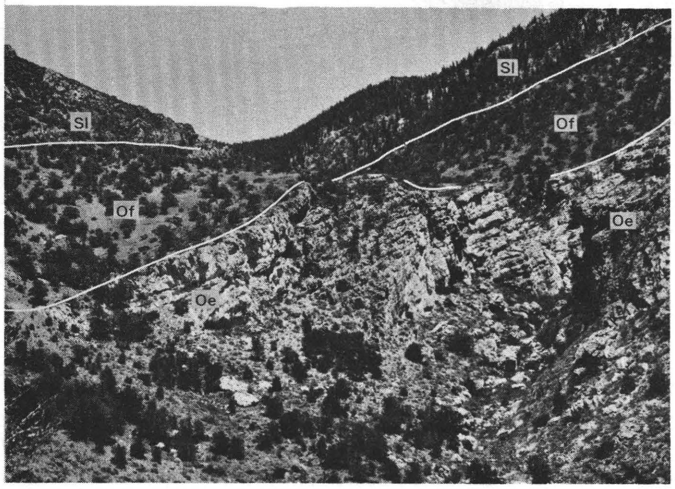


FIGURE 4.—Typical outcrop of white Eureka Quartzite (Oe), a mile east of Cave Creek Reservoir. Note the large subrounded weathered blocks in the canyon bottom. The dark slope above the quartzite is underlain by Fish Haven Dolomite (Of), and the light-gray outcrops farther uphill and above the dark slope to the left are Laketown Dolomite (Sl).

limestone, is brownish gray. The light-colored blocks of siliceous rock are associated here with other fault blocks of formations adjacent to the Eureka Quartzite, another indication that the light-colored blocks are the Eureka Quartzite.

The quartzite is a very light gray to pinkish-gray rock in which most beds are 1–10 feet thick, though some beds are thicker. Some beds are finely laminated and others are crossbedded. Much of the quartzite has been shattered or brecciated and then cemented with quartz, this process making an extremely tenacious rock. The quartzite locally contains many small (¼- to 1½-inch diameter) subspherical pockets filled with pale-yellowish-brown sandy or dolomitic material, and a very small amount of the rock is sandstone rather than quartzite.

The quartzite consists almost entirely of unstrained quartz grains that are mostly 0.05–0.7 mm in diameter but rarely are as much as 5 mm. Some of the quartzite contains a very small amount of sericite and black opaque material between the quartz grains. The quartz grains themselves also enclose very small amounts of apatite, biotite, zircon, iron oxide, and minerals tentatively identified as tourmaline and rutile. The grains are subrounded to subangular and are alined, probably parallel to the bedding. Some grains are only moderately closely packed and are cemented by later optically continuous quartz, but the boundaries of more closely packed grains interpenetrate or interlock. The sandy and vuggy varieties of the rock contain some carbonate cement, probably dolomite.

The thickness of the Eureka Quartzite cannot be measured in the Connors Pass quadrangle, and it is difficult to estimate because the formation is persistently faulted. Even where faults are not obvious, the range of thickness of the formation is great, this diversity implying either faulting or unconformity or both. Faulting would seem to be the main cause of the varied thickness, for elsewhere in the region where an unconformity does overlie the quartzite, the Eureka is not as lenticular as it is in Connors Pass quadrangle; nevertheless, several apparently unfaulted quartzite lenses within the quadrangle, including the one east of Cave Creek Reservoir, are only 100–200 feet thick. About 1½ miles northeast of Aspen Spring, just north of the quadrangle, the quartzite is 300–400 feet thick; well-exposed sections nearby indicate that this may be its complete thickness.

No fossils have been found in the Eureka Quartzite. The formation is underlain, however, by rocks that contain a lower Middle Ordovician fauna and is overlain by rocks that contain an Upper Ordovician fauna. The quartzite is therefore assumed to be of Middle Ordovician age, although it may include some rocks of early Late Ordovician age.

The Eureka Quartzite occurs throughout the Schell Creek Range and adjacent ranges and apparently varies little in appearance and only slightly in thickness. Ross (1964, p. 1551) suggested that the Eureka Quartzite may have been derived from the Uinta Arch to the east or, less likely, from the north.

FISH HAVEN DOLOMITE

The Fish Haven Dolomite, a very dark brown dolomite 400–500 feet thick, overlies the Eureka Quartzite with apparent conformity. This rock is the darkest of the Paleozoic sequence, and, because it overlies the nearly white Eureka Quartzite, it is easy to identify even at a distance. I apply a nomenclature from sections in faraway northeastern Utah, rather than from nearby Eureka, Nev., to the rocks of Late Ordovician and Silurian age because lithologies and thicknesses change more markedly toward Eureka than they do to the northeast. The formation was named the Fish Haven Dolomite by Richardson (1913) for its exposures near the town of Fish Haven in northeasternmost Utah.

The Fish Haven Dolomite is found only in the north-central part of the Connors Pass quadrangle, where it is exposed along a discontinuous narrow belt high on both flanks of the ridge between Cleve Creek and Cave Creek and forms scattered small bodies along a fault zone low on the west flank of that ridge from near the center of the quadrangle to the north edge of the quadrangle.

Few sections of this formation are both well exposed and apparently complete. On much of the east flank of the ridge just mentioned the dolomite forms a dip slope that, though steep, contains few outcrops. Near the Kolcheck mine the dolomite lies successively upon the Eureka Quartzite, upon the upper limestone, and upon the shale member of the Cambrian and Ordovician limestone. The contacts of the dolomite mass are there nearly horizontal, but the bedding within it dips steeply. This structural discordance indicates that the basal contact of the dolomite is a low-angle fault rather than an unconformity, and because this relation prevails over an extensive area, the dolomite is probably much broken by unrecognized faults. Low on the west flank of the ridge, however, and in the canyon that joins Cave Creek from the north, about three-quarters of a mile east of Cave Creek Reservoir, the Fish Haven Dolomite is moderately well exposed and is not obviously faulted.

Where the Fish Haven Dolomite overlies the Eureka Quartzite, its lower part commonly forms a bench and gentle slopes. The upper part of the Fish Haven forms steeper slopes that are broken in places by low cliffs, and it invariably weathers to small blocky rubble, which thinly covers much of the gentle slopes formed by the lower part of the dolomite.

Most of the Fish Haven Dolomite is very dark brownish gray to dark gray, but toward the top the dark beds alternate with light-gray dolomite, which increases in abundance upward. The thickness of beds in the formation averages about 1–2 feet but is as much as 5 feet in some places. Some beds contain fine laminae that are commonly much fainter than those in the younger Simonson Dolomite. The faint laminae near the top of the thick dolomite bed and left of the tree in figure 5 may be compared with the abundant conspicuous laminae shown in figure 8. Minute fractures are abundant and are at least as closely spaced as bedding planes. The dolomite is coarsely crystalline and has few primary sedimentary features. Chert makes up less than 5 percent of the rock; it forms scattered light-gray to dark-gray roughly ellipsoidal nodules a few inches long. Quartz grains like those in the underlying quartzite occur only near the bottom of the formation. The dark dolomite has a strong fetid odor when broken.

The Fish Haven is 300–400 feet thick in the canyon east of Cave Creek Reservoir, where it does not seem to be faulted. In several other places it seems to be 500–700 feet thick, but inasmuch as these places are near severely faulted rocks little reliance can be placed on the larger estimates.

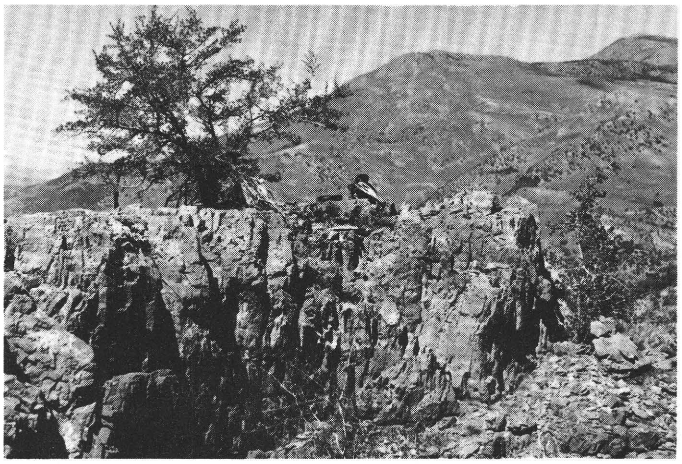


FIGURE 5.—Typical bed of Fish Haven Dolomite east of upper Steptoe Creek, 1.5 miles south of the north border of the quadrangle.

The Fish Haven Dolomite contains a few poorly preserved fragments of corals, gastropods, and brachiopods that are replaced by coarsely crystalline light-colored dolomite or by quartz. The chain coral *Catenipora* sp. was identified in two collections by W. A. Oliver, Jr. (written commun., 1960), who dated them as probable Late Ordovician in age although a similar coral appears in the Silurian. This chain coral is probably the *Halysites* sp. of many other reports on the formation. J. T. Dutro, Jr. (written commun., 1961), also identified a single brachial valve of a *Plaesiomys* that is probably related to *Plaesiomys subquadrata* (Hall), a widespread species of Late Ordovician age. In other areas the formation is more adequately dated as Late Ordovician, and specifically as Richmond.

SILURIAN SYSTEM

LAKETOWN DOLOMITE

The Laketown Dolomite, a light-gray coarse-grained dolomite 600–700 feet thick, conformably overlies the Fish Haven Dolomite. It was named by Richardson (1913) for exposures near Laketown in northeastern Utah.

The Laketown Dolomite underlies small areas in the north half of the quadrangle; some of these areas are on or near the crest of the range, others are scattered along the fault zone at the foot of the west flank of the range between Cave Creek and the north edge of the quadrangle, and several are in Cooper Canyon. The formation is relatively accessible and is well exposed high above the dry falls on the Eureka Quartzite in the large canyon that joins Cave Creek from the north, about three-quarters of a mile east of Cave Creek Reservoir, and also along a gully on the steep northwest-facing slope just west of the Kolcheck mine.

The formation has few outstanding characteristics that are noticeable at a distance. Its base forms the upper parts of the small cliffs that begin near the top of the Fish Haven Dolomite. On steep slopes, most of the formation is exposed in discontinuous cliffs where its bedding is only faintly visible.

Where the Laketown Dolomite is best exposed, its base intergrades with the underlying Fish Haven Dolomite, in which the spacing and thickness of the beds of dark dolomite gradually diminish upward toward the contact. The base of the Laketown Dolomite is arbitrarily placed at the top of a brown bed above which dark colors are subordinate to light colors, but this horizon does not anywhere form a reliable time-stratigraphic boundary.

The Laketown Dolomite is largely light gray to yellowish gray or very pale orange on both fresh and weathered surfaces, but beds of light-brownish-gray to brownish-gray dolomite, much like those in the underlying formation, continue to appear at wide intervals for several hundred feet above the contact. Most of the rock is coarsely crystalline, but in the upper part of the formation there are a few beds of light-gray fine-grained dolomite resembling the overlying Sevy Dolomite and a few beds of medium-gray to brownish-gray coarsely laminated fine-grained dolomite. Although most of the dolomite contains only small scattered nodules of yellowish-gray chert, chert and other siliceous material make up more than half of one layer about 30 feet thick.

The original textures and structures of the Laketown Dolomite have been greatly modified by diagenesis. The few faintly laminated beds contain indications of fine bedding and some crossbedding; elsewhere the bedding is generally indistinct and the beds are at least moderately thick. The competence of the rock is therefore mainly controlled by the spacing of small fractures. Some laminated beds contain concentrically layered ellipsoidal bodies having cores of silica, which may be remnants of algal structure. Fragments of corals and brachiopods are scarce and, where present, are replaced by coarse-grained dolomite. Vugs commonly less than 2 inches across are scattered through much of the moderately coarse grained dolomite; many of them are lined with coarse crystals of dolomite and a few contain quartz.

The thickness of the Laketown Dolomite is about 600–700 feet, as estimated from the section east of the Cave Creek Reservoir. Laketown of similar thickness underlies about a mile of the crest of the high ridge between Cave Creek and Cleve Creek just north of the place where the road reaches the crest of the ridge north of Cave Mountain. Here, however, the internal structure of the Laketown is probably complex, for the beds commonly dip 10–40° more steeply than the basal contact.

Two collections of poorly preserved fossils were made from the upper part of the formation along the crest of the ridge north of Cave Mountain. W. A. Oliver, Jr., and Helen Duncan (written commun., 1960) reported that the collections contain *Favosites* sp., *Catenipora*? sp., horn corals, and phacelloid rugose corals, which they regarded as probable Silurian.

During Silurian time the Connors Pass quadrangle presumably was on a moderately stable shelf between an actively subsiding geosynclinal area to the west and a more stable belt to the east. Intense diagenetic dolomitization has all but obliterated the initial character of the Silurian sediments. The fact that many of the fossil remnants in the sediments are coralline indicates at least a moderately warm sea; the absence of terrigenous detritus suggests that this sea was bordered by lowland areas.

DEVONIAN SYSTEM

SEVY DOLOMITE

The Sevy Dolomite, a fine-grained light-gray dolomite at least 900 feet thick, overlies the Laketown Dolomite. Nolan (1935) named it for the excellent exposures in Sevy Canyon in the northern Deep Creek Range, Utah, and since then the formation has been most thoroughly studied by Osmond (1954).

The Sevy Dolomite underlies many small areas on the crest and the west flank of the Schell Creek Range, the steep west flank of the prominent hill 1–3 miles south of the mouth of Steptoe Canyon, and the west flank of the range just south of the mouth of the canyon of Cooper Wash. No complete section of the Sevy can be found in this quadrangle, for it appears only in much-faulted areas and is not generally well exposed. A section can be pieced together, however, for the basal beds, at least, are exposed on the north wall of the large canyon 2.5 miles east-northeast of Cave Creek Reservoir, most of the middle beds are exposed on the hill south of Steptoe Creek, and the beds near the top are exposed south of Cooper Wash.

The Sevy Dolomite generally forms gentle to moderately steep slopes broken by many persistent narrow ledges and a few less continuous cliffs. The ledges and intervening benches are much more regular and continuous than those formed by other rocks. Only the slopes formed by part of the Guilmette Formation of Middle and Late Devonian age, and by part of the Ely Limestone, of Mississippian, Pennsylvanian, and

Permian age, remotely resemble those formed by the Sevy Dolomite, but the ledgy units in these other formations are brownish or yellowish gray, whereas those in the Sevy are very light gray.

In most places the base of the Sevy is marked rather sharply by the main body of fine-grained dolomite, but in a few places similar beds are also present in the uppermost part of the dominantly coarse-grained underlying Laketown Dolomite, and exposures are too scarce for one to follow the contact in detail.

The Sevy Dolomite is a medium-light-gray to dark-gray rock that weathers to a distinctive very light gray to light gray. The beds underlying the benches are faintly light brownish gray, but this variation in hue is apparent only nearby, for at a distance the Sevy is masked by the color of the ledges and of the rubble derived from them. The rock is very fine grained to aphanitic. The beds generally break along small fractures rather than along the distinct bedding planes that are spaced 1–5 feet apart. The bench-forming beds are faintly laminated. Finely crenulated stylolites in the uppermost beds are exposed near Cooper Wash. Most of the Sevy contains little detrital material, but the darker laminae appear to be slightly silty.

Where best exposed on the steep slopes of the hill south of the mouth of Steptoe Canyon, the Sevy is divisible into two ledge-forming units separated by a cliff-forming unit. The lower ledge-forming unit is about 340 feet thick. The beds in it are commonly about 1–2 feet thick, but some of those near the top are as much as 5 feet thick. Some of the ledge-forming beds in this unit contain a sedimentary breccia, and some of the intervening bench-forming beds are faintly laminated and are separated by minor disconformities. The middle, cliff-forming unit is 170–220 feet thick. Some of it is very indistinctly bedded, and some of it is massive and weathers to slightly rounded bosses. Toward the more gently sloping north and south spurs of the hill, the cliff is so subdued that this unit is no longer mappable; the cliffs are only on exceptionally steep slopes. The upper ledge-forming unit, which is at least 340 feet thick, resembles the lower one except that some of its beds form rounded ledges rather than the usual blocky ones. In this section the upper unit is overlain by a low-angle fault that cuts out the overlying Simonson Dolomite, part of the Guilmette Formation, and possibly also the uppermost beds of the Sevy Dolomite itself. A mile farther south some intraformational conglomerate and minor disconformities appear near the top of the Sevy.

Along Tamberlaine Canyon and on the high ridge southwest and south of Kolcheck Basin, there is a sandy layer, less than 20 feet thick, near the top of the Sevy. Southwest of the basin, quartz sand is very abundant in this layer, which there includes a few light-gray to very pale brown quartzite beds, but elsewhere makes up no more than 50 percent of this layer. Osmond (1954, p. 1918–1920, fig. 9) showed only a few feet of sandy dolomite near Cooper Wash, but he remarked that a dolomite containing less than 30 percent of quartz sand is hard to recognize in the field as a sandy bed. In some places the grains are frosted. The quartz grains in this layer range in diameter from 0.05 to 0.5 mm and are commonly subrounded to well rounded. Rocks containing abundant quartz also contain more angular grains that interpenetrate and are partly cemented by secondary silica, which may have been derived from the outer parts of the angular interpenetrating grains. In other words, angularity of these grains is considered a diagenetic feature. The quartz sand is very clean and contains only a few grains of iron oxide and leucoxene; the quartz grains enclose a very few small crystals of rutile, tourmaline, and apatite.

The thickness of the Sevy Dolomite is estimated to be about 900 feet near Steptoe Canyon, but inasmuch as the top of the formation is faulted and the position of the base is uncertain, this thickness is a minimum. If the measured dip of the Sevy near Steptoe Canyon is 5° too low, the formation may be as much as 1,050 feet thick. No more than 500 feet of the Sevy is exposed in any block northeast of Cave Creek.

The Sevy Dolomite contains few fossils, and only three collections were made from it in the Connors Pass quadrangle. W. A. Oliver, Jr. (written commun., 1961), identified *Ferestromatopora* sp. and an amphiporoid stromatopora in one collection, *Clathrodictyon* sp. in the second collection (written commun., 1959), and indeterminate brachiopod fragments in the third. J. T. Dutro, Jr. (written commun., 1961), identified *Cryptonella?* sp. in the third collection. These fossils have a long age range, but the ages are all compatible with those assigned to the formation in other areas. Nolan (1935, p. 19) dated the type Sevy Dolomite as Devonian, and probably Middle Devonian, because of the unconformity at its base and its gradation at the top into well-dated Middle Devonian rocks. Osmond (1954, p. 1928–1929) favored a Late Silurian and Early Devonian age for the Sevy, placing less emphasis on the nature of the top of the formation than Nolan did and more on the occurrence of *Halysites* sp., *Halysites* cf. *H. catenularia* Linné, which he found low in strata correlated with the Sevy Dolomite in the Ninemile Canyon area of the southern part of the Egan Range. He also inferred that the Sevy was transgressive, being older in the west than in the east. I regard the Sevy Dolomite in this quadrange as mostly Early Devonian and in minor part Middle Devonian, but it may contain some rock older than Early Devonian.

Sevy Dolomite occurs throughout most of eastern Nevada and western Utah, where it is of uniform lithology and is 500–1,500 feet thick, the thinner sections being mostly near the Connors Pass quadrangle. It is noteworthy that the fine-grained Sevy Dolomite lies near the middle of the thickest part of the middle Paleozoic dolomite sequence and that it overlies the only widespread unconformity in these rocks.

The origin of the Sevy Dolomite seems to be somewhat different from that of the adjacent coarse-grained dolomite. The removal of dolomite during the widespread hiatus preceding deposition of the Sevy and the resolution of this dolomite by the sea perhaps caused especially high concentrations of magnesium in the Sevy sea (Osmond, 1954, p. 1930) and the primary deposition of the fine-grained dolomite. Such a situation would have been ideal for the deposition of primary dolomite, or perhaps for relatively rapid replacement of a calcareous sediment.

SIMONSON DOLOMITE

The Simonson Dolomite, a coarse-grained brown dolomite 600–700 feet thick that includes some strongly laminated beds, conformably overlies the Sevy Dolomite. This formation, which is slightly more extensive than the underlying formation, was named the Simonson Dolomite by Nolan (1935, p. 19) for exposures in Simonson Canyon in the northwest part of the Deep Creek Range. Osmond (1954) divided the Simonson into four informal members: a basal buff member, a lower alternating member, a brown cliff-forming member, and an upper alternating member.

The Simonson Dolomite is exposed in several small areas along the crest and the west flank of the Schell Creek Range between Cave and Cleve Creek, along the east side of Cooper Wash, and along Tamberlaine Canyon. The best section, and probably the one measured by Osmond, lies on the spur south of the mouth of the canyon, tributary to Steptoe Creek, that runs through the northernmost outcrops of volcanic rocks along the west flank of the range (pl. 1). Slopes on the Simonson Dolomite are darker than those on the Sevy Dolomite and are only faintly marked by ledges.

The base of the formation is marked by the lowest appearance of a yellowish-gray to light-brownish-gray medium- to coarse-grained thick-bedded to massive dolomite. In the lower part of the basal buff member, a few fine-grained dolomite beds like those in the Sevy Dolomite alternate with the coarser grained rock, and the upper part of the member is faintly laminated and somewhat thinner bedded than the rocks below it.

In the lower alternating member, conspicuous laminae of light-gray to light-brownish-gray dolomite alternate with laminae of dark-brown dolomite (fig. 6).

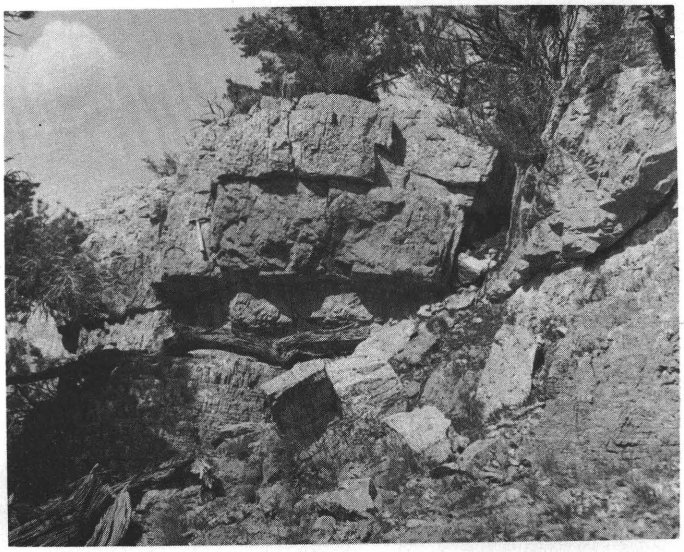


FIGURE 6.—Typical ledge in the lower alternating member of the Simonson Dolomite on the south wall of the canyon tributary to Cooper Wash at its mouth. Laminae are far more abundant and conspicuous than in the Fish Haven Dolomite.

Individual laminae are paper thin to almost an inch thick. The contacts between the laminae are commonly very regular, but in a few places they are highly irregular, having undulations as high as a quarter of an inch. Prevailingly dark beds in units a few feet to a few tens of feet thick form small benches and alternate with equally thick or slightly thicker cliffy units of lighter color. Minor unconformities are beneath some of the brown laminae within and possibly between the units. The brown laminae are commonly coarser grained than the gray laminae, and they contain fossil fragments. Some brown laminae also contain biostromal forms or are almost coquinas; others are bioclastic. The dark-brown units thicken upward within the member, and the lower alternating member grades into the overlying brown cliff-forming member.

In the southern part of the Cooper Wash area, the brown cliff-forming member consists of three separate brown cliffy dolomite units that are separated by weaker laminated rocks. Along Tamberlaine Canyon and 1.5 miles east of Cave Creek Reservoir, this member is relatively inconspicuous; in other places this member cannot be recognized. In some places there appears to be a single member composed of laminated dolomite that is of more than the normal thickness. Either the thicknesses of the three upper members are varied, or in places the brown cliff-forming member is absent and the alternating members are juxtaposed. The brown cliff-forming member contains some biostromal beds and some dark-gray chert lenses 1–2 inches thick and many feet long. In Tamberlaine Canyon the stratum that most resembles the brown cliff-forming dolomite member contains many thin beds of dark-gray fine-grained limestone.

The upper alternating member resembles the lower one, except that it contains widely scattered chert nodules. Near the top it grades upward from brownish-gray laminated dolomite to a medium-gray faintly laminated or unlaminated limy dolomite.

In the Cooper Wash area the Simonson Dolomite is estimated from structure sections to be 600–700 feet thick. The minimum thicknesses of the individual members are: basal buff member, 200 feet; lower alternating member, 150 feet; brown cliff-forming member, 100 feet; and upper alternating member, 150 feet. In what is presumably the same section, Osmond (1954, fig. 12) measured thicknesses of 207, 108, 87, and 131 feet for these respective members.

No identifiable fossils were collected from the Simonson within the quadrangle, although fragments of corals and brachiopods are locally abundant. Elsewhere in this region the formation is sparsely fossiliferous, and as a whole its fauna is unquestionably Middle Devonian. The age of the base of the formation cannot be accurately determined because of the general absence of fossils at this horizon, but regionally the top of the formation is at or just beneath the *Martinia kirki* zone or the next younger *Stringocephalus* zone, both of Middle Devonian age. In detail the upper contact appears to fluctuate slightly with respect to these zones, which apparently range through several hundred feet of strata. In the Connors Pass area and in the Hamilton area (Humphrey, 1960), about 40 miles to the west, fossils probably referable to the *Martinia kirki* zone appear in the lowest limestone that is assigned to the overlying Guilmette Formation or its equivalent. But in several other places the Simonson Dolomite contains fossils of the younger *Stringocephalus* zone. It seems more likely that our knowledge of the fossils and their range is not yet complete than that the four very persistent and distinctive members of the Simonson Dolomite cross time horizons so much at random.

The coarsely crystalline dolomite in the Simonson was probably derived from a calcareous mud, some remnants of which are still preserved as limestone lenses, for the obliteration of most fossils and of much sedimentary detail indicates that the dolomite is secondary. The origin of the light-gray beds is more problematical, for they are not as coarse grained as the dolomite and they do not contain remnants of fossils or limestone. Osmond (1965, p. 1952–1953) believed that these differences recorded a periodically changing environment, and he described in detail the kinds of fluctuation that may have occurred near sea level. He ascribed the gray beds to deposition or reworking at, or even above, sea level and believed that the brown beds were deposited at slightly greater depths, more favorable to organic life. One might wonder whether the color differences could be due to sorting of pigmented material during replacement by dolomite, as well as, or instead of, due to initial differences in the organic content of the sediments. The many minor disconformities in the alternating members of the formation indicate current activity, but the thinness of the laminae indicates that the currents were very gentle. Osmond explained this apparent contradiction as a result of periodic damping of wave activity, and offered several intriguing suggestions for the cause of this damping. The origin of the laminae may be alternatively explained, however, by changes in the local physio-chemical environment rather than by changes in the regional geographic environment.

GUILMETTE FORMATION

The Guilmette Formation, a sequence of limestone, dolomite, and a small amount of sandstone, is about 2,000 feet thick and conformably overlies the Simonson Dolomite. It was named by Nolan (1935, p. 20) for exposures in Guilmette Gulch, in the northeast part of the Deep Creek Range. In the Connors Pass quadrangle the distribution of the dolomite within the formation is highly irregular, and the informal members that have been mapped are therefore of merely local value.

The Guilmette Formation underlies about 9 square miles of the quadrangle. It is extensively exposed in the topographically high area between Cave and Cleve Creeks and between Tamberlaine Canyon and the Taylor mining district; small outcrops also occur along Steptoe Creek, Cave Creek, and Grasshopper Canyon. The formation is so much faulted that no complete section has been found, and the rapid and irregular lateral changes in lithology make it difficult to piece together a meaningful representative section. A comparatively unbroken section of a large part of the formation is fairly well exposed on the upper slopes of the east side of Cooper Wash about 1–2 miles north of the section in the Simonson Dolomite. A less complete but more accessible and more fossiliferous section extends across the south end of the prominent ridge 1 mile east of the Taylor mining district; the only outcrops of the highest member of the formation are near the north end of the ridge. A large part of the formation, not cut by any large faults, is fairly well exposed on the east slope of a small valley about 2 miles southeast of the mouth of Steptoe Creek.

The mixed limestone and dolomite generally form narrow benches and gentle slopes that alternate with discontinuous cliffs. The limestone in the lower part of the formation forms higher cliffs, and areas in which dolomite is abundant contain long irregular cliffs diversified by knobs, buttresses, and chutes. Most of the sandy, silty, and shaly beds form gentle slopes, but in many places the sandstone marker bed near the middle of the formation forms a small cliff.

Most of the Guilmette Formation consists of thick-bedded fine-grained gray limestone, thin-bedded fine-grained silty limestone, dark-brown coarse-grained dolomite, light- to medium-gray fine-grained dolomitic limestone, light-brownish-gray coarse-grained dolomite, and pale-yellow-brown limy siltstone, in various proportions and successions. The base of the formation is placed at the bottom of the lowest thick limestone bed, or the dolomitic limestone, whichever is lower, but this horizon is an irregular surface and probably ranges through several score feet of section. Silty rocks generally first appear near the middle of the formation and become more abundant upward. Dolomite is moderately abundant near the base of the formation, most abundant near the middle, and scarce near the top. A thin, somewhat lenticular sheet of quartz sandstone or quartzite slightly above the middle of the formation makes a distinctive marker. The bulk of the formation can be divided in most places into three informal members, called in ascending order a, b, and c, but these members cannot be identified where the proportion of coarse-grained dolomite is very high or where the local structure is complex. These rocks are overlain between Cooper Canyon and Cooper Wash by a thick sequence of reef limestone, thin-bedded limestone, siltstone, and subordinate sandstone, quartzite, and conglomerate, which I call member d.

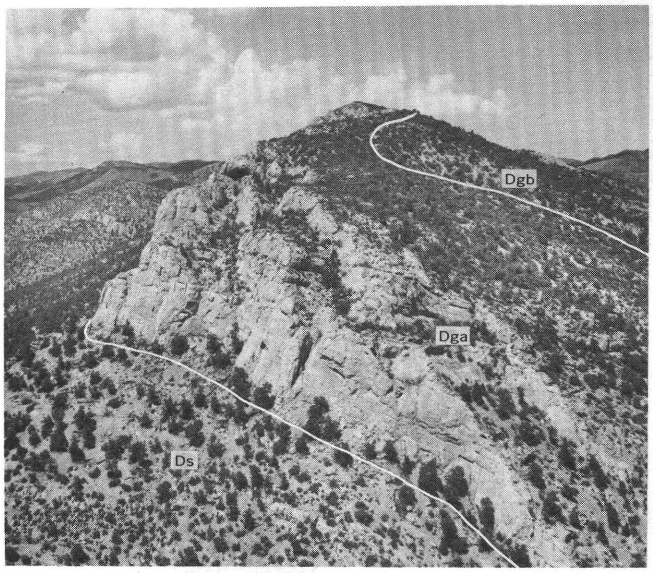


FIGURE 7.—Typical exposure of basal thick-bedded limestone of member a of the Guilmette Formation (Dga). The dip slope to the right is underlain by the overlying member b (Dgb). The gentle slope, broken by small ledges, in the left foreground is underlain by the upper part of the Simonson Dolomite (Ds). View is northward across the canyon tributary to Cooper Wash at the mouth of the canyon.

MEMBER A

The lowest member, a, of the Guilmette Formation is about 500–600 feet thick. East and southeast of the mouth of the canyon of Cooper Wash and northeast of Cave Creek it consists of a cliff-forming limestone, but between the mouths of the canyons of Cooper Wash and of Steptoe Creek it is mainly dolomite and the slopes are gentler. Part of the transition between the two facies of member a is illustrated in figure 7, a photograph of an area where the nearer cliffs consist mainly of limestone and the distant slopes are underlain by beds that are at about the same stratigraphic horizon but are more dolomitic. The base of the formation lies at the foot of the prominent cliff, and much of the rock exposed in the dip slope to the right of the cliff belongs to member b. East of the mouth of the canyon of Cooper Wash the cliffy limestone is underlain by a thin fine-grained thin-bedded gray dolomitic limestone, which, however, is faintly laminated and so much fractured that its structural relations to the cliffy limestone are uncertain. Southeast of the mouth of this canyon, member a consists mainly of sparsely fossiliferous and slightly bluish gray cherty fine-grained limestone. Here thick lenses of cliff-forming thick-bedded to massive limestone are separated by thin-bedded limestone. Inasmuch as faults are along the periphery of at least the southernmost lens, it is uncertain whether the lenses are primary features or were formed by shearing. Northeast of Cave Creek, siltstone partings and silty dolomite appear low in member a, which there contains widely scattered chert nodules and sparse remnants of brachiopods, gastropods, and crinoids.

Between the mouths of Cooper Wash and Steptoe Creek, member a is more dolomitic than elsewhere.

East of the wash a few dark-brown coarse-grained dolomite beds 3–15 inches thick crop out about a mile north of the last cliffy limestone. Just northwest of the wash, on the southwest spur of hill 9,085, dark-brown dolomite, gray dolomite, and limestone are about equally abundant, and it is therefore hard to separate member a from member b. On this hill, member b is underlain by about 800 feet of massive to very thick bedded coarsely crystalline light-pinkish-gray to light-brownish-gray dolomite. Although a minor thrust fault, which brought some of the Guilmette Formation over Chainman Shale and some of it over other rocks of the Guilmette Formation, complicated the situation, much of the light-brownish-gray dolomite is probably in member a. Between the 9,085-foot hill and Steptoe Creek, member a is more than 300 feet thick and consists mainly of dark-brown dolomite.

At several places in this area of abundant dolomite, local sedimentary features show some interesting relations between several varieties of dolomite. East of Cooper Wash some of the brown dolomite beds contain algal heads or stromatoporoids that enclose pockets of light-gray dolomite. On the southwest spur of hill 9,085 is a lens of intraformational conglomerate, 1 foot thick and 5 feet long, that consists of small angular pebbles of gray dolomite, dark-brown dolomite, and chert in a gray matrix of calcareous dolomite. Along the front of the range northeast of the hill there is more intraformational sedimentary breccia and conglomerate, consisting of gray-dolomite fragments in a matrix of dark-brown dolomite. The breccia grades laterally from larger rectangular blocks of gray dolomite separated by small septa of brown dolomite to unbroken gray dolomite. The relations between the several kinds of coarse-grained rock suggest that such rocks were formed at various times, but all of them are younger than the fine-grained dolomite and possibly younger than the chert. Furthermore, the fine-grained dolomite and chert were sufficiently consolidated to break into angular fragments before the coarse dolomite was formed. Clearly, however, inferences about the origin of dolomite drawn from one outcrop in this area cannot be applied too generally.

MEMBER B

Member b of the Guilmette Formation is 300–700 feet thick and consists of alternating layers of dark-brown coarse-grained dolomite and shaly or thin-bedded limestone. In some places this member also contains a small amount of the light-brownish-gray coarse-grained dolomite and a thin marker bed of sandstone or quartzite.

The limestone is commonly a slightly bluish to medium-gray rock containing crinkly beds ¼–4 inches in thickness, separated by yellow-brown silty partings. Many of the limestone beds are fossiliferous, and many fossils weather out and lie on the surface. *Atrypa* spp. and athyrid and spiriferid brachiopods are very common, and some corals and brachiopods are generally present. This fossil assemblage and weathering habit occur most commonly in rocks of the lower or middle parts of member b, but some speciments of *Atrypa* are found in a slightly younger assemblage, and some of the fossils of member c also weather free from the enclosing rock.

Dolomite is most abundant in member b in hill 9,085 and is least abundant south of Cooper Wash. Most of it is dark brown and thin bedded to moderately thick bedded, but some of the light-brown massive dolomite on hill 9,085 may belong to member b. Small dark-brownish-gray chert nodules are widely scattered in dark-brown dolomite. Locally the dark-brown dolomite interfingers with the limestone; the contacts between these rocks, which are mostly sharp, generally extend along the more conspicuous bedding planes, but in some places they cut across them. Faint laminae, along which the rock does not part, end abruptly against the dolomite, but some laminae also continue across the contact. Most sedimentary details obviously were obliterated during the replacement of the limestone by dolomite. Tongues of dolomite replace slightly more shaly limestone rather than thick-bedded limestone.

A quartzitic sandstone bed that has been mapped as a marker lies either within member b or in the basal part of member c. It is commonly 15–30 feet thick, but in some places it is absent and in other places it is as much as 60 feet thick. The rock is light gray and weathers grayish pink to very pale orange. It is faintly laminated, and generally parts along planes 2 inches to 4 feet apart.

In thin section it is seen that clastic grains generally make up about 50 percent of the rock, or rarely as much as 70 percent. They consist almost entirely of quartz, which contains a few inclusions of tourmaline, apatite, and zircon, and also very small particles of black opaque iron oxide(?), tourmaline, zircon, and leucoxene. The quartz grains are 0.1–0.9 mm in diameter and are subangular to well rounded. The grains are commonly loosely packed, with point-to-point or line-to-line contacts, in a calcite cement. Less commonly the quartz grains interlock or are separated by a little quartz or clay cement.

The upper part of member b contains less dolomite and more silty limestone than the lower part, and it includes some yellowish-brown limy siltstone. It is as fossiliferous as the lower part, but contains more corals and fewer brachiopods.

MEMBER C

Member c of the Guilmette Formation is about 400–600 feet thick and is more uniform in character than the underlying members. Most of it consists of silty limestone and limy siltstone, but it also contains some thin layers of dolomite. The limestone is generally fine grained, slightly bluish gray, and thin bedded although a few layers less than 20 feet thick contain beds as much as 4 feet thick. Some of these layers contain abundant corals and algae or stromatoporoids and may be biostromes. Some of the limestone also contains a little nodular to slightly lenticular dark-gray chert that weathers light brown.

MEMBER D

In one small area on the east side of the divide between Cooper Canyon and Cooper Wash, about 600 feet of limestone, interbedded with small amounts of sandstone, quartzite, and conglomerate, overlies member c of the Guilmette Formation with apparent conformity and is unconformably overlain by the Pilot Shale. These rocks are member d of the Guilmette Formation.

The basal 50 feet of member d consists of platy limestone, shale, sandstone, quartzite, conglomerate, and sedimentary breccia. The coarser clastic rocks appear to be lenticular, for they are less extensive along their strike than the finer grained rocks. The quartzite, the most abundant of the clastic rocks, is light gray to light olive gray and weathers pale reddish brown. As seen in thin sections, about 90 percent of the quartzite consists of angular to subrounded quartz grains only 0.02–0.2 mm in diameter. A few clastic grains of zircon and tourmaline are also present. The grains are cemented by fairly abundant clay, sericite, and traces of iron oxide, calcite, and chalcedony. The breccia, forming beds 4–12 inches thick, consist of medium-gray to brownish-gray quartzite fragments. The small grain size, poor sorting and rounding of grains, and large clay content and weathering color of the sandy rocks are more characteristic of those of Mississippian age than of those of middle Paleozoic age; apparently they are harbingers of a changing tectonic environment.

The upper 550 feet of member d has more shaly limestone than the lower part and contains a few thick biostromal limestone beds resembling the rocks in member c. Near the top of member d there are conspicuous outcrops of reef limestone containing a rich fauna of partly silicified corals.

FAUNA, AGE, AND CORRELATION

The limestone beds of members b, c, and d of the Guilmette Formation are among the most fossiliferous rocks in the area. Brachiopods are very abundant and well preserved in the shaly limestone of the lower part of member b. Many of the shells are phosphatic, and the shells weather free with both valves intact. Higher in member b corals are more abundant, the brachiopods more varied, and the *Atrypa* perhaps somewhat larger. In member c the shaly beds contain abundant brachiopods and the limy beds contain corals. Silicified *Cladopora*? is especially abundant near the top of the member. Corals are even more abundant near the top of member d than in member c. All fossils identified from the Guilmette are listed in table 3.

J. T. Dutro, Jr., stated:

> The fossils of the Guilmette Formation are predominantly of Late Devonian age although the poorly preserved material from member a may well be Middle Devonian. Member b contains a predominantly brachiopod assemblage that is of early Frasnian age and probably represents the "*Spirifer*" *argentarius* zone of Merriam (1963, p. 53). All the fossils from members c and d are also of Frasnian age, and representatives of both the *Pachyphyllum* and "*Martinia*" *nevadensis* zones of Merriam are present. As Merriam has indicated (1963, p. 54), these two are facies assemblages of approximately the same age significance. The brachiopod-rich assemblage in member c can be compared favorably with the Sly Gap fauna of New Mexico (Stainbrook, 1948). Common species and genera between the two assemblages are: *Calvinaria bransoni* Stainbrook, *Devonoproductus vulgaris* Stainbrook, *Douvillina*, *Gypidula munda* Stainbrook, *Nervostrophia*, *Nudirostra* spp., *Pugnoides schucherti* Stainbrook, *Spinatrypa*, *Thomasaria*, *Warrenella* spp., *Alveolites*, *Macgeea*, *Pachyphyllum*, *Tabulophyllum*, and *Phacellophyllum*. No representatives of Merriam's *Cyrtospirifer* zone (post-Frasnian) have been identified from this quadrangle.

ENVIRONMENT

During the early part of Guilmette time, the quiet deposition that had been going on in much of the eastern Great Basin since Middle Ordovician time continued. In the Connors Pass area, as elsewhere through Guilmette time, progressively less dolomite and more limestone were deposited, this change suggesting that concentration of magnesium in the sea was significantly reduced. As magnesium content was reduced, precipitation of, or replacement by, the available magnesium depended increasingly on local environment, and consequently the dolomite beds are less continuous.

The macroscopic sedimentary features indicate that both the light-brownish-gray and the dark-brown coarse-grained dolomite types replaced limestone and the fine-grained gray dolomite. At least the dark-brown dolomite preferentially replaced shaly dolomite, and perhaps the light-brownish-gray dolomite also preferentially replaced the more massive and coralline limestone. Locally, pebbles and blocks of each coarse-grained type are embedded in the other, so there is not a clear-cut age relation between them. Both contain sedimentary fragments of the fine-grained gray dolomite, in which inclusions of the coarse-grained dolomite do not appear. Apparently the fine-grained gray dolomite is older than the coarse-grained types, and conceivably it is a primary dolomite. If these deductions are valid, deposition of primary dolomite, or very early replacement dolomite, ceased generally during early Guilmette time and ceased entirely in about middle Guilmette time, whereas the replacement by secondary dolomite reached a climax during middle Guilmette time and continued into late Guilmette time.

The sandy rocks of member d are the oldest deposits in the Connors Pass area that reflect orogenic movements occurring near the end of Guilmette time in northwest Utah and south-central Nevada. The reef limestone above these clastic beds indicates that the seas were shallow, and thus for a while the clastic sediments could not be transported westward beyond the downwarping area near the Thomas and Dugway Ranges and did not spread more widely until downwarping had ceased and the basin was filled with sediments.

DEVONIAN AND MISSISSIPPIAN SYSTEMS

PILOT SHALE

The Pilot Shale, a dark shale and siltstone as much as 480 feet thick, overlies the Guilmette Formation with apparent disconformity. Spencer (1917) named this shale for outcrops at Pilot Knob, west of Ely. The Pilot Shale commonly forms the lowest of three conspicuous units, two of shale and a middle one of limestone, largely of Mississippian age. The Pilot is the lower part of Hague's White Pine Shale (1892). The name White Pine was later abandoned and replaced in the Eureka district by the Pilot, Joana, and Chainman Formations (Nolan and others, 1956). The Pilot Shale has been widely recognized in eastern Nevada and western Utah, but has nowhere been intensively studied.

In the Connors Pass quadrangle the Pilot Shale occupies about 2 square miles in the northern two-thirds of the west side of the Schell Creek Range. It generally forms narrow benches, gentle slopes, and saddles between outcrops of resistant limestone formations. In many places it forms a veneer on the dip-slope side of ridges eroded in the Guilmette Formation (fig. 7). In many other places it forms talus-covered slopes below cliffs of the overlying Joana Limestone. Slopes on the Pilot Shale are generally treeless.

Outcrops of Pilot Shale are few and small, and nowhere is as much as 5 percent of the formation exposed. The best exposures in the area are those on the south side of Square Top Hill, south of the lower part of Steptoe Creek, and those just north of the large silicified area in the Taylor mining district. Many slopes formed by Pilot Shale are covered with large chips and small plates of siltstone typical of many of the exposures, which, though useful in identification of the formation, mask the rarely exposed shalier rocks.

The contact between the Pilot Shale and the Guilmette Formation is commonly smooth and unfaulted. The thicknesses of both units are fairly uniform and, because the transition from limestone to shale is gradational, the contact easily can be assumed to be conformable. The base of the Pilot Shale was placed at the top of the highest thick limestone of the Guilmette Formation, a position probably slightly higher than that chosen by Clark and Becker (1960, fig. 2). At one place, however, in a small valley that drains eastward from the divide between Cooper Wash and Cooper Canyon, the Pilot Shale is underlain by member d of the Guilmette Formation. The contact there is covered, but a disconformity is inferred because of the very small distribution of member d and because the contact dips 20–30° more gently than the Pilot and Guilmette rocks adjacent to the contacts. The relations resemble those on the bedding-plane thrust faults common in the older Paleozoic rocks, but there is much less evidence of movement along this contact than along most of the contacts known to be faulted. Presumably, therefore, this contact is a disconformity and member d is preserved only very locally as a hill beneath the Pilot Shale.

The Pilot Shale consists mainly of shale and siltstone but includes some quartzite, limestone, and argillite. The proportions of these constituents are unknown because the finer grained rocks are generally masked by surface debris from the coarser grained ones. Shale and siltstone are probably about equal in abundance, and the other kinds of rock together probably make up less than 5 percent of the formation.

Siltstone is more abundant in the lower part of the formation than in the upper part. It is dark gray and commonly weathers to pale-yellow-brown chips and plates, mostly 2–4 inches across but some as much as 10 inches across. Some of the plates are reddish brown, especially near the base of the formation where a few thin beds of limestone are present.

240-303 O - 67 - 3

TABLE 3.—*Fauna of the*

[Most brachiopods identified by J. T. Dutro, Jr. (written commun., 1960, 1961), corals by W. A. Oliver, Jr. (written commun., 1959, 1960, 1961), and a few mixed collections by
argentarius zone; 10, *Phillipsastraea* o r *Martinia*

Unit (pl 1.)	Member a			Member b															Member c, lower part										
Age	Middle Devonian			Middle Devonian or Late Devonian						Late Devonian																		
Faunal zone	?		8?	8? or 9						9	9?	9		9?	?		9	10?	10									
Field No.; U.S. Geol. Survey colln. number in parentheses	58D48	58D50	58D73	59D225	60D422	60D375	60D533	60D539	60D436	60D378	(5803–SD) 59D256	60D393	60D394	(5797–SD) 59D223	59D249	(5801–SD) 59D248	60D480	60D484	60D398	(5804–SD) 59D257	60D379	60D506	60D481	60D537	60D356	60D380	60D391	60D420	60D460
Brachiopods:																													
Atrypa devoniana Webster																													cf.
Atrypa sp			×?			×				×		×											×	×					
Calvinaria cf. *C. bransoni* Stainbrook																													
Calvinaria sp																							×?	×	×?	×?		×?	
Crurithyris sp	×	×?																				×?							×?
Cryptonella sp																													
Cyrtina sp																		×?											
Cyrtospirifer placitus Stainbrook																		cf.											
Devonoproductus cf. *D. vulgaris* Stainbrook																													×
Devonoproductus sp																						×							
Douvillina sp																													×?
Eleutherokomma sp					×?			×						×						×?							×?		
Elytha sp																										×?			
Gypidula aff. *G. munda* Stainbrook																													
Hypothyridina sp. a (of Merriam, 1940)																													×
Hypothyradina sp																													
Nervostrophia sp																													×
Nudirostra aff. *N. carya* (Crickmay)																								×				×	
Nudirostra sp																										×			
Productella sp			×?								×?																		
Pugnoides aff. *P. schucherti* Stainbrook																													
Pugnoides sp																													
Schizophoria sp																													
Spinatrype montanensis (Kindle)													cf.														aff.		
Spinatrypa sp. (large)												×					×												

See footnote at end of table.

Guilmette Formation

C. W. Merriam (written commun., 1959). Zones are numbered in ascending order in the Devonian rocks, as applied in the Eureka, Nev., area; 8, *Tylothyris* zone; 9, *Spirifer nevadensis* zone (Nolan and others, 1956; Merriam, 1940)]

Unit (pl 1.)	Member c, middle part		Member c, upper part																				Member d, lower part	Member d, upper part			
Age	Late Devonian																										
Faunal zone	10		10							10?	10				?	10							10	10?	10		
Field No.; U.S. Geol. Survey colln. number in parentheses	(5813–SD) 59D302	60D396	(5798–SD) 59D224	(5805–SD) 59D258	(5806–SD) 59D259	60D400	60D401	60D388	59D300	60D437	60D482	60D381	60D382	60D403	60D509	60D510	60D383	60D402	60D399	60D495	60D462	60D355	59D260	(5808–SD) 59D261	(5809–SD) 59D262	(5811–SD) 59D264	59D263
Brachiopods:																											
Atrypa devoniana Webster							cf.							aff.													
Atrypa sp										×	×					×				×					×	×?	
Calvinaria cf. *C. bransoni* Stainbrook														×													
Calvinaria sp			×	×													×?	×?	×?		×?						
Crurithyris sp																×?											
Cryptonella sp	×																										
Cyrtina sp																											
Cyrtospirifer placitus Stainbrook					aff.																						
Devonoproductus cf. *D. vulgaris* Stainbrook												×				×											
Devonoproductus sp							×																				
Douvillina sp																											
Eleutherokomma sp							×?																				
Elytha sp																											
Gypidula aff. *G. munda* Stainbrook																				×							
Hypothyridina sp. a (of Merriam, 1940)													×														
Hypothyradina sp										×?																	
Nervostrophia sp																											
Nudirostra aff. *N. carya* (Crickmay)																											
Nudirostra sp								×							×		×							×?			
Productella sp																											
Pugnoides aff. *P. schucherti* Stainbrook														×							×						
Pugnoides sp																				×							
Schizophoria sp														×							×						
Spinatrypa montanensis (Kindle)													cf.														
Spinatrypa sp. (large)																											

TABLE 3.—*Fauna of the*

Unit (pl 1.)	Member a			Member b															Member c, lower part										
Age	Middle Devonian			Middle Devonian or Late Devonian						Late Devonian																		
Faunal zone	?		8?	8? or 9						9	9?	9		9?	?		9	10?	10										
Field No.; U.S. Geol. Survey colln. number in parentheses	58D48	58D50	58D73	59D225	60D422	60D375	60D533	60D539	60D436	60D378	(5803–SD) 59D256	60D393	60D394	(5797–SD) 59D223	59D249	(5801–SD) 59D248	60D480	60D484	60D398	(5804–SD) 59D257	60D379	60D506	60D481	60D537	60D356	60D380	60D391	60D420	60D460
Spinatrypa sp					×		×	×		×				×		×		×				×							
"*Spirifer*" cf. "*S.*" *argentarius* Meek													×																
"*Spirifer*" cf. "*S.*" *strigosus* Meek																						×							
Spiriferoid brachiopod, indet						×	×		×	×																			
Stropheodonta sp											×?							×?					×?						
Tenticospirifer sp																													
Tylothyris sp			×?																										
Thomasaria cf. *T. altumbona* Stainbrook																													×
Warrenella cf. *W. eclectea* Crickmay																							×						
Warrenella sp																													×?
Corals:																													
Alveolites sp																													
Aulopora sp				×?																×									
Auloporid coral																													
Acinophyllum sp																	×?												
Breviphyllum sp																						×?							
Chonophyllum sp. cf. *C. infundibulum* (Meek)																													
Cladopora sp																					×?								
Horn corals indet				×																×									
Macgeea sp																			×	×?	×?		×[1]						
Pachyphyllum sp																													
Pachypora sp																	×												
Phacellophyllum sp																					×		×[1]						
Phaceloid coral species indet				×																									
Ramose favositoid coral																				×									
Rugose corals indet																													
Striatopora sp																													
Syringopora sp																													
Tabulophyllum sp																			×	×		×?							
Thamnopora sp				×?																		×							
Thamnoporid coral																													
Gastropods:																													
Murchisonia sp	×	×?																											
Bellerophontid gastropod (small)	×	×																											
Platyceratid gastropod, indet																													×

See footnote at end of table.

Guilmette Formation—Continued

Unit (pl 1.)	Member c, middle part		Member c, upper part																				Member d, lower part	Member d, upper part			
Age	Late Devonian																										
Faunal zone	10		10							10?	10				?	10							10	10?	10		
Field No.; U.S. Geol. Survey colln. number in parentheses	(5813–SD) 59D302	60D396	(5798–SD) 59D224	(5805–SD) 59D258	(5806–SD) 59D259	60D400	60D401	60D388	59D300	60D437	60D482	60D381	60D382	60D403	60D509	60D510	60D383	60D402	60D399	60D495	60D462	60D355	59D260	(5808–SD) 59D261	(5809–SD) 59D262	(5811–SD) 59D264	59D263
Spinatrypa sp.					×																×						
"*Spirifer*" cf. "*S.*" *argentarius* Meek																											
"*Spirifer*" cf. "*S.*" *strigosus* Meek							×					×													×		
Spiriferoid brachiopod, indet.	×																										
Stropheodonta sp.					×	×?													×?								
Tenticospirifer sp.												×															
Tylothyris sp.																											
Thomasaria cf. *T. altumbona* Stainbrook																											
Warrenella cf. *W. eclectea* Crickmay				×															×								
Warrenella sp.			×?														×?	×?									
Corals:																											
Alveolites sp.		×					×				×											×			×		×
Aulopora sp.																											
Auloporid coral								×																			
Acinophyllum sp.																											
Breviphyllum sp.						×			×?																		
Chonophyllum sp. cf. *C. infundibulum* (Meek)											×?																
Cladopora sp.									×	×		×										×	×		×	×	
Horn corals indet.			×																						×	×	
Macgeea sp.												×											×?	×?	×	×	
Pachyphyllum sp.	×	×																							×		
Pachypora sp.																									×		
Phacellophyllum sp.										×	×											×					×
Phaceloid coral species indet.																											
Ramose favositoid coral																											
Rugose corals indet.																						×					
Striatopora sp.																									×?		
Syringopora sp.		×																									
Tabulophyllum sp.			×	×		×		×?	×		×		×										×		×		
Thamnopora sp.		×					×															×					
Thamnoporid coral														×													
Gastropods:																											
Murchisonia sp.																											
Bellerophontid gastropod (small)																											
Platyceratid gastropod, indet.																											

TABLE 3.—*Fauna of the*

Unit (pl 1.)	Member a			Member b															Member c, lower part										
Age	Middle Devonian			Middle Devonian or Late Devonian						Late Devonian																			
Faunal zone	?		8?	8? or 9						9	9?	9		9?	?		9	10?	10										
Field No.; U.S. Geol. Survey colln. number in parentheses	58D48	58D50	58D73	59D225	60D422	60D375	60D533	60D539	60D436	60D378	(5803-SD) 59D256	60D393	60D394	(5797-SD) 59D223	59D249	(5801-SD) 59D248	60D480	60D484	60D398	(5804-SD) 59D257	60D379	60D506	60D481	60D537	60D356	60D380	60D391	60D420	60D460
Gastropod indet. (high and medium spired)																													
Gastropod indet. (small, high spired)									×																				
Gastropod indet.																				×									
Others:																													
Pelecypod indet. (large)																							×						
Tentaculites sp.																										×			
Styliolina		×																											
Stictostroma sp.															×?														
Stromotoporoid indet						×																							
Fenestrate bryozoans, indet.																													
Echinoderm debris	×																												

[1] Either one or the other coral present.

Shale is most common in the upper part of the formation. It is generally a dark-gray to olive-gray noncalcareous silty clay shale that breaks into very thin hard plates or fissile chips and is faintly color laminated. In one place the soil formed on the shale swells when wet. A few beds 1–2.5 inches thick of argillite having a dull to semivitreous luster are scattered in the upper part of the formation. Near the top there are also some pale-red to moderate-red siltstone and quartzite beds as much as 3 inches thick, and here some of the shale is calcareous. One typical siltstone bed high in the Pilot Shale has a grain size of 0.02–0.04 mm. Most of the grains are subangular to subrounded quartz and calcite but include some iron oxide, chlorite, tourmaline, and a hydrocarbon. The fairly abundant cement consists of calcite, clay, and iron oxide.

Semiquantitative spectrographic analyses were made of five samples of Pilot Shale for comparison of their trace-element content (table 5) with that of the Chainman Shale. Sample 23 is a platy, fissile, paper shale from near the middle of the formation, sample 24 a slightly silty shale from its upper part, and sample 25 a silty shale about 50 feet above its base. Sample 26 is a black thin, platy, paper shale having a little argillite; it was taken from three outcrops within a stratigraphic range of 10 feet near the middle of the formation. Sample 27 is from a blocky shale about 80 feet from its top.

The thickness of the formation has been scaled from several structure sections, because the exposures are too poor to make direct measurements very meaningful. Where not obviously faulted, the formation ranges in thickness from 320 to 480 feet; about half of the scaled estimates are at 400 feet, which seems a reasonable average for the area.

No fossils have been found in the Pilot Shale within the Connors Pass quadrangle, but there the age of the

Guilmette Formation—Continued

Unit (pl 1.)	Member c, middle part		Member c, upper part																				Member d, lower part	Member d, upper part			
Age	Late Devonian																										
Faunal zone	10		10							10?	10				?	10							10	10?	10		
Field No.; U.S. Geol. Survey colln. number in parentheses	(5813–SD) 59D302	60D396	(5798–SD) 59D224	(5805–SD) 59D258	(5806–SD) 59D259	60D400	60D401	60D388	59D300	60D437	60D482	60D381	60D382	60D403	60D509	60D510	60D383	60D402	60D399	60D495	60D462	60D355	59D260	(5808–SD) 59D261	(5809–SD) 59D262	(5811–SD) 59D264	59D263
Gastropod indet. (high and medium spired)								×																			
Gastropod indet. (small, high spired)																											
Gastropod indet.																										×	
Others:																											
Pelecypod indet. (large)																											
Tentaculites sp.								×																			
Styliolina																											
Stictostroma sp.																											
Stromotoporoid indet.										×															×		
Fenestrate bryozoans, indet.														×													
Echinoderm debris													×														

formation is bracketed between the early or middle Late Devonian age of the underlying Guilmette Formation and the Early Mississippian age of the overlying Joana Limestone. Some fossils of Late Devonian and others of Early Mississippian age have been found in the Pilot Shale in many nearby stratigraphic sections, and at several outcrops there is an unconformity within the formation or at its base.

The Pilot Shale is probably a finer grained facies of the predominantly argillaceous and quartzitic units in the lower part of the Eleana Formation (Poole and others, 1961), which was deposited near another high area about 150 miles to the southwest. During latest Devonian time, currents carried silt over a large area and very little calcium carbonate was deposited. Slightly finer material, perhaps winnowed out of the coarser clastic material deposited near the orogenic centers, continued to be deposited during earliest Mississippian time.

MISSISSIPPIAN SYSTEM

JOANA LIMESTONE

The Joana Limestone is a thick-bedded to massive crinoidal limestone 300–500 feet thick that conformably overlies the Pilot Shale. This formation was named the Joana Limestone by Spencer (1917), for the Joana mine a few miles west of Ely.

The Joana Limestone underlies about 3–4 square miles in the Connors Pass quadrangle; it occurs chiefly in the western half of the quadrangle but also in a small area at an altitude of about 7,600–8,000 feet that is west of Cleve Creek. It generally forms small rugged knobs, ridges, and cliffs that contrast markedly with the gentle slopes formed by the adjacent shale formations. Where the underlying shale has been deeply eroded, as along Cave Creek and Steptoe Creek, the cliffy outcrop is emphasized; there many cliffs overhang, making caves and natural shelters along their bases. Gentle dip slopes on the limestone are largely stripped bare of colluvium and are interrupted by large rounded ledges having strongly pitted and fluted surfaces. The middle and upper parts of the formation contain limestone layers that weather to narrow regular ledges and slabby debris.

Typical outcrops of the formation are most accessible at the junction of Cave Creek and Steptoe Creek. A more complete and unfaulted section is moderately well exposed 1–2 miles south of Cave Creek, and a thick section of the limestone may be seen on the west side of the upper part of Cooper Wash.

Much of the Joana Limestone is medium gray, thick bedded, bioclastic, crinoidal, and slightly cherty. The base of the formation is well exposed on the south side of Square Top Hill along Steptoe Creek and on the ridge about 1.5 miles N. 10° W. of the peak containing Taylor bench mark. Despite the slightly limy top of the Pilot Shale and the platy bedding at the base of the Joana Limestone, the contact is sharp. South of Cave Creek, where it is most fully exposed, the formation consists of five units, consisting alternately of platy and massive limestone. The lowest unit consists of only 15–20 feet of platy limestone. The basal beds of this unit are rarely more than 4 inches thick, and are partly nodular, are generally bioclastic, and in places contain brachiopods and horn corals. In the Cave Creek area a zone 10 feet thick contains abundant long lenses and beds of chert. Above the basal platy unit there is the lower massive limestone, about 200 feet thick, which contains scattered pods of dark-gray chert and scraps of fossils; this unit varies considerably in thickness, possibly because it is indistinguishable from the upper massive unit where the middle platy limestone is absent. The middle unit, 30–50 feet thick, consists of bioclastic silty platy limestone in beds 0.5–4 inches thick. The beds become thicker upward, and the middle unit grades into the upper massive limestone unit, about 150 feet thick, but less conspicuous and perhaps less extensive than the lower massive unit. At one outcrop, half a mile northeast of hill 9,085, rocks of the upper massive limestone unit contain abundant coral fragments and some colonial coral masses that resemble biostromal structures. The upper platy unit is about 60 feet thick and resembles the middle one. The beds are generally thinner toward the top of the upper platy unit, and they locally contain scour channels and minor unconformities. The unit contains much calcarenite rich in crinoid debris and some calcilutite. In some places where the Joana is thin but not obviously faulted, not all of these units can be recognized, possibly because obscure bedding-plane faults are more widespread in the formation than has been supposed, or because the lithology differs laterally, or because the upper contact is an unconformity that cuts out some of the units.

Where the Joana Limestone seems to be most complete, it is at least 480 feet thick, as scaled from several structure sections. Near Steptoe Creek and Cave Creek, however, where neither of the contacts is obviously faulted, there is only about 300–400 feet of this limestone. On the east side of the upper part of Cooper Wash, there is only 100–200 feet of Joana, again without obvious faults along the contacts. A similar thinning of Cambrian formations in this area was explained structurally, but the structural features are locally exposed near the Cambrian formations; whereas the Joana Limestone also thins in outcrops northwest of the quadrangle that are clearly unfaulted. Provisionally, then, most of the contacts of the Joana Limestone, even where it is thin, are mapped as normal stratigraphic contacts, on the assumption that the initial thickness of the formation differed widely.

Fossil fragments and single fossils are abundant in the formation, but assemblages of unbroken fossils are only moderately common in the platy limestone units and are scarce in the massive limestone units. All fossils identified from the Joana are listed in table 4.

TABLE 4.—*Fauna of Joana Limestone*

[Identified by Helen Duncan and MacKenzie Gordon, Jr. (written commun., 1962, 1963), and J. T. Dutro, Jr. (written commun., 1961)]

Field No.	58D12	59D226	59D265	59D291	59D301	60D354	60D439
U.S. Geol. Survey loc.	------	21235–PC	21236–PC	21244–PC	21247–PC	------	21258–PC
Brachiopods:							
Camarotoechia? sp.	------	------	------	------	------	×	------
Chonetes	------	------	------	------	×	------	------
Composita? sp.	------	------	------	------	------	×	------
Dimegelasma? sp. indet.	------	×	------	------	------	------	------
Eumetria? sp.	------	------	------	------	------	×	------
Eumetria cf. *E. verneuiliana* (Hall)	------	×	------	------	------	------	------
Marginatia sp.	------	------	------	------	×	------	------
Productelloid brachiopod	------	------	×	------	------	------	------
Productoid brachiopod cf. *Marginatia* sp. indet.	------	×	------	------	------	------	------
Schizophoria? sp. indet.	------	×	------	------	------	------	------
Spirifer cf. *S. centronatus* Winchell	×	×	×	------	×	------	------
Spirifer sp. indet.	×	------	×	------	------	------	------
Strophomenoid brachiopod (highly inflated form)	------	------	×	------	------	------	------
Strophomenoid brachiopod indet.	×	×	------	------	------	------	------
Corals:							
Homalophyllites sp.	×	------	------	------	×	------	------
Lithostrotionella sp. A	------	------	------	×	------	------	×
Vesiculophyllum sp.	×	------	------	------	×	------	------
Zaphrentites? sp. indet.	------	------	------	------	×	------	------
Others:							
Pelmatozoan debris	------	------	------	------	------	×	------
Fenestrate bryozoan indet.	------	×	------	------	------	------	------
Straparollus (*Euomphalus*) sp.	------	×	------	------	------	------	------
Crinoid pinnule	------	------	------	×	------	------	------
Loxonema sp. indet.	------	------	------	×	------	------	------
Straparollid gastropod indet.	------	------	------	×	------	------	------
Straparollus sp. indet.	------	------	------	------	×	------	------

Most of the fossils were collected to aid in solution of local structural problems and came from undetermined stratigraphic horizons; USGS colln. 21236-PC, however, is from the basal 12 feet of the formation, and USGS collns. 21235-PC and 21258-PC, collected a few hundred feet apart, are probably from the middle platy limestone unit. Most of the collections are assigned an Early Mississippian age, but some are not sufficiently diagnostic for so specific an assignment.

The widespread and relatively uniform bioclastic lithology of the Joana Limestone indicates a general return of the quiet conditions of deposition that prevailed during much of middle Paleozoic time. Within the Connors Pass quadrangle there are abundant signs of current activity and some evidence of small reefs. Only in the Cherry Creek Range, 60 miles northwest of the quadrangle (Langenheim, 1960), where the formation is very thin or absent, is there any suggestion of continued crustal instability; however, Langenheim did not mention any bedding-plane faults. In the Silver Island Range (Schaeffer and Anderson, 1960), 120 miles north of the quadrangle, the Joana Limestone rests unconformably on the Guilmette Formation.

CHAINMAN SHALE

The Chainman Shale, a dark-gray shale about 1,100 feet thick, conformably overlies the Joana Limestone. Spencer (1917) named this formation the Chainman Shale for exposures near the Chainman mine, a few miles west of Ely. The formation consists of weak rocks and therefore generally forms benches and valleys. It is widely known for its characteristic fauna of pelecypods and cephalopods. Many geologists have described the formation in specific localities, but it has not yet been comprehensively studied.

The Chainman Shale underlies almost 10 square miles in the western half of the quadrangle. It occurs chiefly in an area lying between Tamberlaine Canyon, the head of Cooper Canyon, and the southwestern corner of the quadrangle, but also in small areas northeast of Cave Creek, east of the crest of the Schell Creek Range along the southern edge of the quadrangle, and west of the North Fork of Cleve Creek. Particularly noteworthy in the distribution pattern of the shale is the fact that the basal contact is more irregular than the upper contact.

The formation underlies most of the broader benches and larger basins in the area. Slopes eroded on it are usually very gentle, except where they are cut by young gullies or are immediately beneath the Ely Limestone. In many places they are veneered with Tertiary and Quaternary gravels or conglomerates, this cover showing that the formation has underlain low areas for a long period of time. Some of these gentle slopes are interrupted by one or more small ridges or alined small knolls consisting of sandstone and quartzite interbedded with the shale, and fragments of these resistant rocks mantle the shale over large areas. A characteristically irregular, hummocky, slump topography is formed on the shale and is represented on the map (pl. 1) by an overprint symbol.

The Chainman is nowhere sufficiently well exposed to provide a basis for anything more than a general outline of its stratigraphy. Some exposures of its upper part can readily be seen in cuts along U.S. Highway 6–50–93 about a mile west of Connors Pass. Exposures of rocks low in the formation are most accessible at the dam of Cave Creek Reservoir and at a place about three-quarters of a mile farther east, on the west side of the valley tributary to Cave Creek from the south. Rocks in the middle and upper parts of the formation are conveniently exposed in Cooper Canyon a few hundred yards south of the head of the rough narrow road running south from the corral in the center of the basin centering in sec. 35, T. 15 N., R. 65 W., and also along the northwest side of the small volcanic plug near the road on the east side of the same basin.

The Chainman consists chiefly of dark-gray noncalcareous highly carbonaceous clay shale, but contain a small amount of coarser clastic rocks. Its base appears to be fairly sharp, but it is everywhere slumped or covered. The formation is roughly divisible into a lower shale and siltstone unit probably 200–400 feet thick, a middle shale unit at least 500 feet thick that contains few sandstone beds, and an upper shale unit about 300 feet thick that contains numerous small lenses of sandstone and quartzite and a little limestone. These units, however, are so broadly gradational and so much disturbed by landslides that they are not mappable.

Siltstone and mudstone are interbedded with the shale near the base of the formation, and some layers of silty rock are several tens of feet thick. The beds in the lower shale and siltstone unit are commonly ½–12 inches thick, have a roughly conchoidal fracture, and weather to irregular blocky slabs. Some of the siltstone is slightly micaceous and weathers pale yellow brown. The shale interbedded with it contains small carbonaceous nodules, a few of which have pyrite cores. Rocks in the upper part of this unit or near the base of the overlying middle shale unit contain a few cephalopods.

In many places near the middle of the Chainman, there are a few selenite veins. A lenticular marker bed of sandstone and quartzite about 50 feet thick appears near the middle of the formation in several places, such as on the high bench northeast of the Taylor mining district. This rock is moderately thick bedded and weathers to a pale-yellowish-brown blocky rubble.

Sandstone and quartzite in groups of beds a few feet to 10 feet thick are most common in the upper shale unit of the Chainman Shale, but even there they do not exceed 5 percent of the rock. In most places there are three to six such groups of beds, and the more prominent and extensive groups are locally mapped as markers. They are not extensive, probably having been deposited as lenses and some of them possibly having been pinched off by faulting and folding. These beds are commonly light gray, greenish gray, or pale yellowish gray on fresh fractures, and they weather brownish gray or olive gray. The sandstone beds are 1–22 inches thick but are most commonly about 4 inches thick. Faint laminae within them show crossbedding, and many surfaces have ripple marks, chiefly of current ripples. In thin section the diameters of the grains are seen to range from 0.05 to 0.5 mm and average about 0.1 mm. The grains are well sorted and alined, and most of them are subangular to subrounded, but some well-rounded and secondarily enlarged quartz grains are usually present.

Quartz grains make up 80–95 percent of the sandstone and quartzite. Other constituents are chert, iron oxide, leucoxene, zircon, apatite (as inclusions in quartz), tourmaline, muscovite, plagioclase (of intermediate An content), and in a few specimens possibly also pyroxene, sphene, and rutile. The quartz grains interlock or are closely packed, and the cementing material consists of clay and iron oxide, quartz, chalcedony, or calcite. One typical quartzite sample contains 15 percent by weight of soluble and clay-sized material and 0.3 percent of heavy minerals, of which zircon, leucoxene, and tourmaline of several varieties, including schorlite, each make up 25–30 percent; very small amounts of muscovite and oligoclase are attached to the tourmaline. These quartzite and sandstone beds are finer grained than those in member b of the Guilmette Formation and in older rocks, and

TABLE 5.—*Semiquantitative*

[*Italic* type indicates anomalously high values; --, not looked for; 0, looked for but

Sample loc. (fig. 14)	23	24	25	26	27	28	29	30	31	32	33	34
Field No. / Lab. No.	59D254 5088S	60D357 5090S	61D647 61-2972	61D649 61-2973	61D650 61-2974	59D165 5087S	60D353 5089S	60D440 5091S	61D569 61-2955	61D572 61-2956	61D577 61-2957	61Y595 61-2958
Formation	Pilot Shale					Chainman Shale						
Analyst	E. F. Cooley [1]		J. C. Hamilton [2]			E. F. Cooley [1]			J. C. Hamilton [2]			
Si	15	10	7	>10	>10	7	20	30	>10	>10	>10	>10
Al	--	--	3	3	3	--	--	--	7	7	7	3
Fe	2	3	.7	.7	.7	3	2	3	7	1.5	1.5	.7
Mg	>5	>5	7	7	1.5	1	>5	1	1.5	7	1.5	7
Ca	--	--	>10	>10	.7	--	--	--	.7	7	.3	7
Na	--	--	.3	.15	.07	--	--	--	.3	.3	.3	.15
K	--	--	3	3	3	--	--	--	1.5	3	1.5	3
Ti	.3	.3	.07	.07	.15	.3	.7	.7	.3	.3	.3	.07
P	--	--	0	0	0	--	--	--	0	0	0	0
Mn	.05	.1	.03	.015	.015	.03	.07	.01	.15	.03	.03	.03
Ag	<.0001	<.0001	0	0	0	*.002*	<.0001	.0001	0	0	0	0
B	.005	.005	.003	.003	.007	*.01*	*.01*	*.01*	.003	.007	.007	.003
Ba	.03	.05	.03	.015	.03	.01	.03	.07	.015	.07	.015	.03
Be	<.0001	<.0001	0	0	.00015	.0001	.0001	.0001	.00015	0	.00015	0
Co	<.001	.001	0	0	.0007	<.001	<.001	.001	.0015	.0007	.0015	0
Cr	.005	.01	.0015	.003	.003	*.1*	.007	.015	.015	.007	.015	.003
Cu	.005	.003	.0015	.003	.007	*.02*	.005	.015	.003	.003	.003	.003
Ga	<.002	<.002	.0003	.0003	.0003	<.002	<.002	<.002	.0015	.0007	.003	0
La	<.005	<.005	0	0	0	<.005	<.005	<.005	0	0	.003	0
Li	--	--	0	0	0	--	--	--	0	0	.03	0
Mo	<.0005	<.0005	0	0	.003	*.005*	<.0005	.001	.0007	.0007	0	0
Nb	--	--	0	0	0	--	--	--	0	0	.0015	0
Ni	.002	.002	.0007	.0015	.015	*.01*	.005	.007	.003	.003	.007	.0015
Pb	.002	.002	0	0	.0015	.003	.002	.002	.0015	0	.003	0
Sb	0	0	0	0	0	0	0	0	0	0	0	0
Sc	<.001	.001	.0007	.0007	.0007	.001	.001	.001	.0015	.0007	.0015	.0007
Sr	.02	.03	.015	.007	.03	.05	.01	.005	.03	.03	.03	.015
V	.005	.007	.003	.003	.015	*.1*	.007	.01	.007	.007	.015	.0015
Y	.002	.002	.0015	0	.0015	*.01*	.005	.003	.0015	.0015	.003	0
Yb	--	--	.00015	.00015	.00015	--	--	--	.0003	.00015	.0003	.00015
Zn	<.02	<.02	0	0	0	*.2*	<.02	<.02	0	0	0	0
Zr	.01	.007	.007	.003	.007	.01	.03	.02	.015	.015	.015	.003

[1] Analyses reprted in percent. Elements looked for but not found (in parts per million): As <1000, Bi <10, Cd <50, Ga <20, Ge <20, In <50, Nb <50, Sb <200, Sn <10, Ta <50, Ti <100, and W <50.

[2] Analyses reported to nearest numbers in the series 7, 3, 1.5, 0.7, 0.3, 0.15, and so forth, in percent. These numbers are midpoints of group data on a geometric scale, and the groups contain the quantitative value for about 60 percent of the analyses Elements also looked for but not found: As, Au, Bi, Cd, Ce, Dy, Fr, Eu, Gd, Ge, Hf, Hg, Ho, In, Ir, Lu, Nd, Os, Pd, Pr, Pt, Re, Rh, Ru, Sn, Sm, Ta, Tb, Te, Th Tl, Tm, U, and W.

their grains are more angular and contain more material other than quartz.

Some of the shale in the upper unit contains limy nodules and mudstone beds having an abundant fauna of gastropods, pelecypods, and goniatites. This upper shale unit also contains some thin beds of argillite and, in its uppermost 50–100 feet, some lenses of grit and of conglomerate associated with a small amount of greenish-gray quartzite. The conglomerate contains pebbles of limestone and siliceous material. Just above the quartzite there is generally a little limestone or coquina containing abundant silicified brachiopods and corals that weather out and cover the slope. Near the northwest side of the small intrusive body along Cooper Canyon, in sec. 36, limestone containing the silicified fossils is interbedded with black shale containing the pelecypod-goniatite fauna; rocks of intermediate lithology have a mixed fauna.

The uppermost 40–60 feet of the upper unit contains beds that grade into the overlying Ely Limestone. Brownish-gray sandy and bioclastic limestone beds, making up 25–50 percent of the material, alternate with beds of dark-gray shale. A few of the more limy beds contain chert nodules. The top of the Chainman is most conveniently placed at the base of the lowest thick bed of cherty limestone, a contact that can almost everywhere be located on the map within 200 feet, and usually within 50 feet. I suspect, however, that the contact would coincide more closely with the systemic boundary if it were placed at the top of the highest black shale bed, a less convenient horizon slightly higher in the section.

The trace-elements contents of 21 samples were semiquantitatively analyzed by spectrographic methods (table 5). Sample 28 contained 20 times as much zinc, chromium, vanadium, and silver as other samples, and minor anomalies were found in other samples. This cursory study showed no apparent relation between the distribution of the trace elements and the areal or stratigraphic position of the sample or the distance of the sample from mineralized areas, intrusives, or major structures.

Much of the Chainman Shale is unfossiliferous, but at a few horizons it is abundantly fossiliferous. In some places the dark-gray shale contains a rich and well-preserved fauna of pelecypods, gastropods, and goniatites, and the limier beds contain some corals and brachiopods. Collections from the shale, includ-

spectrographic analyses of black shale

not found; >, greater than; <, less than; d, detected but concentration uncertain]

35	36	37	38	39	40	41	42	43	44	45	46	47	48
61Y596 61-2959	61Y597 61-2960	61Y598 61-2961	61Y599 61-2962	61D602 61-2963	61D605 61-2964	61D608 61-2965	61D609 61-2966	61D610 61-2967	61D624 61-2968	61D626 61-2969	61D628 61-2970	61D629 61-2971	61D703 61-2975
Chainman Shale—Continued													
J. C. Hamilton—Continued [2]													
>10	7	7	>10	>10	>10	7	>10	>10	>10	>10	>10	>10	>10
3	1.5	3	3	>10	3	7	>10	7	3	7	7	3	3
.7	.7	.7	1.5	1.5	1.5	.7	1.5	3	1.5	3	7	1.5	1.5
7	3	7	7	1.5	.7	1.5	1.5	1.5	3	1.5	1.5	7	.7
>10	>10	>10	>10	.7	7	7	.3	.7	7	.3	.7	7	1.5
.15	.15	.3	.15	.3	.3	.3	.3	.3	.3	.3	.3	.3	.7
3	1.5	3	3	3	3	.7	3	3	3	3	.7	3	1.5
.3	.07	.15	.07	.7	.15	.15	.3	.3	.15	.3	.3	.15	.3
0	0	0	0	0	0	*1.5*	0	0	0	0	0	0	.7
.03	.015	.03	.03	.007	.007	.007	.015	.07	.015	.07	.3	.03	.007
0	0	0	0	.00015	0	.00015	0	0	0	0	0	0	0
.003	0	.003	.003	.007	.007	.007	.007	.007	.007	.007	.007	.003	.007
.015	.015	.03	.007	.015	.007	.007	.015	.015	.015	.03	.007	.015	.03
0	0	0	0	.0003	.00015	0	.00015	.0003	0	.00015	.00015	0	0
.0007	0	0	0	.0007	0	0	.0015	.003	0	.0015	.0007	.007	0
.015	.015	.007	.003	.015	.015	.015	.015	.015	.015	.015	.007	.003	.015
.0015	.003	.0015	.003	.0003	.003	.003	.003	.003	.003	.003	.003	.003	.003
.0003	0	.0007	.0003	.003	.0015	.0007	.003	.0015	.0007	.0015	.0015	.0007	0
0	0	0	0	.007	0	0	.003	.003	0	.003	0	0	.003
0	0	0	0	.03	0	0	.03	.03	0	0	0	0	0
0	0	0	0	0	.0007	.0007	0	.0007	0	.0015	.0007	0	0
0	0	0	0	.0015	0	0	0	0	0	0	0	0	0
.007	.003	.0015	.0015	.007	.007	.007	.007	.007	.003	.007	.003	.003	.003
0	0	0	.0015	0	0	0	.003	.003	0	.003	.003	0	0
0	0	0	0	0	0	0	0	0	0	0	*.07*	0	0
.0007	0	.0007	.0007	.003	.0015	.0007	.0015	.0015	.0007	.0015	.0015	.0007	.0007
.03	.07	.03	.07	.03	.03	.07	.03	.03	.03	.03	.03	.007	.015
.007	.007	.003	.003	.015	.007	.015	.015	.015	.007	.03	.007	.007	.007
.0015	.0015	.0015	.0015	.003	.003	.003	.003	.003	.0015	.003	.003	.0015	.003
.00015	.00015	.00015	.00015	.0003	.0003	.0003	.0003	.0003	.0003	.0003	.0003	.00015	.0003
0	0	0	0	0	d	.03	0	0	0	0	0	0	0
.015	.007	.015	.007	.015	.015	.003	.007	.015	.015	.015	.015	.007	.015

ing one, field No. 200, made in 1957 by Mackenzie Gordon, Jr., R. K. Hose, C. A. Repenning, and H. R. Christner, contains about 550 specimens; the fauna are listed in table 6. W. L. Youngquist (1949, p. 278–279) previously reported on another collection from the Connors Pass quadrangle.

The collections are all of Late Mississippian age and range from the *Goniatites granosus* zone to the upper part of the *Eumorphoceras bisulcatum* zone. Gordon thought possibly some of the basal beds of the formation are missing in this area, inasmuch as fossils of an older zone in nearby stratigraphic sections have not been found here, and the *Goniatites granosus* zone is here only about 300–500 feet above the base of the formation instead of considerably higher as it is in other areas.

The Chainman Shale occurs throughout the Schell Creek Range and adjacent ranges, and some of its lithologies reflect more distant orogenic events. The sandstone, quartzite, and conglomerate lenses of the Chainman Shale in the Connors Pass quadrangle are inferred to be the thin extremities of tongues of the Diamond Peak Formation, which was derived from the west (Nolan and others, 1956), or its correlative, the Scotty Wash Quartzite; inasmuch as some of the conglomerate contains siliceous pebbles, it also was probably derived at least in part from the west.

TABLE 6.—*Fauna of the Chainman Shale*

[Identified by Mackenzie Gordon, Jr. (written commun., 1961, 1963), Gordon and Helen Duncan (written commun., 1962)]

Faunal zone	*Goniatites granosus*			*Eumorphoceras bisulcatum*															
Unit	Lower shale and siltstone			Middle shale		Upper shale													
U.S. Geol. Survey loc.	19986–PC	19985–PC	18800–PC	19987–PC	19988–PC	19989–PC	18802–PC	18801–PC	18815–PC	--------	--------	18808–PC	18804–PC	18805–PC	18806–PC	18809–PC	18812–PC	18813–PC	17180–PC
Field No.	61D617	60D500	59D267	61D640	61D698	61D702	59D336	59D334	59D333	58D63	58D18	59D309	59D163	59D164	59D266	59D322	59D327	59D328	59D200
Brachiopods:																			
Chonetes sp.							×												
Cleiothyridina cf. *C. suborbicularis* (Hall)												×							
Cleicthyridina cf. *C. sublamellosa* (Hall)										×				×					
Composita sp. indet.										×						?			
Composita subquadrata (Hall)														×					
Diaphragmus n. sp. A									×	×									×
Flexaria aff. F. *arkansana* (Girty)									×										
Flexaria sp. indet.								×										×	
Hustedia sp.														×					
Inflatia inflata (McChesney)																			×
Inflatia n. sp. A									×	×		×	?	×	×				×
Inflatia sp. indet.											×						×	×	
Kozlowzkia? sp.																	×		
Leiorhynchus carboniferum Girty				×		?	×	×	×										×
carboniferum polypleurum Girty?																			×
Dinoproductid indet.										×	×								
Martinia? sp. indet.															×				
Orbiculoidea sp.								×											
Orthotetes kaskaskiensis (McChesney)?																			×
Ovatia sp. indet.																		×	
Petrocrania chesterensis (Miller and Faber)														×					
Punctospirifer transversus (McChesney)													?	×					
Punctospirifer? sp. indet.									×										
Reticulariina campestris (White)?																		×	
Reticulariina sp. indet.														×			×		
Rhipidomella nevadensis (Meek)																		×	×
Schizophoria cf. *S. resupinoides* (Cox)													×	?					
Schizophoria cf. *S. texana* Girty									×					?					
Schizophoria sp.																	×		
Spirifer sp.											×	×							×
Strophomenid indet.											×						×		
Werriea kaskaskiensis (McChesney)?																			×
Cephalopods:																			
Cravenoceras hesperium Miller and Furnish				×	×														
Cravenoceras merriami Youngquist								×	?										
Dolorthoceras? sp.				×															
Euloxoceras sp.								×											×
Mitorthoceras crebriliratum (Girty)	×																		
Mitorthoceras perfilosum Gordon	×																		
Neoglyphioceras cloudi utahense (Miller, Youngquist, and Nielsen)		×	×																
Neoglyphioceras subcirculare (Miller)		×																	
New ammonoid genus, n. sp.				×															
Rayonnoceras sp.	×																		×
Reticycloceras cf. *R. croneisi* Gordon								×											
Stroboceras sp.								×											
Tylonautilus cf. *T. gratiasus* (Girty)							×												

See footnote at end of table.

TABLE 6.—*Fauna of the Chainman Shale*—Continued

[Identified by Mackenzie Gordon, Jr. (written commun., 1961, 1963), Gordon and Helen Duncan (written commun., 1962)]

Faunal zone	*Goniatites granosus*			*Eumorphoceras bisulcatum*															
Unit	Lower shale and siltstone			Middle shale		Upper shale													
U.S. Geol. Survey loc	19986-PC	19985-PC	18800-PC	19987-PC	19988-PC	19989-PC	18802-PC	18801-PC	18815-PC	--------	--------	18808-PC	18804-PC	18805-PC	18806-PC	18809-PC	18812-PC	18813-PC	17180-PC
Field No	61D617	60D500	59D267	61D640	61D698	61D702	59D336	59D334	59D333	58D63	58D18	59D309	59D163	59D164	59D266	59D322	59D327	59D328	59D200
Gastropods:																			
Bellerophontid																			×
Gastropod indet (high spired)				×															
Glabrocingulum quadrigatum Sadlick and Nielsen				×				×											
Glabrocingulum binodosum Sadlick and Nielsen																			×
Mourlonia sp							×	×	×										
Naticopsis sp													×						
Trepospira sp								×											×
Pelecypods:																			
Allorisma? sp. indet														×					
Astartella sp																			×
Aviculopecten spp				×	×			×	×										×
Cypricardella? spp				×		×		×											
Nuculoides sp				×				×											
Paleoneilo sp				×		×	×												
Phestia sp				×		×	×	×											
Pelecypod indet						×	×	×											×
Polidevcia sp		×																	
Posidonia cf. *P. wapanuckensis* (Girty)				×	×			×											
Promytilus sp				×				×											
Sphenotus sp								×											
Others:																			
Archimedes sp. indet														?		×			
Fistuliporoid bryozoan																			×
Rhomboporoid bryozoan indet											×							×	
Stenoporoid bryozoan														×[1]				×	×
Trepostomatous bryozoan indet. (laminar forms)											×								
Trepostomatous bryozoan indet. (ramose forms)										×	×						×		
Amplexizaphrentis? sp												×		×					
Amplexoid? coral															×				
Horn coral																	×		
Zaphrentoid corals																			×
Crinoid columnals							×							×					
Coleolus? sp								×											
Leioclema?											×								
Pelmatozoan fragments																×	×	×	

[1] Encrusting and ramose types.

MISSISSIPPIAN, PENNSYLVANIAN, AND PERMIAN SYSTEMS

ELY LIMESTONE

The Ely Limestone, a light-gray cherty limestone about 2,500 feet thick, overlies the Chainman Shale. The contact is apparently conformable in a few places and is structurally discordant in many places. The name Ely Limestone is used here essentially as Pennebaker (1932) defined it. It consists mainly of Pennsylvanian and Permian rocks that lie between the Chainman Shale and a thick sequence of Permian sandstone, but it includes some rocks of Mississippian age.

The Ely Limestone underlies more than 15 square miles in the Connors Pass quadrangle, occurring chiefly in the southern and western parts of the quadrangle but also occupying a small area near the mouth of the canyon of Cleve Creek. It forms moderately prominent ridges that contrast strongly with the benches and valleys underlain by Chainman Shale. Small, moderately continuous, conspicuously yellowish-gray ledges and benches are typical of the lower part of the formation (fig. 8), but they are fewer and less continuous on slopes formed on the upper part.

No stratigraphic section of the Ely Limestone was measured because, like the Chainman Shale, it is much faulted and inadequately exposed. Moderately good exposures of the formation are accessible along the highway at Connors Pass; the clastic upper unit of the formation lies just east of the pass and the limy lower unit half a mile to the west of it. In both places, however, the beds are much disturbed by faults and by numerous unmapped shears. Much of

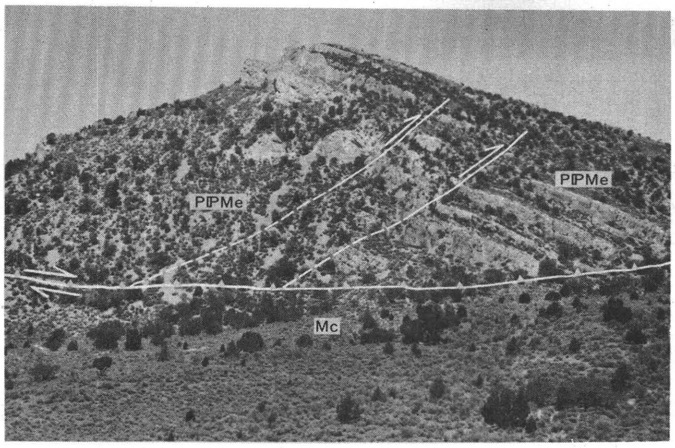


FIGURE 8.—Ledges of Ely Limestone (PIPMe) in a klippe, underlain by almost flat-lying Chainman Shale (Mc) on the south side of the peak at Taylor bench mark. The peak rises about 500 feet above the bench formed by the shale.

the lower part of the formation is fairly well exposed on the west flank of the peak on which Taylor bench mark lies and on the south slope of the ridge a mile north of The Narrows on the lower part of Steptoe Creek. Much of the crest of the ridge south of Connors Pass is underlain by the upper part of the formation, but at that place there are few exposures and there is some evidence of minor structural complications.

The Ely Limestone is divisible into two informal units, a limestone unit that makes up the lower two-thirds of the Ely, and a clastic unit that consists of the upper third. These units are not mapped separately although locally a marker bed, b, shows the base of the clastic unit. Much of the limestone is shaly, and calcareous shale increases in abundance upward. The clastic unit contains some siltstone, sandstone, dolomite, and a very small amount of conglomerate.

The base of the Ely is almost everywhere faulted; the angular discordance on the faults ranges from 0° to almost 90°, and the stratigraphic thickness of rocks that are missing along the fault may in some places be as much as 2,000 feet. Roadcuts about half a mile west of Connors Pass also show several wedges of Chainman Shale that have been squeezed up into the limestone. Obviously then, there is doubt about the completeness of all sections near the base of the formation. In some places, however, at which there is no evidence of faulting along the base, clastic limestone somewhat similar to the limestone interbedded near the top of the underlying shale occurs in the lower 20–60 feet of the Ely Limestone. This rock consists of alternating beds of light-medium-gray bioclastic limestone, brownish-gray bioclastic limestone, coquina, and limy siltstone. The lowest gray limestone beds, 4–12 inches thick, generally contain chert.

In the limestone unit, most of the limestone is relatively pure, crystalline, and fine- to medium-coarse-grained, but some of it is bioclastic or crinoidal. Toward the top of this part of the formation, moderately thick-bedded limestone in beds 20–40 feet thick alternate with equally thick units of platy limestone (fig. 8). Most of the limestone contains a little chert either in small nodules and lenses or in subspherical bodies 1–2 feet in diameter. The chert is light gray to brownish gray and weathers to a pale yellowish gray. A few units of rock 10–30 feet thick contain as much as 20 percent of lenticular and bedded chert. The cherty rocks form many of the yellowish bands that are conspicuous on some of the slopes. The subspherical nodules are most common in the ledgy units or just above them. Many are made up of concentric layers of competent and incompetent material and resemble an onion in structure, and some contain vestiges of *Chaetetes* or possibly of algae.

About 1,400 feet above the base of the formation, near the top of a layer that contains the onionlike nodules, abundant subspherical silicified colonies of *Chaetetes* 6 inches to 2 feet in diameter provide the very useful marker bed a (pl. 1). Generally there is only a single *Chaetetes*-bearing limestone bed, but in some places there are two beds 20 feet apart, and in two places (in one of which no structural complications were apparent) there are two beds at least 100 feet apart. Where *Chaetetes* colonies disappear abruptly along a bed, chert nodules resembling them in size and shape continue.

The proportion of clastic rocks increases in the upper one-third of the formation, the clastic unit, in some places to such an extent that it is difficult to identfy the contact between the Ely Limestone and the overlying Arcturus Formation. Limy siltstone and sandstone make up about 75 percent of the upper 500–800 feet of the formation, and thin lenses of limestone only 25 percent. The siltstone and sandstone, which are very pale yellowish brown to very pale orange, are friable and fractured and are rarely well exposed. The limestone lenses are moderately thick bedded and are coralline; they may possibly represent biotstromal features. Large fusulines are abundant in some of the shaly limestone, and a few beds of this rock actually fusuline coquinas or fusuline-granule arenite. The clastic unit of the formation also contains a few beds of fine- to medium-grained light-gray dolomite. In the southern part of the quadrangle, the clastic rocks and fusuline limestone appear so abruptly in the stratigraphic section as to provide another marker horizon. Along Steptoe Creek north of The Narrows, a marker bed of chert-pebble conglomerate and breccia about a foot thick lies close beneath the fusuline limestone and 500–600 feet above the *Chaetetes* marker zone.

Brachiopods and corals are moderately abundant throughout the Ely Limestone, and fusulines are very abundant at some horizons in its clastic unit (table 7).

The fauna of the Ely Limestone in this area ranges in age from Late Mississippian to Early Permian but does not include any Late Pennsylvanian fossils. In some localities the lowest few tens of feet of limestone contain Late Mississippian fossils. Rocks equivalent to the Missouri and at least parts of the Des Moines and Virgil of the midcontinent region may be missing. The conglomerate marker bed, b, at the base of the clastic unit appears to be a basal unit above an unconformity at or near the base of the Permian. An unconformity of this age is widely, but not everywhere, recognized in eastern Nevada and western Utah (Steele, 1960, pl. 4; Dott, 1955).

The distribution of clastic rocks and unconformities in the Ely rocks of the surrounding region indicates that during part of Ely time the Connors Pass quadrangle lay near tectonically active areas, and the rocks within the quadrangle record several regional tectonic events. The slight increase in shaly and silty limestone in the upper part of the limestone unit—that is, in the rocks between the two marker beds—is roughly within the rocks equivalent to the Tomera Formation of Dott (1955), 60–80 miles northeast of the Connors Pass quadrangle, and inasmuch as the abundant conglomerate in the Tomera suggests that the Antler orogenic belt of central Nevada was uplifted, the detrital material in this part of the Ely Limestone in the Connors Pass area presumably was derived from the west. During Middle and Late Pennsylvanian time eastern Nevada and western Utah, as well as the tectonic area to the west, were moderately uplifted, and the Oquirrh Basin in north-central Utah subsided rapidly. Material removed from or carried across the Connors Pass area and the eastern Nevada-western Utah uplift may have been deposited in the adjacent Oquirrh Basin. The clastic unit of the Ely Limestone and probably also the clastics in the overlying Rib Hill Sandstone seem to reflect renewed uplift of the orogenic belt.

PERMIAN SYSTEM

RIB HILL SANDSTONE

The Rib Hill Sandstone, a pale-yellow-brown sandstone about 1,000 feet thick, conformably overlies the Ely Limestone. This sandstone was first named the Rib Hill Formation by Pennebaker (1932); Steele (1960, p. 103) recommended renaming the sandstone the Riepetown Sandstone (for an abandoned town near the present site of Ruth), because the name Rib Hill had also been applied to a quartzite of Precambrian age in Wisconsin, but the name Rib Hill Sandstone is apparently more widely accepted in Nevada and it will be used here.

The Rib Hill Sandstone underlies about 2 square miles of the quadrangle in a zone that extends from the Taylor mining district to the valley southeast of Connors Pass. In small areas where it was difficult to distinguish the Rib Hill Sandstone from silty rocks in the overlying Arcturus Formation, the sandstone was generally mapped with the Arcturus.

The Rib Hill Sandstone forms gentle slopes and lowlands and is poorly exposed. The slopes are strewn with small slightly rounded plates and blocks of friable sandstone which weathers to grayish orange, pale yellow brown, or pale reddish brown, all more intense than any colors of the siltstone in the upper part of the Ely Limestone and also more intense than those of most sandstone and siltstone in the Arcturus Formation. Near the Taylor district the lower part of the Rib Hill contains more reddish-brown rock than the upper part. A part of the Rib Hill is well exposed half a mile north of Connors Pass and also in the small gullies tributary to the main valley in which there is a rough narrow road.

The base of the Rib Hill Sandstone, which is nowhere well exposed, is placed above the highest beds containing much limestone and beneath the lowest relatively pure and intensely colored sandstone. The transition between the Ely and the Rib Hill occurs in a zone that is generally less than 100 feet thick and in some places less than 10 feet thick. The transitional zone contains a little of both the shaly limestone and the lenticular bodies of thick-bedded coralline limestone common in both the underlying and overlying formations.

Most of the Rib Hill consists of very fine grained to fine-grained pure to moderately impure quartz sandstone. Dolomite beds 6–18 inches thick and sandy and shaly limestone beds are common throughout the formation but make up less than 1 percent of it. The dolomite is fine grained and medium gray but weathers light gray, much like the dolomite in the adjacent formations. Most of the sandstone is thick bedded to massive, but some of it is faintly laminated and crossbedded.

TABLE 7.—*Fauna*

[Identified by Helen Duncan and Mackenzie Gordon, Jr. (written commun., 1963), Gordon (written commun.,

Epoch	Late Mississippian							Early Pennsylvanian														Middle Pennsylvanian																	
Age	Chester							Morrow														Atoka						Des Moines											
Position in formation	Basal unit (~100 ft)							Between basal unit and marker bed a																				Between marker beds a and b												
Field No.	59D298	59D162	59D304	61D699	59D329	58D20[1]	61D737	59D268	59D269	59D286	59D307	59D308	59D315	60D431	60D561	59D323	60D568b	61D643	61D668	58D17	61D739	59D272	59D321	59D271	61D634	61D684	61D670	59D297	59D314	59D273	59D283	61D691	61D692	61D701	61D704	58D24	58D61	61D742	61D748[2]
U.S. Geol. Survey loc.	19063-PC	21234-PC	21249-PC	21125-PC	18814-PC		20593-PC	21237-PC	21238-PC	21243-PC	21250-PC	21251-PC	21253-PC	21257-PC	21259-PC	21255-PC	21260-PC	21119-PC	21120-PC		20594-PC	21240-PC	21254-PC	21239-PC	21118-PC	21122-PC	21121-PC	21246-PC	21252-PC	21241-PC		21123-PC	21124-PC	21126-PC	21127-PC			20595-PC	20597-PC
Brachiopods:																																							
Antiquatonia cf. *A. hermosana* (Girty)																																X							
aff. *A. coloradoensis* (Girty)																					X																		
sp.												X																		X									
Avoniid indet.																																					X		
Beecheria sp. indet.									X			X																											
Calliprotonia? sp.																																							
Chonetes sp. indet.							X																																X
Cleiothyridina cf. *C. orbicularis* (McChesney)									X			X																									•	X	
cf. *C. sublamellosa* (Hall)	X																																						
sp. indet.																				X				X						X									
Composita cf. *C. ozarkana* Mather												X																											
subquadrata (Hall)		X	X																		X																		
subtilita (Hall)									X																X?			X?	X?										
sp. indet.	X																		X	X?															X				
sp. (narrow form)															X																								
sp. (transverse form)															X																								
Compositoid indet.					X?																															X	X		
Condrathyris sp. indet.																								X															
Crurithyris sp. indet.																																						X?	
Derbyia aff. *D. robusta*																			X																				
sp. indet.												X																							X				
Desmoinesia aff. *D. muricatina* (Dunbar and Condra)																																						X	X
Diaphragmus n. sp. B	X				X		X																																
Dictyoclostid indet.			X																					X												X			
Echinoconchus sp. indet.																																				X?			
Flexaria cf. *F. arkansana* (Girty)					X																																		
sp.		X					X		X	X?		X																											
Grandaurispina sp.																																							
Hustedia sp.												X																											
Hystriculina sp. indet.																						X?			X										X?				
Inflatia n. sp. A							X																																
sp.			X																																				
Kozlowskia nuda (Lane)																																X							
n. sp.	X		X																																	X			
Linoproductus prattenianus (Norwood and Pratten)																												X	X?	X?		X?							
sp. indet.								X	X			X		X	X				X?	X																		X	
Lissochonetes sp.																		X																					
Marginiferid indet.																				X																			
Mesolobus mesolobus decipiens (Girty)																												X											
Neospirifer sp. indet.																									X?				X							X			
cf. *N. dunbari* King																																X							
Orbiculoidea sp.					X				X																X														
Ovatia n. sp. A		X	X?																																				
Productoid indet.										X										X												X							
Reticulariina campestris (White)	X			X	X					X?		X						X		X?	X																		
sp. indet.							X																																
Rhipidomella nevadensis (Meek)			X																																				
sp.												X																							X				
Rhynchonelloid brachiopod indet.																									X														
Rhynchopora sp.																																						X	
Rugoclostus sp. indet.										X?		X?			X																								
Schizophoria cf. *S. resupinoides* (Cox)																					X																		
Schizophoria cf. *S. texana* Girty	X		X												X				X																				
sp. indet.					X					X?																													
Spirifer aff. *S. increbescens* (Hall)							X																																
n. sp.																					X																		
occiduus Sadlick												X		X	X		X	X	X?	X																			
opimus Hall, var.																					X																		
sp.		X		X	X																														X				
Spiriferoid indet.																					X															X			
Spiriferinid (fragment)																																							
Strophomenid indet.					X				X																														
Torynifer sp. indet.							X																																

See footnotes at end of table.

of the Ely Limestone

1963) Gordon and Duncan (written commun., 1962, 1963), and R. C. Douglass (written commun,. 1959, 1961, 1962)]

Epoch	Unconformity	Early Permian																										
Age		Wolfcamp																										
Position in formation		Above marker bed b																										
Field No.		61D751	61D751	59D242	59D241	59D274	59D295	59D303	59D337	60D568a	59D341	58D60	61D748[2]	61D757	61D758	61D755	61D746	61D752	61D745	61D688	61D744	61D705	61D686	61D687	61D690	61D693	61D694	61D750
U.S. Geol. Survey loc.		20599-PC	f22543	19065-PC	19066-PC	21242-PC	21245-PC	21248-PC	21256-PC	21117-PC	19062-PC	f21681	f22541	f22548	f22549	f22546	f22540	f22544	f22539	19745-PC	f22538	f21965	f21958	f21959	f21960	f21961	19746-PC	f22542
Brachiopods:																												
Antiquatonia cf. *A. hermosana* (Girty)																												
aff. *A. coloradoensis* (Girty)																												
sp.																												
Avoniid indet.																												
Beecheria sp. indet.																												
Calliprotonia? sp.								X?																				
Chonetes sp. indet.							X																					
Cleiothyridina cf. *C. orbicularis* (McChesney)																												
cf. *C. sublamellosa* (Hall)																												
sp. indet.							X		X																			
Composita cf. *C. ozarkana* Mather																												
subquadrata (Hall)																												
subtilita (Hall)																												
sp. indet.									X																			
sp. (narrow form)																												
sp. (transverse form)																												
Compositoid indet.																												
Condrathyris sp. indet.																												
Crurithyris sp. indet.				X																								
Derbyia aff. *D. robusta*																												
sp. ident			X																									
Desmoinesia aff. *D. muricatina* (Dunbar and Condra)				X																								
Diaphragmus n. sp. B																												
Dictyoclostid indet.							X																					
Echinoconchus sp. indet.																												
Flexaria cf. *F. arkansana* (Girty)																												
sp.																												
Grandaurispina sp.						X	X																					
Hustedia sp.																												
Hystriculina sp. indet.																												
Inflatia n. sp A																												
sp.																												
Kozlowskia nuda (Lane)																												
n. sp.							X		X																			
Linoproductus prattenianus (Norwood and Pratten)																												
sp. indet.										X																		
Lissochonetes sp.																												
Marginiferid indet.																												
Mesolobus mesolobus decipiens (Girty)																												
Neospirifer sp. indet.																												
cf. *N. dunbari* King																												
Orbiculoidea sp.																												
Ovatia n. sp A																												
Productoid indet.																												
Reticulariina campestris (White)																												
sp. indet.																												
Rhipidomella nevadensis (Meek)																												
sp.									X?																			
Rhynchonelloid brachiopod indet.																												
Rhynchopora sp.																												
Rugoclostus sp. indet.																												
Schizophoria cf. *S. resupinoides* (Cox)																												
Schizophoria cf. *S. texana* Girty																												
sp. indet.																												
Spirifer aff. *S. increbescens* (Hall)																												
n. sp.																												
occiduus Sadlick										X																		
opimus Hall, var.																												
sp.																												
Spiriferoid indet.									X																			
Spiriferinid (fragment)									X																			
Strophomenid indet.																												
Torynifer sp. indet.																												

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TABLE 7.—*Fauna of the*

Epoch	Late Mississippian							Early Pennsylvanian														Middle Pennsylvanian																	
Age	Chester							Morrow														Atoka						Des Moines											
Position in formation	Basal unit (~100 ft)							Between basal unit and marker bed a																		Between marker beds a and b													
Field No	59D298	59D162	59D304	61D699	59D329	58D20[1]	61D737	59D268	59D269	59D286	59D307	59D308	59D315	60D431	60D561	59D323	60D568b	61D643	61D668	58D17	61D739	59D272	59D321	59D271	61D634	61D684	61D670	59D297	59D314	59D273	59D283	61D691	61D692	61D701	61D704	58D24	58D61	61D742	61D748[2]
U.S. Geol. Survey loc	19063-PC	21234-PC	21249-PC	21125-PC	18814-PC		20593-PC	21237-PC	21238-PC	21243-PC	21250-PC	21251-PC	21253-PC	21257-PC	21259-PC	21255-PC	21260-PC	21119-PC	21120-PC		20594-PC	21240-PC	21254-PC	21239-PC	21118-PC	21122-PC	21121-PC	21246-PC	21252-PC	21241-PC		21123-PC	21124-PC	21126-PC	21127-PC			20595-PC	20597-PC
Bryoza:																																							
Archimedes sp				X																																			
Cystodictya sp																									X														
Fenestella sp																																X							
Fenestellid bryozoan																									X														
Fistuliporoid brozoan																				X									X?										
Penniretipora sp																													X			X							
Polypora sp																			X					X								X		X					
Rhomboporoid bryozoan indet																			X						X							X							
Stenoporid bryozoan										X											X																		
Stenoporid bryozoan (encrusting form)															X		X		X													X							
Stenoporid bryozoan (large ramose form)	X																																						
Stenoporid bryozoan (small ramose form)																													X										
Trepostomatous bryozoan indet. (ramose form).	X				X														X										X										
Trepostomatous bryozoan indet																																							
Corals:																																							
Amplexizaphrentis? sp			X																																				
Auloporid coral																																X							
Caninoid coral sp. indet										X																				X		X			X				
Caninia trojana Easton?																																		X					
Chaetetes sp																						X	X		X	X													
Diphyphyllum connorsensis Easton																																							
Horn coral indet												X												X					X										
Kleopatrina ftatateeta (McCutcheon and Wilson)																																							
Lithostrotionella sp																																							
dilitata Easton?																																							
"*Lonsdaleia*" cf. *L. illipahensis* Easton																																							
Multithecopora sp. *A*								X			X		X				X?																						
sp. B																											X?	X				X?			X?				
Orygmophyllum? sp											X						X?	X																					
Syringopora sp																																		X					
mccutcheonae Wilson and Langenheim																																							
Thysanophyllum princeps Easton																																							
sp. indet																																							
Foraminifera:																																							
Climacammina sp																																							
Endothyrid																																							
Pseudofusulina sp																																							
Pseudoschwagerina sp																																							
Schubertella sp																																							
schubertellid																																							
Schwagerina sp																																							
Textularid																																							
Triticites sp																																							
Others:																																							
Ameura sp. (pygidium)																																		X					
Astartella sp. indet																		X?																					
Limipecten sp																																							X
Peruvispira sp																																							
Bellerophontacean gastropod, indet							X																																
Gastropod indet							X																														X		
Trilobite, indet																X														X									
Crinoid plate																													X										
Crinoid columnals										X																													
Pelmatozoan debris	X																X	X	X																		X		
Ostracodes																																					X		

[1] From same locality as field No. 61D737.

[2] Mixed collection, specimens largely from float from rocks in stratigraphic interval of about 50 feet near hilltop on inferred fault slice of Ely Limestone.

Ely Limestone—Continued

Epoch	Unconformity	Early Permian																										
Age		Wolfcamp																										
Position in formation		Above marker bed b																										
Field No		61D751		59D242	59D241	59D274	59D295	59D303	59D337	60D568a	59D341	58D60	61D748²	61D757	61D758	61D755	61D746	61D752	61D745	61D688	61D744	61D705	61D686	61D687	61D690	61D693	61D694	61D750
U.S. Geol. Survey loc		20599-PC	f22543	19065-PC	19066-PC	21242-PC	21245-PC	21248-PC	21256-PC	21117-PC	19062-PC	f21681	f22541	f22548	f22549	f22546	f22540	f22544	f22539	19745-PC	f22538	f21965	f21958	f21959	f21960	f21961	19746-PC	f22542
Bryozoa:																												
Archimedes sp																												
Cystodictya sp																												
Fenestella sp									×																			
Fenestellid bryozoan																												
Fistuliporoid brozoan																												
Penniretipora sp																												
Polypora sp																												
Rhomboporoid bryozoan indet							×		×																			
Stenoporid bryozoan									×																			
Stenoporid bryozoan (encrusting form)										×																		
Stenoporid bryozoan (large ramose form)																												
Stenoporid bryozoan (small ramose form)																												
Trepostomatous bryozoan indet. (ramose form)																												
Trepostomatous bryozoan indet									×																			
Corals:																												
Amplexizaphrentis? sp																												
Auloporid coral							×																					
Caninoid coral sp. indet																												
Caninia trojana Easton?																												
Chaetetes sp																												
Diphyphyllum connorsensis Easton		×		×?																								
Horn coral indet									×																			
Kleopatrina ftatateet a (McCutcheon and Wilson)								×			cf.																	
Lithostrotionella sp																											×	
dilitata Easton?										×																		
"*Lonsdaleia*" cf. *L. illipahensis* Easton											×																	
Multithecopora sp. *A*																												
sp. B																												
Orygmophyllum? sp																												
Syringopora sp																												
mccutcheonae Wilson and Langenhiem					×			×?												×							×	
Thysanophyllum princeps Easton					×					×										×								
sp. indet				×?																								
Foraminifera:																												
Climacammina sp																						×	×	×		×		
Endothyrid																×												
Pseudofusulina sp			×										×	×	×			×		×	×		×?					×
Pseudoschwagerina sp			×?																			×		×		×?		
Schubertella sp													×			×	×			×			×	×				×
Schubertellid				×													×	×										
Schwagerina sp																×	×						×		×			
Textularid			×										×	×			×	×										×
Triticites sp												×										×?				×		
Others:																												
Ameura sp. (pygidium)																												
Astartella sp. indet																												
Limipecten sp																												
Peruvispira sp							×																					
Bellerophontacean gastropod, indet																												
Gastropod indet																												
Trilobite, indet																												
Crinoid plate																												
Crinoid columnals							×	×																				
Pelmatozoan debris																												
Ostracodes																												

The sandstone is 60–95 percent quartz grains and 1–2 percent heavy minerals; the remainder is clay material, calcite cement, and a little silica cement. The heavy minerals include, in order of decreasing abundance, tourmaline, leucoxene, zircon, rutile(?), and iron oxide, a suite resembling in type and abundance those in all sandy rocks above the sandstone marker in member b of the Guilmette Formation. Sericite and possibly chert are also present in very small quantity. The average grain size of much of the rock is 0.06 mm; most of the small grains are angular, but the larger grains of quartz and the grains of heavy minerals are more or less rounded. The quartz grains are mostly well sorted and many are frosted. The calcite cement is irregularly distributed; so many of the quartz grains interlock.

Fossils are scarce in the Rib Hill Sandstone, but a few were collected from limestone within 100 feet of the base (table 8). Some of the fusulines were identified by R. C. Douglass as *Schwagerina*(?) sp., of a type thus far reported only from the basal beds of the Riepetown Sandstone (here called the Rib Hill Sandstone) of D. R. Shawe's section A (written commun., 1961) in the Egan Range. Both at that locality and in the Connors Pass quadrangle the enclosing rocks contain fossils of Early Permian age, or Wolfcamp age as the term is used in west Texas; thus, the Rib Hill Sandstone is probably of Wolfcamp age.

TABLE 8.—*Fauna of lowermost 100 feet of the Rib Hill Sandstone*

[Identified by R. C. Douglass (written commun., 1961, 1962), Mackenzie Gordon, Jr., and Helen Duncan (written commun., 1963)]

Age	Wolfcamp			
Field No., abbreviated	280	743	632	700
U.S. Geol. Survey loc	19064	f22537	f21957	f21964
Crurithyris sp	×			
Trepostomatous bryozoan (small ramose form)	×			
Fusuline indet	×			
Monodiexodina sp		×		
Pseudoschwagerina sp		×		×
Schubertella sp		×		
Schwagerina sp			×?	
Climacammina				×

The area in which the Rib Hill Sandstone has been mapped is limited to part of White Pine County, Nev., but Steele (1960, p. 103) recognized the lithology farther to the north and west. The Rib Hill Sandstone seems to be a local highly clastic facies near the base of a group of rocks of a mixed lithology similar to that of the Arcturus Formation, and as it increases in limestone and dolomite content the Rib Hill loses its lithologic identity. To the west the correlative of the Rib Hill Sandstone is the lower part of the Carbon Ridge Formation of Nolan, Merriam, and Williams (1956), which contains considerable conglomerate as well as more limestone than the Rib Hill. To the east the Rib Hill Sandstone loses its lithologic identity near the Utah-Nevada border. In the Confusion Range it is probably represented by part of the Arcturus Formation of Hose and Repenning (1959), and at Gold Hill by a small part of the Oquirrh Formation of Nolan (1935).

ARCTURUS FORMATION

The Arcturus Formation—several thousand feet of heterogeneous rock, containing much limestone, dolomite, shale, and sandstone, and locally much conglomerate and a little gypsum—conformably overlies the Rib Hill Sandstone. These are the youngest Paleozoic rocks in the quadrangle. The name Arcturus Formation was first applied to them in the Egan Range by Lawson (1906, p. 293), but the formation was redefined by Spencer (1917, p. 28) and given a type locality near the Arcturus claims in the Robinson mining district just west of Ely. The name is applied here, as it was by Steele (1960, p. 103), to the rocks above his Riepetown Sandstone (here called the Rib Hill Sandstone), and below the Kaibab Limestone (of Leonard age). The Arcturus Formation in the Connors Pass quadrangle differs from that of the type locality, just west of Ely, chiefly in probably containing a tongue of conglomerate many hundreds of feet thick in its upper part. The top of the formation is not exposed within the quadrangle.

The Arcturus Formation underlies more than 4 square miles in the southern part of the quadrangle, chiefly in the southern part of the drainage basin of Cooper Canyon and in the drainage basin southeast of Connors Pass. The formation is in general so poorly exposed that its internal structure is obscure and its thickness therefore hard to determine. No subdivision that is widespread within the Connors Pass quadrangle has been made, but locally a conglomerate member has been mapped. A basal limestone unit is also referred to in the following description, but most of the formation is described lithologically rather than stratigraphically. Rocks characteristic of the formation are well exposed along both the old and the new alinements of U.S. Highway 6–50–93. Most of these rocks form gentle slopes and low hills, but the limestone near the base forms low ridges and narrow ledges.

Outcrops of Arcturus rocks other than this limestone unit are scarce and usually consist of limestone, dolomite, or conglomerate.

LITHOLOGY

A basal limestone unit of the Arcturus forms lenses as much as 400 feet thick. It consists of moderately thick-bedded coralline limestone, shaly limestone, and interbedded siltstone and sandstone. This unit is thickest about a mile southeast of Connors Pass, but it is virtually absent half a mile farther south, where there is no evidence of bedding-plane faults. The individual limestone beds or groups of beds are lenticular; many of them are about 50 feet thick and 300 feet long and resemble biostromal reefs. The limestone is fine grained to moderately coarse grained, and much of it is bioclastic or crinoidal. Fossil fragments are common, and small silicified spines, possibly of echinoids, are conspicuous on many weathered surfaces. Light-gray chert, which weathers white to pale yellowish brown or yellowish gray is widespread but rarely abundant. It assumes a large variety of shapes, such as concentrically layered subspherical bodies, amoeboid nodules 6–12 inches in diameter, lenses parallel to bedding, and small pelletlike bodies. Locally, groups of limestone beds 5–20 feet thick contain as much as 75 percent chert, in cores surrounded by siliceous siltstone or limestone, but the interbedded siltstone and sandstone resembles that in the main body of the formation. In many ways, then, the basal part of the Arcturus Formation resembles the upper part of the Ely Limestone, but the two can be distinguished by their stratigraphic relation to the Rib Hill Sandstone and generally by their fossils. Where the basal limestone of the Arcturus is absent or is thin bedded, the base of the formation grades into the underlying Rib Hill Sandstone.

The main part of the Arcturus is much more heterogeneous than the basal limestone unit; within the quadrangle it is estimated to be at least 1,500 feet thick and possibly much thicker. It consists of a mixture of sandstone, limestone, dolomite, and shale, which are mostly very pale yellowish brown, though some are white or pale red. The term "marl" may be appropriate for much of the formation. In the exposures along the highway east of Connors Pass the beds of the main part of the formation are 2–12 inches thick, but elsewhere the beds vary more in thickness.

Sandstone typical of the formation contains 65–85 percent of very well sorted quartz grains, appreciable amounts of detrital tourmaline, leucoxene, zircon, rutile(?), plagioclase, and some calcite that may be detrital. The grains are cemented with calcite, clay minerals, iron oxide, and a little silica, distributed somewhat irregularly. The grains range from very fine sand to fine sand; those surrounded by cementing material are subrounded to rounded, whereas the others are angular and form a compact polygonal mosaic without interpentrating boundaries. Such a packing suggests that the close mutual accommodation of grains was obtained by the solution of parts of the grains, perhaps along the pressure surfaces between the grains. Inasmuch as overgrowths are relatively scarce, silica must have been removed from parts of the rock; indeed, the general friability of the rock indicates that silica must have been removed from much of the unit. Yet only a moderate amount of silica appears to have been removed in this manner, for there is no obvious difference in size between the grains forming the mosaics and those surrounded by cementing material. In the rocks that have a mosaic texture, the shapes of the grains and possibly also their size characteristics are diagenetic features, and the diagenetic process may have ceased when the polygonal mosaic was completed and the pores were filled.

The limestone in the main part of the formation is similar to that in the basal limestone unit except that it occurs in thinner and more scattered lenses. Corals and echinoid(?) and crinoid debris are common, and fusulines occur in many of the shaly limestone beds.

Fine-grained light-gray to pale-yellowish-brown dolomite, similar to that in the Rib Hill Sandstone and the top of the Ely Limestone, forms a few thin widely separated lenses; this rock seems more abundant than it really is because, being especially competent, it forms ledges.

Shale is probably also common, but it is very rarely exposed. Near the conglomerate lens north of the highway, shale makes up about a quarter of a poorly exposed section about 400 feet thick. Some of the weathered shale beneath the conglomerate is veneered by a powdery soil, which is capped by a brown crust resembling that formed on gypsiferous ground; presumably this shale is here gypsiferous. There is no conspicuous gypsiferous zone in this area, however, as there is in the Egan Range southwest of Ely. In this part of the Schell Creek Range, more gypsum may have once been present in the upper parts of the Arcturus that presumably were cut out by the low-angle fault above the Arcturus northeast of Connors Pass.

CONGLOMERATE MEMBER

In the upper part of the Arcturus Formation along the highway east of Connors Pass, there is a reddish-brown conglomerate that is probably more than 500 feet thick at that place but wedges out rapidly to the northwest. The thickness body of conglomerate lies almost entirely south of the highway at the big bend 1–2 miles east of the pass, where it seems to be faulted down into other rocks of the Arcturus Formation. A thin conglomerate bed is also exposed half a mile northwest of the bend. These beds of conglomerate on both sides of the highway are here referred to as the conglomerate member of the Arcturus Formation. Much of the conglomerate underlies gentle terrain and weathers to small knobs and large boulders, but the thin lenses of this rock are more friable and leave only residual pebbles in the soil.

As a whole, the conglomerate member is interbedded with other rocks in the Arcturus Formation, and in some places an unconformity underlies the member. Between the old and new alinements of the highway, for a distance of about 1,000 feet, the base of the conglomerate cuts gently across an underlying limestone bed that dips about 20° more steeply than the conglomerate. The exposures here are inadequate for one to determine whether the unconformity represents local channeling at the base of the conglomerate or is of greater structural significance.

The conglomerate is a thick-bedded moderate-red to light-brown rock consisting of chert and quartzite pebbles in a sandy slightly calcareous matrix. The pebbles are subrounded to rounded and have smooth to polished surfaces. North of the highway the member also contains some grit and granule conglomerate and conglomeratic sandstone. Some fossiliferous Permian limestone and sandstone are apparently interbedded in the main body of conglomerate, but the stratigraphic relations are partly obscured by local faults of unknown throw.

Very little of the upper contact of the conglomerate member is exposed. In a roadcut at the big bend in Highway 6–50–93, younger tuffaceous shale is inferred to be faulted onto the conglomerate, and some blocks of conglomerate are faulted or slumped onto the shale. Northwest of the bid bend in the highway, thin lenses of conglomerate are overlain with apparent conformity by fossiliferous Permian limestone.

FAUNA, AGE, AND CORRELATION

Corals, brachiopods, and fusulines are moderately abundant in some of the limestone beds of the Arcturus, and a few beds contain many gastropods and pelecypods. Fossil fragments abound in all the carbonate rocks, and echinoid(?) spines are conspicuous on many weathered surfaces. All the identified fossils are listed in table 9.

TABLE 9.—*Fauna of the Arcturus Formation*

[Foraminifera identified by R. C. Douglass (written commun., 1959, 1961); corals by Helen Duncan (written commun., 1963); most brachiopods by Mackenzie Gordon, Jr. (Duncan and Gordon, written commun., 1963), some by J. T. Dutro, Jr. (written commun., 1960); gastropods, pelecypods, and a few brachipods by E. L. Yochelson (written commun., 1960)]

Age	Wolfcamp					Wolfcamp or Lenard			
Type of fauna	Mixed [1]	Arcturus [2]							
Approximate position in formation where known (feet above base)	600	----	50	----	----	----	1,000	1,000	----
Field No.	61D588	61D662	61D747	58D23	59D276	61D696	61D586	61D585	59D244
U.S. Geol. Survey loc.	19739–PC	19740–PC	20598–PC	f 21680 20312–PC	18477–PC	f 21962	19748–PC	19747–PC	19067–PC
Brachiopods:									
Aulostegid brachiopod, n. gen., n. sp.	----	----	----	----	×	----	----	----	----
Avoniid productoid indet.	----	----	----	----	----	----	----	×	----
Composita subtilita (Hall)	----	×	----	----	----	----	----	----	----
Heteralosia sp.	----	----	----	----	×	----	----	----	----
Neophricodothyris?	----	----	----	----	×	----	----	----	----
Spiriferinid indet.	----	×	----	----	----	----	----	----	----
Bryozoans:									
Fistuliporoid bryozoan (lamellar growth form)	----	×	----	----	----	----	----	----	----
Stenoporoid bryozoan (lamellar growth form)	----	×	----	----	----	----	----	----	----
Tabulipora arcturusensis Gilmore?	×	----	----	----	----	----	----	----	----
Corals:									
Heritschioides sp.	----	----	×	----	----	----	----	----	----
Kleopatrina ftatateeta (McCutcheon and Wilson)	×	----	----	----	----	----	----	----	----
Lithostrotionella dilitata Easton?	----	----	----	×	----	----	----	----	----
Lophamplexus? sp. indet.	----	×	----	----	----	----	----	----	----
Roemeripora? sp.	----	----	----	×	----	----	----	----	----
Foraminifera:									
Bradyina sp.	----	----	----	----	----	×	----	----	----
Climacammina sp.	----	----	----	----	----	×	----	----	----
Schubertella sp.	----	----	----	----	----	×	----	----	----
Millerellids	----	----	----	×	----	----	----	----	----
Schwagerina sp.	----	----	----	×	----	×	----	----	----
Gastropods:									
Amphiscapha sp. indet.	----	----	----	----	----	----	----	×	----
Bellerophon sp. indet.	----	----	----	----	----	----	----	×	----
Euphemites sp. indet.	----	----	----	----	----	----	----	×?	×
Gastropod indet. (moderately high spired form)	----	----	----	----	----	----	----	×	----
Glyptospira sp. indet.	----	----	----	----	----	----	----	×	----
Knightites (*Retispira*) sp. indet.	----	----	----	----	----	----	----	×	×?
Omphalotrochus cf. *O. whitneyi* (Meek)	----	----	----	----	----	----	×	----	----
Permophorus? sp. indet.	----	----	----	----	----	----	----	×	----
Pleurotomaraceans indet. (two genera)	----	----	----	----	----	----	----	×	----
Pseudozygopleurid indet.	----	----	----	----	----	----	----	×	----
Taosia? sp. indet.	----	----	----	----	----	----	----	×	----
Others:									
Echinoid spines or plates	----	----	×	----	----	----	----	----	----
Pectinoid pelecypod	----	----	----	----	----	----	----	----	×
Paleonucula sp. indet.	----	----	----	----	----	----	----	----	×
Plagioglypta sp.	----	----	----	----	----	----	----	×	----
Scaphopod indet.	----	----	----	----	----	----	----	----	×

[1] *Kleopatrina* supposedly is a late Ely fossil, the *Tabulipora* is questionably identified with an Arcturus species. They were collected above rocks mapped as the Rib Hill Sandstone.

[2] These collections either require or permit a correlation with an Arcturus fauna.

The Arcturus Formation is identified as Permian on the basis of fossils. Its fauna in the Connors Pass quadrangle is comparable to that of the Wolfcamp Series and perhaps also of a part of the Leonard Series in the West Texas region. The oldest rocks lie well above the base of the Wolfcamp, but, so far as I can judge from the genera identified, the local fauna throughout that stage is uniform.

The age of the conglomerate member here considered to be part of the Arcturus Formation, has been the subject for considerable speculation by geologists who have seen the rock along the highway, and, indeed, the assigned Permian age is tentative. P. J. Barosh (oral commun., 1963), for instance, considered the conglomerate to be Tertiary because he placed less emphasis on the significance of the fault separating the conglomerate from the overlying shale in the roadcut and more emphasis on the significance of the basal unconformity. A Permian age is preferred to a younger age, however, because northwest of the big bend in the highway some conglomerate lenses seem to be interbedded in fossiliferous Permian limestone and because south of that bend in the highway some fossiliferous Permian limestone appears to be interbedded in the main body of the conglomerate.

The name Arcturus Formation is used only in east-central Nevada and adjacent parts of Utah. The Arcturus Formation of the Connors Pass quadrangle is represented in the Confusion Range, about 40 miles to the east, by the middle and upper part of the Arcturus Formation described by Hose and Repenning (1959). Steele (1960) advocated the use of the name Pequop Formation for correlative rocks at Moorman Ranch between Ely and Eureka, Nev., and also north of the Connors Pass. The Arcturus is also correlative with part of the Oquirrh Formation (Nolan, 1935) at Gold Hill, Utah.

The lithology of the Arcturus Formation in the Connors Pass quadrangle suggests deposition in a shelf area, and comparison with correlative rocks in nearby areas suggests that the shelf lay slightly south of the axis of the basin. The Arcturus within the quadrangle includes at least a small amount of gypsum, which occurs in greater abundance in contemporary rocks (Hose and Repenning, 1959) in the Confusion Range, in part of the Egan Range (A. L. Brokaw, oral commun., 1962), and in the southern part of the Butte Mountains (Douglass, 1960). The formation also includes some conglomerate, similar to that found by Nolan, Merriam, and Williams (1956) in the southern Sulphur Springs Range, by Dott (1955) in the northern Diamond Range about 75 miles to the northwest, by Sharp (1942) and Dott (1955) in the southern Ruby Range about 60 miles to the northwest, by Snelson (1955) and Harlow (1956) in the Spruce Mountain area about 85 miles to the north, and by Nolan (1935) in the "central facies" of the Oquirrh Formation at Gold Hill, Utah. The position of the conglomerate member in the southeastern part of the quadrangle and its wedging out to the northwest suggest that the pebbles came from the south or southeast; however, inasmuch as only a little conglomerate is found in the area, there is no assurance that the source lay in that direction. The composition and the rounding of the pebbles, considered together, suggest that they were transported a long distance. During Arcturus time the Connors Pass quadrangle probably lay in a shelf area along the southeastern margin of a marine basin, a basin that was nearly or entirely landlocked part of the time.

JURASSIC, CRETACEOUS, AND TERTIARY(?) SYSTEMS

Rocks of Mesozoic and earliest Tertiary age are extremely scarce in the quadrangle, as they are in most of the region. Triassic and Lower Jurassic sedimentary rocks are absent in the quadrangle, but they probably once covered the area and may have affected later deformation through their great load. Several monzonitic plugs were intruded into rocks of the adjacent ranges, 10–20 miles distant, during Mesozoic and early Tertiary time. In the Connors Pass quadrangle there are some porphyritic rhyolite dikes that are of Mesozoic or Tertiary age, and a small fault block of shale and tuffaceous sandstone that is probably of Cretaceous or Paleocene age.

PORPHYRITIC RHYOLITE

There are about 30 small rhyolite dikes in the Connors Pass quadrangle. Most are clustered within a few square miles in the Taylor mining district, but three are east of Connors Pass and one is in Cooper Wash 2.5 miles north of the Taylor district. An unexposed dike may lie half a mile to a mile east of the one in Cooper Wash, for the colluvium there contains scattered fragments of a prophyritic rhyolite similar to that of the dikes.

Most of the dikes are only a few feet wide and less than 200 feet long. Several are irregular in plan but most are tabular bodies. The longest dike extends for a mile northwestward from Majors Place and is uniformly 10 feet wide. In the Taylor district there are several dikes, one too small to be mapped, that parallel nearby normal faults and are less sheared than some of the host rocks.

The rocks of most of the dikes are incompetent and are poorly exposed in narrow belts largely covered with platy and blocky rubble. Several dikes near the south end of the Taylor district are better exposed along small ridges, but the northwest end of the dike at Majors Place is marked by a band of white powdery material containing only a few fragments of porphyritic rock.

Where it is least altered the rhyolite is very light gray to very pale orange, but where it is much altered it is either white, pale red, or moderate reddish orange. The rhyolite usually has an earthy or chalky appearance, but in a few dikes it is subvitreous and in the southeast end of the dike near Majors Place it is slightly saccharoidal. Quartz veinlets cut the dike and the slaty host rocks near Majors Place. Quartz phenocrysts are in the rhyolite, and grains of biotite and feldspar can be seen where the rock is not much altered.

Thin sections show that the rhyolite is all porphyritic, having a granular to obscure groundmass dominated by secondary minerals. Phenocrysts generally make up only about 5 percent of the rock and never more than 20 percent. Quartz phenocrysts constitute 1–6 percent, and plagioclase phenocrysts 1–5 percent; the phenocrysts also commonly include a trace to 8 percent of potassium feldspar and as much as 2 percent of biotite. The anorthite content of the plagioclase is An_{0-35}, but it generally cannot be determined because of the intensive alteration. In some specimens the potassium feldspar is sanidine, but in about half it consists of an undetermined potassium feldspar that is intergrown with quartz. Accessory magnetite(?), apatite, and zircon were observed in some specimens, and secondary quartz, sericite, clay minerals, and iron oxide are generally present. In some specimens the plagioclase is partially replaced by albite and sericite; in others it is alomst entirely replaced by sericite, kaolin(?), and calcite. Some of the silica in the groundmass may be chalcedony.

The chemical compositions of two samples of porphyritic rhyolite are listed in table 10.

The porphyritic rhyolite dikes contain no glass and probably contained none prior to alteration; many are granophyric. These dikes, therefore, probably cooled more slowly and at greater depth than those associated with the volcanic rocks. They may be roughly contemporaneous with nearby stocks, as is apparently true of similar porphyritic rhyolite dikes in the southern Snake Range (Drewes, 1958) and in the Ward district of the Egan Range (A. L. Brokaw, oral commun., 1961).

No measure of the absolute age of the dikes is available, nor is their relative age closely defined. The dikes intrude rocks as young as the Arcturus Formation, of Permian age, and fragments similar to the dike rocks—though probably derived from the Snake Range, as indicated by other components associated with these fragments—are included in the Eocene conglomerate west of Majors Place. The dikes are probably also older than some of the Tertiary volcanic rocks, because they were intruded at a depth greater than that at which the glassy volcanic plugs were emplaced, during a period when the region was steadily being exhumed. If the correlation of the porphyritic dikes with the stocks in the adjacent ranges is accepted, the possible range in age may be further narrowed. Radiometric dates of some of the stocks described by Griggs, Whitebread, and Roberts (in Adair and Stringham, 1960, p. 231) range from

TABLE 10.—*Chemical and semiquantitative spectrographic analyses of porphyritic rhyolite*

[Elements looked for in spectrographic analyses but not found: Ag, As, Au, Bi, Cd, Ge, Hf, Hg, In, Ir, Mo, Os, Pd, Pt, Re, Rh, Ru, Sb, Ta, Te, Th, Tl, U, W, and Zn]

Sample loc. (pl. 1)	1	2
Lab. No.	H3186	H3187
Chemical analyses [By C. L. Parker]		
SiO_2	65.54	76.26
Al_2O_3	11.73	12.61
Fe_2O_3	.40	.25
FeO	.59	.44
MgO	.45	.22
CaO	6.83	.71
Na_2O	2.47	3.03
K_2O	4.19	4.79
H_2O+	.85	.52
H_2O-	1.37	.68
TiO_2	.05	.05
P_2O_5	.01	.01
MnO	.19	.05
CO_2	5.06	.04
Cl	.01	.01
F	.12	.10
Subtotal	99.86	99.77
Less O	.05	.04
Total	99.81	99.73
Semiquantitative spectrographic analyses [By P. R. Barnett]		
B	0	0
Ba	.03	.015
Be	.0003	.0003
Ce	0	0
Co	0	0
Cr	0	.00015
Cu	0	.00015
Ga	.0015	.0015
La	0	0
Li	.015	.015
Nb	.0015	.003
Nd	0	0
Ni	.0003	0
Pb	.003	.0015
Sc	.0007	.0007
Sn	.0007	.0007
Sr	.015	.003
V	0	0
Y	.007	.0015
Yb	.0007	.00015
Zr	.003	.003

Permian to Oligocene, and inasmuch as the dikes cut Permian rocks they are probably of Jurassic to Oligocene age.

CRETACEOUS OR TERTIARY SYSTEM

SHALE AND TUFFACEOUS SANDSTONE

A sequence about 50 feet thick, made up of two layers of shale separated by a layer of tuffaceous sandstone, underlies a very small area along U.S. Highway 6–50–93 at the big bend 1.5 miles east of Connors Pass. The rocks are exposed in only one cut, slightly more than 200 feet long and less than 25 feet high, on the north side of the road. If that cut had not been made, these rocks would not have been found, for the surface underlain by them is thickly covered with rubble from the adjacent Arcturus Formation. The rocks seem to lie in a thin fault block between outcrops of the conglomerate member of the Arcturus and outcrops of other Arcturus rocks. Dark-greenish-gray to greenish-black clay shale and mudstone 10 feet thick are in fault contact with underlying sandstone and chert-pebble conglomerate, presumably of the Arcturus Formation, and the shale encloses small blocks of similar chert-pebble conglomerate, probably faulted into or slumped onto parts of it. Faulted onto this shale is a very light gray tuffaceous sandstone about 25 feet thick; beds in this unit are 3–36 inches thick and are gently arched. The upper shale, which is interbedded with mudstone, rests conformably on the tuffaceous sandstone and also contains blocks of conglomerate.

Plant remains that were collected, probably from one of the shale layers in the roadcut, by Neal Smith and Fred Digert (written commun., 1959), were tentatively identified by D. I. Axelrod as:

Sequoia fastigata (Sternberg) Heer
Podozamites sp. cf. *P. lanceolatus* (Lindley and Hutton) Braun
Heliconia sp.
Cercidiphyllum ellipticum (Newberry) Brown

Axelrod dated these forms as Late Cretaceous or early Paleocene. The collection was referred to by Van Houten (1956, p. 2808). Estella Leopold tried in vain to find pollen in several samples of the shale and sandstone.

TERTIARY SYSTEM

BASALTIC ANDESITE VITROPHYRE

A single dike of basaltic andesite vitrophyre intrudes the Ely Limestone low on the north wall of the upper part of Cave Creek. The dike is not exposed, but its presence is shown by a heavy local concentration of small fragments and blocks of dark-gray igneous rock. The distribution of the blocks indicates that the dike is only a few feet wide, is less than 200 feet long, and trends northwestward along the valley wall.

The rock is very dark gray to grayish black and has a very thin yellowish-brown coating on its weathered surface. Rounded inclusions of quartz 5–15 mm long are conspicuous but form only a small percentage of the rock. In thin section the rock is seen to contain 5 percent of phenocrysts and xenocrysts and about 35 percent of small crystals having trachytic texture, all in a glassy groundmass. The larger crystals are 2 percent labradorite phenocrysts and 1 percent each augite phenocrysts, magnesian orthopyroxene phenocrysts, and quartz xenocrysts. Of the small crystals, labradorite makes up about 15 percent of the rock, augite 10 percent, orthopyroxene 9 percent, and reddish-brown opaque material that may be iron oxide or an alteration product of olivine about 1 percent. The quartz is strongly resorbed by the glass, and the orthopyroxene has thin reaction rims of augite. The glass is very pale brown, is crowded with fine crystallities, and has an index of refraction of 1.533, appropriate for andesitic or basaltic glass.

The abundance of unaltered glass indicates that the dike was emplaced at shallow depth and is comparatively young. In this respect the vitrophyre resembles the abundant volcanic rocks in the quadrangle more closely than it does the granophyric porphyritic rhyolite dikes. Fragments of similar but thoroughly weathered basalt or andesite are included in the basal conglomerate underlying the volcanic rocks of Eocene and Oligocene(?) age in this area. The dike of vitrophyre is probably a feeder of the oldest rock of the volcanic sequence but is probably not much older than the other rocks of that sequence. Its similarity in composition and in stratigraphic position to the basal member (*1a*) of the Kalamazoo Volcanics of Young (1960b), further north in the Schell Creek Range, indicates that they are correlative, but none of the local volcanics is of basaltic andesite.

CONGLOMERATE

Pale-reddish-gray conglomerate containing subordinate limestone and tuff occurs in lenticular bodies as much as 1,000 feet thick and rests unconformably on upper Paleozoic rocks. The conglomerate underlies about 3 square miles within the quadrangle, chiefly in six separate bodies alined slightly askew to the crest of the Schell Creek Range. A small amount of the conglomerate also occurs along the west flank of the range at the south end of the Taylor district. Separate bodies of conglomerate lie on separate structural blocks, and many underlie a flow of quartz latite vitrophyre. Slopes eroded on the conglomerate are gentle to moderately steep and are strewn with residual pebbles and cobbles in a reddish-gray or pale-reddish-brown to yellowish-gray soil.

The best readily accessible exposures of this conglomerate are in the gullies tributary to Cave Creek from the south, but nowhere is a complete section well exposed. The thick body of conglomerate south of Cave Creek is divisible into two units, the upper one gray and only slightly indurated, and the lower one reddish gray and more indurated. Their contact is shown on the map by marker bed b. Conglomerate also occurs in smaller bodies that cannot be correlated with either unit. A thin limestone lens is mapped as marker bed a, and lenses of tuff are mapped as part of the conglomerate.

The base of the conglomerate is poorly exposed where it overlies the Chainman Shale, but three-quarters of a mile southwest of Cooper Canyon irregularities along its contact with the Arcturus Formation show that it rests on an old erosion surface of slight local relief. Here the upper 3 feet of limestone in the Arcturus Formation is broken and stained reddish brown, the color of the basal conglomerate. The relief on the unconformity beneath the conglomerate is much less than that on the present surface in this area, on which there is no reddish-brown soil. Between Cave Creek and Cooper Canyon the basal contact is a strongly angular unconformity, but elsewhere it is only slightly unconformable or merely disconformable.

The conglomerate has indistinct bedding planes from several feet to 30 feet apart or it is massive; the fragments are poorly sorted except where there are small lenses of sandstone or roughly alined boulders (figs. 9, 10). Pebbles are generally dominant, cobbles are common and locally dominant, and boulders are present here and there. The conglomerate has a sandy and silty matrix containing much calcium carbonate and finely disseminated brown iron oxide. Boulders and cobbles are largely restricted to the lower 30 feet of the conglomerate just south of Cooper Canyon and to the basal 200–400 feet of the conglomerate south of Cave Creek. The clasts are generally subangular to subrounded, but in places they are moderately well rounded. The degree of rounding of the clasts thus is intermediate between that in the conglomerate member of the Arcturus Formation and that in the younger clastic rocks except the lacustrine gravel.

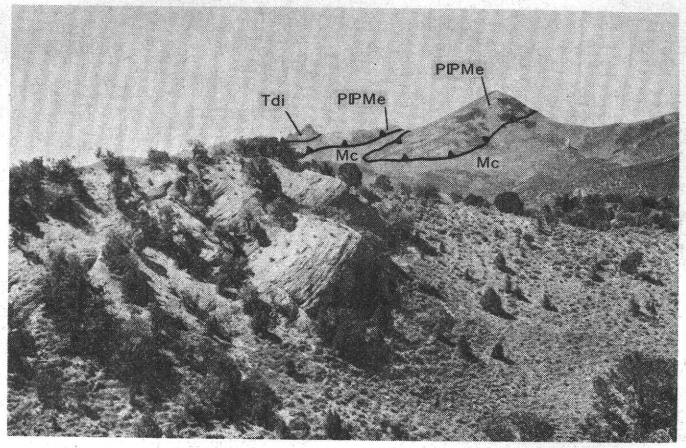

FIGURE 9.—Massive outcrops (foreground) of the lower unit of the Eocene conglomerate, a half mile southwest of Cooper Summit. View is southward toward peak having the Taylor bench mark at its summit. Ledges of Ely Limestone (PIPMe) are in a klippe above the Chainman Shale (Mc), which underlies the treeless bench around the peak. The crag in the middle of the skyline is underlain by the edge of a large plug of dacite vitrophyre (Tdi).

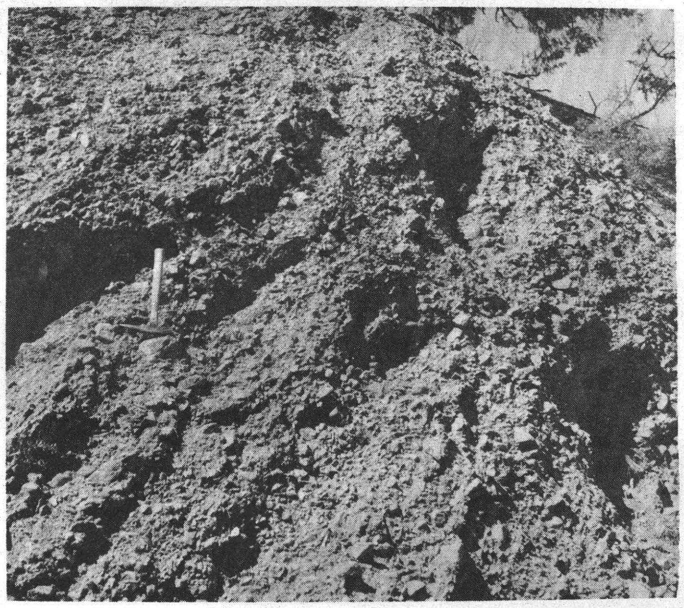

FIGURE 10.—Details of sorting, roundness, and size of clasts in the Eocene conglomerate, half a mile southwest of Cooper Summit.

Most of the clasts consist of cherty fossiliferous limestone, derived from the Ely Limestone and possibly from the Arcturus Formation, or of sandstone from the Rib Hill Sandstone and adjacent formations. Clasts of greenish-gray quartzite, probably derived from the Chainman Shale, and of nearly white fine-grained vitreous quartzite probably derived from the Eureka Quartzite, are generally present but are nowhere abundant. There are some fragments of other kinds of rock: (1) on the east flank of the Duck Creek Range, Prospect Mountain Quartzite, Pole Canyon Limestone, and Upper Cambrian and Lower Ordovician limestone; (2) southwest of Majors Place, Prospect Mountain Quartzite, Pole Canyon Limestone, unmetamorphosed fossiliferous limestone from upper part of the Lincoln Peak Formation, unmetamorphosed and unsheared Upper Cambrian and Lower Ordovician limestone, granite or quartz monzonite, and altered porphyritic rhyolite. The granitic cobbles are more deeply weathered than similar rocks on alluvial fans at the foot of the southern Snake Range —an indication that the environment in which the Eocene gravel was deposited was different from the Pleistocene environment. The granitic rock consists of about 40 percent quartz, 35 percent slightly granophyric orthoclase, 20 percent albite (An_{2-6}), and 3 percent muscovite. The upper unit of the conglomerate on both sides of Cave Creek contains some limestone fragments having nodules of a bluish-opaline material. Rare fragments of phyllite and one of granite were found on slopes underlain by conglomerate in the basin of Cooper Canyon. The conglomerate contains almost no fragments of volcanic rocks except in the tuff lenses. The middle part of the conglomerate where exposed on the south wall of the upper part of Cave Creek contains some fragments of weathered basaltic andesite, possibly derived from a dike similar to the one on the north valley wall. The conglomerate lenses south of Clear Spring contain a few cobbles and pebbles of a fresh tough dark-gray andesite vitrophyre of unknown source. This rock contains phenocrysts of hornblende and plagioclase in a glass having a refractive index of 1.518.

Much of the red color of the conglomerate is imparted by the fragments from the Rib Hill Sandstone, but some may be derived from the red soil that appears to have underlain the conglomerate in places. The upper unit of the conglomerate is grayer, perhaps because the red soil had been stripped from the source area by the time this part was laid down, and also because it contains fewer fragments of the Rib Hill Sandstone.

One lens of limestone is interbedded low in the conglomerate on the east wall of the lower part of the South Fork of Cave Creek. It forms a ledge 1,500 feet long and commonly about 10 feet thick but as much as 30 feet thick in some places. The limestone is very finely crystalline, is yellowish gray to very pale orange, and has beds commonly 2–6 inches thick. Some bedding planes are nodular and locally the top of the limestone contains a breccia of limestone flakes resembling fragments of those formed between mud cracks.

Much tuff is interbedded with the conglomerate at Clear Spring, and south of Cave Creek more than a dozen lenses of tuff appear in the upper unit of the conglomerate. The largest lenses of distinctly tuffaceous rock are mapped as tuff lenses within the Eocene conglomerate. Tuff may also be mixed with the matrix of some of the conglomerate, but is thoroughly masked by other fine-grained clastic material.

The tuff is gritty or sandy, rich in crystal fragments, and white, very light gray, yellowish gray, or pinkish gray. It is generally massive, but locally it contains some thin beds of well-sorted tuffaceous sandstone. In thin section, 20–25 percent of the tuff is seen to consist of crystals or crystal fragments; thus the rock is 6–10 percent quartz, 8–10 percent andesine near An_{45}, 2–3 percent sanidine(?), and 2–4 percent biotite. Minor accessories are magnetite, zircon, sphene, apatite, and allanite. The rest of the tuff(?), consists of vesicular, partly devitrified glass, clay minerals, cristobalite, and small amounts of lithic fragments, iron oxide, and chalcedony. The glass of one tuff body has an index of refraction of 1.503.

The reddish-gray conglomerate is the first post-Permian deposit in the area that contains evidence of source area, mode of transport, and conditions at the site of deposition. Most of the fragments were derived from rocks now exposed within the area, but the granitic cobbles are foreign to the area, and the limestone in the conglomerate southwest of Majors Place is unmetamorphosed, whereas the nearest limestone in place is distinctly altered. The persistent scattered clasts of quartzite believed to be from the Eureka Quartzite probably came from outside the area of the present Schell Creek Range, for the Eureka has been eroded from only a small part of that area. The sizes of the fragments and their degree of roundness and sorting also indicate that they came from distant sources, probably outside the quadrangle.

Quantitative estimates of the distance of transport and the location of the actual source areas remain conjectural. Inasmuch as fragments now being deposited at the distal ends of fans 5–10 miles from the present range front are more angular than most of the clasts in the conglomerate, these subrounded clasts must have been carried farther than the fragments on the present fans—possibly two to three times as far. The combination of rocks represented by the fragments in the conglomerate near Majors Place is not now exposed at any place nearer than in the southern Snake Range, about 20 miles to the east. The northern Snake Range is one of the few areas in which the

Paleozoic rocks have been largely removed; it is also the largest of these areas and one of the nearest. The red conglomerate therefore presumably was derived largely from sources east of the Schell Creek Range, possibly in the Snake Range.

That debris was probably carried by streams to an old piedmont area is indicated by the sorting and bedding of the deposits. There is no indication that the site of deposition had as much local relief as it has now, and there is some evidence south of Cooper Canyon that the surface beneath the conglomerate was deeply weathered and gentle in relief. Some relief is indicated by the great variations in the thickness of the conglomerate over short distances; not all of this variation can be the result of later erosion. The formation and preservation of the weathered rinds on the granite or quartz monzonite cobbles indicate that deposition was less rapid or the climate was wetter than at present, these conditions perhaps affecting the red color in the conglomerate. The limestone lens was probably deposited in a lake, the life of which may have been terminated by desiccation, as indicated by the capping layer of curled limestone flakes.

Inasmuch as the conglomerate is unfossiliferous, it cannot be dated exactly. From its general appearance and its relation to volcanic rocks of generally similar composition, it is tentatively correlated with the Kinsey Canyon Formation of Young (1960b, p. 163–164), 15 miles to the north, and with either the conglomerate low in the Sheep Pass Formation of Eocene age or that low in the Garrett Ranch Group of Oligocene(?) age described by Winfrey (1960, p. 126–133), 30–50 miles to the southwest. Similar conglomerate appears in the Egan Range, just west of the quadrangle, and low in the White Sage Formation of Eocene(?) age (Nolan, 1935, p. 42), at the north end of the Deep Creek Range. If the correlation of the reddish-gray conglomerate in the Connors Pass quadrangle with that in the Sheep Pass Formation or with that in the Garrett Ranch Group or with that in the White Sage Formation proves valid, the conglomerate is of Eocene age and it is more likely to be post-Eocene than pre-Eocene.

QUARTZ LATITE VITROPHYRE

The conglomerate just described is conformably overlain by glassy effusive quartz latite that underlies about 1 square mile of the high area between Cave Creek and Cooper Canyon and also smaller areas north of Cave Creek. At each locality there is only one lava flow. The lava south of Cave Creek is about 300 feet thick, but at the other places it is much thinner.

Most of this lava is very resistant and underlies sharp ridges and a mesa, but some of it underlies low knobs. The thick flow south of Cave Creek is polygonally jointed and forms a low cliff above an apron of blocky talus. On relatively flat areas the rock weathers to a grus interspersed with residual blocks.

The basal 15 feet of the flow, best exposed just above Summit Spring, is a gray breccia of glassy fragments in a grassy matrix. Most of the flow is pale reddish gray on fresh factures, but weathers brownish gray, pale reddish brown, or dusky reddish gray. It has a strong platy structure formed by alined, flattened, and stretched vesicles, which are commonly less than 2 inches long and make up 10–20 percent of the rock. Many of the phenocrysts of feldspar, biotite, and quartz also lie parallel to the platy structure. Near the top of the flow the rock is more vesicular, and in places the flow layers are mammillary rather than platy. Small angular inclusions of tuff are locally abundant.

Phenocrysts make up about 40 percent of the rock and the groundmass is an unaltered or little altered flow-laminated glass. The proportions of the phenocrysts do not differ widely: 10–20 percent of the rock consists of plagioclase phenocrysts, which have an anorthite content of An_{20-55} but commonly are only An_{35-48}, 10–15 percent quartz phenocrysts, and 2–10 percent sanidine(?) phenocrysts. The rock contains 0.5–1 percent sphene and a very small amount of magnetite (?), apatite, zircon, and allanite. Few of the phenocrysts are more than 4 mm long, and most are only about 0.5 mm in diameter. The larger phenocrysts are generally subhedral or fragmental, but the smaller ones as well as the grains of accessory minerals are commonly euhedral. Quartz and, more rarely, plagioclase have the embayed or rounded shapes of partly resorbed crystals.

The glass is very pale brown and contains abundant microlites. Its index of refraction has a narrow range, 1.501–1.505, and is most commonly 1.502–1.504, a range that barely overlaps that of the glass of the other volcanic rocks. Most of the glass is replaced in small part by a radially fibrous mineral, and in some of the specimens it is almost completely devitrified. A small amount of cristobalite fills or lines vesicles in many speciments, and opal and chalcedony are also common.

Throughout the report I present the full range of the anorthite content, as measured in all thin sections of a given mapped unit, such as the quartz latite vitrophyre flow. This range extends from the most sodic zone in one plagioclase crystal to the most calcic zone in any other plagioclase crystal. Where there is abundant data, such as for the quartz latite vitrophyre (measurements were made on about 70 plagioclase crystals in 11 thin sections), a second range of anorthite content is given to show the common composition of the plagioclase.

The anorthite content was determined by measuring extinction angles of grains in which the (001) and (010) crystal faces are oriented normal to the stage, and by applying these results to the appropriate high-temperature curve for volcanic rocks or low-temperature curve for plutonic rocks, as given by Tröger (1956, p. 111).

Indices of refraction of glass were obtained by use of a sodium lamp and immersion oils calibrated to about 0.004. These oils were mixed to about the nearest 0.001, and appropriate temperature corrections were applied.

Chemical and spectrographic analyses of five specimens of quartz latite vitrophyre from the lava flow are presented in table 11. The analyses are so similar that the specimens could all represent a single flow.

The quartz latite vitrophyre is probably only a little younger than the conglomerate, which is dated as Eocene. Zircon from this flow is given an age of 50 millions years ±10 million years by lead-alpha radiogenic methods (T. W. Stern, written commun., 1964). Most potassium-argon ages given of volcanic rocks in eastern Nevada are slightly younger (30–40 my) and probably are more reliable.

TABLE 11.—*Chemical and semiquantitative spectrographic analyses of rocks from Egan Range (Shawe, 1961) and from Schell Creek Range*

[Spectrographic analyses of samples from localities 3–8 by P. R. Barnett, who reported to the nearest number in the series 7, 3, 1.5, 0.7, 0.15, and so forth, in percent. These numbers represent midpoints of group data on a geometric scale. Elements looked for but not found: Ag, As, Au, Bi, Cd, Ge, Hf, Hg, In, Ir, Mo, Os, Pd, Pt, Re, Rh, Ru, Sb, Ta, Te, Th, Tl, U, W, and Zn. In samples from Egan Range, Li also was looked for but not found]

Locality	Egan Range	Schell Creek Range							Egan Range
Rock name	Rhyolite	Quartz latite vitrophyre							Rhyolite
Rock type	Welded tuff	Lava flows						Intrusive	
Sample loc. (pl. 1)	------------	------------	3	4	5	6	7	8	------------
Lab No	Average of four samples	Average of five samples	H3179	H3180	H3181	H3182	H3188	H3178	Average of four samples
Chemical analyses									
SiO_2	69.71	71.44	72.13	70.60	71.85	70.64	71.97	75.89	73.47
Al_2O_3	14.11	13.84	13.36	14.26	13.71	13.99	13.88	13.10	13.56
Fe_2O_3	1.12	1.08	1.06	1.26	1.22	1.11	.77	.30	.45
FeO	1.41	1.06	1.06	1.06	.99	1.13	1.06	.36	.32
MgO	.81	.69	.66	.70	.72	.70	.66	.10	.09
CaO	2.85	2.49	2.40	2.39	2.56	2.59	2.50	.86	.90
Na_2O	2.51	2.69	2.78	2.33	2.81	2.71	2.83	3.51	3.50
K_2O	3.84	4.29	4.01	4.82	4.05	4.39	4.17	4.69	4.89
H_2O^+	1.81	.72	.79	1.00	.61	.77	.45	.32	2.20
H_2O^-	1.00	.67	.62	.70	.59	.64	.79	.42	.21
TiO_2	.45	.39	.37	.41	.39	.41	.38	.05	.05
P_2O_5	.10	.10	.20	.01	.09	.11	.09	.02	.03
MnO	.05	.05	.09	.06	.04	.04	.04	.06	.08
CO_2	.02	.01	.01	.01	.00	.01	.01	.01	.01
Cl	.02	.02	.02	.02	.02	.02	.01	.01	.02
F	.07	.08	.08	.08	.07	.08	.07	.13	.12
Subtotal	------------	------------	99.64	99.71	99.72	99.34	99.68	99.83	------------
Less O	------------	------------	.03	.03	.03	.04	.03	.05	------------
Total	------------	------------	99.61	99.68	99.69	99.30	99.65	99.78	------------
Semiquantitative spectrographic analyses									
B	0	0	0	0	0	0	0	0.0015	0.0011
Ba	.19	.07	.07	.07	.07	.07	.07	.007	.003
Be	.00015	.00018	.00015	.00015	.00015	.0003	.00015	.0003	.0003
Ce	.023	.018	.015	.015	.015	.03	.015	0	0
Co	.0003	.0003	.0003	.0003	.0003	.0003	.0003	0	0
Cr	.0003	.00015	.00015	.00015	.00015	.00015	.00015	.00015	<.0003
Cu	.0003	.00027	.0003	.0003	.0003	.0003	.00015	.00015	<.0003
Ga	.0009	.0015	.0015	.0015	.0015	.0015	.0015	.003	.0023
La	.011	.009	.007	.007	.007	.015	.007	0	0
Li	0	0	0	0	0	0	0	.015	0
Nb	.0019	.0015	.0015	.0015	.0015	.0015	.0015	.003	.003
Nd	.009	.004	0	0	0	.015	.007	0	0
Ni	<.0003	.00006	0	0	0	0	.0003	0	0
Pb	.0009	.0015	.0015	.0015	.0015	.0015	.0015	.003	.003
Sc	.0007	.0007	.0007	.0007	.0007	.0007	.0007	.0007	.0003
Sn	0	0	0	0	0	0	0	.0007	.0007
Sr	.06	.07	.07	.07	.07	.07	.07	.007	.003
V	.004	.003	.003	.003	.003	.003	.003	0	0
Y	.0023	.0015	.0015	.0015	.0015	.0015	.0015	.003	.003
Yb	.00023	.00015	.00015	.00015	.00015	.00015	.00015	.0003	.0003
Zr	.019	.015	.015	.015	.015	.015	.015	.007	.005

Welded tuffs of somewhat similar chemical and mineral composition from the Egan Range were described by Shawe (1961, p. 178–181); analyses of them (table 11) show that the quartz latite vitrophyre flows of the Schell Creek Range resembled the welded tuff to the west in the Egan Range more closely than that tuff resembles the intrusive rhyolite in the Egan Range, also described by Shawe. The chemical similarity between the intrusive rhyolite described by Shawe and the intrusive quartz latite vitrophyre is even more impressive.

LATITIC ROCKS

The flows of quartz latite vitrophyre are conformably overlain by a thick sequence of tuff. Two small lenses of welded tuff and two intrusive bodies of vitrophyre and a lava flow compose the formation that underlies an area of about 1 square mile near Cooper Summit and a small area half a mile northeast of Cave Creek Reservoir. The tuff forms gentle to moderately steep slopes between the more competent lava flows, these slopes being so thickly strewn with blocks of tuff and other volcanic rocks that outcrops are few. The tuff is best exposed along the road 1 mile south of Cooper Summit.

The tuff is almost everywhere a massive, blocky-weathering, very light gray rock, but locally it forms rounded bosses having pitted surfaces. It consists predominantly of fine-grained volcanic ash and 20–35 percent of crystals and crystal fragments, but near the south end of the quadrangle it also contains 20–30 percent of small lithic fragments, as much as 4 inches in diameter, of glass, vitrophyre, and white tuff. The vitrophyre fragments resemble the quartz latite more closely than they do the overlying dacite.

The tuff contains 8–15 percent plagioclase crystals, which have a composition range of An_{33-48}, 2–7 percent sanidine(?), 6–20 percent partly resorbed quartz, and 2–3 percent biotite. The accessory minerals include sphene, which is abundant, and magnetite(?), zircon, apatite, and allanite, which are all scarce. Shards of vesicular glass are abundant and some pumiceous fragments are also present. The index of refraction of the glass is 1.492–1.501—mostly lower than, but slightly overlapping, the range in the quartz latite vitrophyre flow. Devitrified glass and clay alteration minerals obscure much of the groundmass.

WELDED TUFF LENSES

Two lenses of welded tuff lie at or just above the base of the latite tuff. One is well exposed on the ridge west of the road about a mile south of Cooper Summit, and the other is poorly exposed three-quarters of a mile northeast of Cave Creek Reservoir. The lens south of Cooper Summit is scarcely half a mile long and 100 feet thick, and it apparently grades in all directions into normal tuff. The rock has a strong platy parting and contains flat vitrophyre inclusions, both parallel to the bedding. The lens thins gradually northward and is probably continuous with a layered well-indurated tuff unit about 15 feet thick, which lies about 50 feet above the base of the latite tuff. Although the vitric fragments which make up most of the layered tuff are not markedly elongate like those in the welded tuff, some are subrounded and have concentric flow structures, these features suggesting that the fragments are slightly flattened clots or volcanic bombs. The layered tuff also contains devitrified glass, opaline vein material, and numerous fragments of phenocrysts. The fragments and crystals are crushed in some places, especially where they are in contact with one another.

The mineral composition of the welded tuff is identical with that of the surrounding tuff and much like that of the quartz latite flow. The refractive index of the glass is 1.494–1.500. In thin section many of the glassy fragments are seen to be spindle shaped, the ratio of length to thickness ranging from 5:1 to 10:1. Some large glass fragments have a faint perlitic fracture that is not common in the other vitrophyres of this area.

INTRUSIVE BODIES OF QUARTZ LATITE VITROPHYRE

The latite tuff is cut by two small bodies of quartz latite vitrophyre. One of these underlies an elliptical area at Cooper Summit. Its shape and the absence of nearby faults indicate that it is intruded, rather than faulted, into the tuff. The rock is about 50 percent phenocrysts in a strongly flow-laminated glassy groundmass. Plagioclase phenocrysts (An_{45-48}) make up 30 percent of the rock, and phenocrysts of quartz, biotite, and hornblende, 8, 6, and 4 percent, respectively; the accessories are sphene, magnetite(?), apatite, and allanite. The refractive index of the glass is 1.501. Mineralogically, then, this intrusive rock closely resembles the quartz latite vitrophyre of the flows except for the presence of hornblende and the absence of potassium feldspar phenocrysts; in these respects the rock is transitional with the younger dacite flows.

A small area near the mouth of the canyon of Cooper Wash also is underlain by quartz latite vitrophyre that seems to be intrusive. Near the north and east edge of this body there are some steeply inclined peripherally striking flow laminae and some coarse vitrophyre breccia and agglomerate. Most of the petrographic features of this body of quartz latite vitrophyre are like those of the lava flows beneath the latite tuff, but the groundmass is partly cryptocrystalline, and the refractive index of the remaining glass can only be determined to two significant figures as 1.51. Alteration to clay minerals is moderately intense, but is much less than in the dikes of porphyritic rhyolite.

The intrusive quartz latite vitrophyre rock (table 11) differs slightly in chemical composition from the associated extrusive rock. It contains significantly more silica, soda, and fluorine, and less lime, total iron oxides, magnesia, and water. The chemical and possibly also the petrographic differences may be related to the greater alteration of the intrusive rock than of the extrusive rock, these relations indicating that post-eruption fluids may have issued from the vent. It is possible, however, that the composition of the latite magma differed from place to place.

UPPER LAVA FLOW OF QUARTZ LATITE VITROPHYRE

At the mouth of Cooper Wash a thick body of latite tuff is apparently overlain by a lava flow that somewhat resembles the quartz latite vitrophyre flow beneath the tuff at Cave Creek, but which contains a greater abundance and variety of silica minerals, is brecciated, and is affected by a pervasive clay-mineral alteration. There are several possible correlations between it and the volcanic rocks along Cave Creek. It seems most likely that the tuff bodies are contemporaneous and the lava flow slightly younger, for quartz latite cuts the tuff in both places, and the spreading of tuff is less restricted by topography than is the spreading of lava flows. The vitrophyre agglomerate between the quartz latite vitrophyre intrusive body and the tuff along Cooper Wash indicates that the intrusive body may be a plug in the vent from which the tuff was ejected, which is perhaps also the vent of the younger flow. Regardless of the details of the relations between them, the tuff and the quartz latite vitrophyre rocks seem to be genetically associated and of nearly the same age.

ENVIRONMENT, AGE, AND CORRELATION

The initial period of major volcanism, which began in this area with small outbursts of tuff and continued with the extrusion of the quartz latite vitrophyre flows, culminated with a large explosive outburst that deposited a thick blanket of tuff over a large area. Some of the tuff was erupted early and was welded, but, judged from the slight deformation of the glassy fragments and the presence of unoxidized biotite, the welding was not intense and the temperature was not very high. Some of the glassy bombs that were deposited at the horizon of the welded tuff body and just beyond its edge apparently were so soft that they flattened slightly under the weight of the rapidly accumulating mantle of tuff; some of the more rigid fragments and phenocrysts were crushed under this pressure. The welded tuff may have been formed in a similar manner from a larger accumulation of hot viscous glassy fragments that were under moderately high lithostatic pressure. Regardless of origin, the welded tuff bodies differ from the extensive sheets of welded tuff studied elsewhere in this part of the Great Basin by Mackin (1960) and Cook (1960) in that they are extremely local. Near the end of this large explosive outburst, or just after it, a latite magma having slight dacite affinities intruded some of the tuff and locally poured out over it.

Biotite from the intrusive body of quartz latite vitrophyre near the mouth of the canyon of Cooper Wash has been dated by the potassium-argon method as 38 million years (± 10 percent) (S. S. Goldich, written commun., 1964). This determination confirms an Eocene age, and probably indicates a late Eocene age. The sample was collected at sample locality 8 (pl. 1).

The tuff does not seem to have any correlative in the Kalamazoo Volcanics of the northern Schell Creek Range, except possibly an unanalyzed tuff. In the Egan Range, however, Shawe (1961) described a welded tuff that resembles in phenocrysts and chemical composition the lava flows believed to be associated with the welded tuff in the Cave Creek area.

DACITE VITROPHYRE

Some of the youngest volcanic rocks in the area lie conformably on the latite tuff or on the conglomerate of Eocene age, whereas some lie unconformably on rocks of late Paleozoic age or are intruded into upper Paleozoic rocks. These youngest volcanics have the chemical and mineral composition of dacite and a vitrophyric texture, but are varied within these limitations; I may therefore have mapped rocks of somewhat different age and origin as dacite vitrophyre. East of Cooper Summit the dacite vitrophyre consists of three flows, separated by thin layers of tuff, that are altogether more than 1,200 feet thick; elsewhere it also includes dikes and volcanic necks. The flows, tuff, and intrusive rocks are distinguished on the map, however, as separate units of the dacite vitrophyre.

Dacite vitrophyre underlies many small areas, which have a combined extent of about 3 square miles. Most of these areas lie between the upper part of Steptoe Creek and Majors Place; others lie on the west flank of the Schell Creek Range south of the Taylor district, and one is Rattlesnake Knoll, in Spring Valley near U.S. Highway 6–50.

The flows of quartz dacite vitrophyre differ greatly in topographic expression: they form high steep ridge crests east of Cooper Summit, gently rounded hills south of Cooper Canyon, and gentle piedmont slopes, hardly distinguishable from the adjacent gravel-covered slopes, southwest of the Taylor district. The steep slopes are covered with colluvium and blocky talus, but the gentler ones are mantled with grus and scattered residual blocks.

Fresh surfaces of quartz dacite vitrophyre are medium gray to grayish red, and the weathered surfaces are moderate yellowish brown, moderate brown, pale reddish brown, to brownish black. The fresh rock is generally darker than the quartz latite vitrophyre. The dacite is slightly vesicular, and the more competent rock has a platy parting. Roughly polygonal joints stand at right angles to the parting and partly cause weathering of the dacite into blocks. Where blocks are abundant and outcrops are absent, as in some places southwest of the Taylor district and just south of Cooper Canyon, the source of the blocks may be an agglomerate body. Phenocrysts of feldspar, biotite, and hornblende are common, and in places there are also phenocrysts of quartz. In some places the dacite flow includes fragments of sedimentary rocks, and of igneous rocks resembling the dacite.

The only place where inclusions of sedimentary rocks are especially common in the dacite vitrophyre is at Rattlesnake Knoll. The inclusions are subangular to subrounded and range in size from grit to blocks 12 inches across although most are only 1–2 inches across. They are scattered in an igneous matrix, and in places they are roughly sorted into layers parallel to the nearly flat lying partings. The fragments consist of quartzite, rhyolite, fine-grained and recrystallized massive limestone, shaly limestone, and phyllite, all common in the alluvial fans in this area as well as in some of the fans on the east side of Spring Valley. The nearly horizontal position of the partings, the size sorting of some fragments, and the great variety of the fragments, seem to indicate that this dacite vitrophyre is a lava flow that picked up debris from the surface of an underlying fan.

The abundance of the phenocrysts is widely varied —from 15 to 50 percent. The relative abundance of different kinds of phenocrysts is also widely varied: plagioclase phenocrysts (An_{35-50}, common range; An_{28-95}, extreme range) make up 8–25 percent of the rock, quartz 0–15 percent, biotite 0–8 percent, hornblende 0–5 percent, augite 0–3 percent, and hypersthene 0–5 percent. The groundmass is generally glassy, but in some specimens it is cryptocrystalline. In a typical specimen, andesine phenocrysts make up 20 percent of the rock, quartz phenocrysts 10 percent, biotite 5 percent, hornblende 3 percent, and augite 1 percent. The accessory minerals include sphene, magnetite(?), apatite, zircon, and allanite. The dacite differs from the latite by its complete lack, among the phenocrysts, of potassium feldspar, its occasional lack of quartz and biotite, its greater abundance of hornblende, and its content of augite and hypersthene.

The characteristics of the phenocrysts also vary widely. Most of them are less than 2 mm in diameter but some are more than 4 mm. The plagioclase crystals are commonly subhedral or fragmental, and some are embayed or rounded by resorption. Normal zoning, oscillatory zoning, and, in one specimen, reverse zoning occur in the plagioclase. The range of composition between zones in most of the plagioclase phenocrysts is small, but in several specimens the anorthite content between zones varies more than 15 percent. The quartz phenocrysts are generally fragmental, and many of them are embayed or rounded by resorption. Biotite and hornblende are commonly subhedral to euhedral phenocrysts, most of which are sheathed with iron oxide particles and thus are slightly oxidized, but a few of which are altered to oxyhornblende and oxybiotite and thus are intensely oxidized. Some of the hornblende is slightly altered to actinolite or to chlorite and calcite, and some of the biotite is replaced by epidote, sericite, and iron oxide. A few grains of biotite have reaction rims of hornblende. Augite and hypersthene are commonly euhedral.

The groundmass is a banded clear to pale-brown glass, which contains a few microlites and some pockets of devitrified glass. The index of refraction of the glass is commonly 1.508–1.513, but in one specimen it is 1.501. Some of the groundmass contains a little cristobalite and is pervasively altered to clay minerals.

Chemical analyses of eight samples of dacite vitrophyre are shown in table 12, and the places where they were collected are shown on the geologic map (pl. 1). The results of the analyses are very similar, and as a group the analyses are distinct from those of the quartz latite vitrophyre shown in table 11; there is more variation, however, within the dacite samples than within the latite samples.

INTRUSIVE ROCKS

The intrusive bodies of quartz dacite in the Connors Pass quadrangle range from dikes about 100 feet long and 30 feet wide to a volcanic neck southeast of Crethers Springs that is about three-quarters

TABLE 12.—*Chemical and semiquantitative spectrographic analyses of dacite vitrophyre*

[Chemical analyses by C. L. Parker. Looked for in spectrographic analyses but not found: Ag, As, Au, B, Bi, Cd, Ge, Hf, Hg, In, Li, Mo, Pd, Pt, Re, Sb, Sn, Ta, Te, Th, Tl, U, W, and Zn. Spectrographic analyses of samples 9, 10, 11, and 14 by P. R. Barnett, who reported to the nearest number in the series 7, 3, 1.5, 0.7, 0.3, and 0.5 in percent, and also looked for, but did not find: Ir, Os, Rh, and Ru. Spectrographic analyses of samples 12, 13, 15, and 16 by J. C. Hamilton, who reported to the nearest number in the series 1, 0.7, 0.5, 0.3, 0.2, 0.15, and 0.1 (which represent approximate midpoints of group data on a geometric scale). About 30 percent of amounts determined present by semiquantitative spectrographic analyses agree with amounts determined by chemical analyses]

Rock type	Intrusive	Lava flows							
Sample loc. (pl. 1)	9	10	11	12	13	14	15	16	Average of 8 samples
Lab. No	H3177	H3185	H3184	H3660	H3661	H3183	H3658	H3659	
Chemical analyses									
SiO	67.83	67.31	66.03	63.01	69.73	65.41	66.71	67.67	66.71
Al_2O_3	14.64	14.81	15.17	15.17	14.17	15.10	14.97	14.53	14.82
Fe_2O_3	2.27	3.11	3.16	3.25	2.69	2.24	2.41	1.59	2.59
FeO	1.72	.80	1.36	1.62	.64	2.16	1.70	1.80	1.48
MgO	1.19	1.11	1.23	1.85	.82	1.37	1.26	1.04	1.23
CaO	3.75	3.30	3.76	4.97	2.98	4.14	4.18	3.08	3.77
Na_2O	3.01	2.91	2.99	2.71	2.94	2.91	3.06	2.94	2.93
K_2O	3.40	3.84	3.42	2.98	3.98	3.42	3.45	3.75	3.53
H_2O+	.47	.53	.53	1.99	.47	1.37	.32	2.06	.97
H_2O-	.51	1.08	1.02	.64	.60	.26	.42	.20	.59
TiO_2	.60	.63	.70	.77	.53	.72	.70	.56	.65
P_2O_5	.16	.16	.20	.22	.14	.23	.21	.14	.18
MnO	.09	.05	.12	.09	.03	.08	.08	.07	.08
CO_2	.05	.00	.01	.32	.01	.17	.30	.01	.11
Cl	.02	.02	.01	------	------	.03	------	------	------
F	.08	.08	.07	------	------	.06	------	------	------
Subtotal	99.79	99.74	99.78	99.59	99.73	99.67	99.77	99.44	99.64
Less O	.03	.03	.03	------	------	.04	------	------	------
Total	99.76	99.71	99.75	------	------	99.63	------	------	------
Semiquantitative spectrographic analyses									
Ba	0.15	0.07	0.07	0.1	0.1	0.07	0.15	0.1	0.95
Be	.00015	.00015	.00015	.00015	.0002	.00015	.0002	.0002	.00017
Ce	.015	.015	.015	.02	.02	.015	.02	.015	.017
Co	.0007	.0007	.0007	.001	.007	.0007	.001	.001	.0019
Cr	.0015	.0015	.0007	.0007	.0007	.0007	.001	.0007	.0009
Cu	.0003	.0003	.0003	.001	.001	.0003	.0005	.0007	.0006
Ga	.0015	.0015	.0015	.002	.003	.0015	.003	.003	.0021
La	.007	.015	.007	.007	.015	.007	.01	.007	.009
Nb	.0015	.0015	.0015	.002	.002	.0015	.002	.002	.0018
Nd	.007	.007	.007	<.01	.015	.007	.015	<.01	------
Ni	.0007	.0015	0	0	0	0	0	0	.0003
Pb	.0015	.0015	.0015	.002	.002	.0007	.003	.002	.0018
Sc	.0015	.0015	.0015	.0015	.001	.0015	.0015	.001	.0014
Sr	.07	.07	.07	.07	.07	.07	.07	.05	.07
V	.015	.007	.007	.01	.007	.007	.01	.007	.0088
Y	.003	.0015	.003	.003	.003	.003	.003	.003	.0028
Yb	.0003	.00015	.0003	.0003	.0005	.0003	.0005	.0003	.0003
Zr	.015	.03	.03	.02	.02	.03	.05	.015	.026

of a mile in diameter. Most of them are clustered just south of Cooper Canyon. Intrusive contacts, chilled margins, and steeply inclined peripheral flow structures can be seen along the margins of most of these bodies. Many of the intrusive bodies are on the north-trending faults. The rocks of a few are moderately altered. Similarly altered bodies without structural evidence of intrusion occur in this area, but these may only have been volcanic rocks penetrated by gases escaping through vents.

The physical and chemical properties of the less altered intrusive rocks differ little from those of the flows. The index of refraction of glass from chilled margins is 1.510–1.513. The more altered rocks, however, are very light gray and are thoroughly fractured. Their groundmasses are cryptocrystalline and are pervaded with clay minerals. The plagioclase crystals are not albitized, as are some of those in the older porphyritic rhyolite dikes.

The large intrusive neck southeast of Crethers Springs is nearly circular in plan and consists of two kinds of rock. Along the west and north sides of the neck the rock is unaltered and highly competent, and it there underlies a rugged crescentic ridge. Along the west side, where the contacts are locally very steep, a wide-chilled margin and a peripheral breccia zone contain fragments of volcanic rock and of upper Paleozoic sedimentary rock. Strong joints dipping 35°–65° outward indicate that, aside from the local steep exposures, the contact dips inward as in a funnel. The core and south and east sides of the neck, however, are underlain by moderately altered incompetent rock. On these sides the contact is unexposed, but across one steep canyon on the northeast side it appears to be steep or to incline inward. The altered rock grades into the unaltered rock, and in places it encloses unaltered rock.

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Half a mile southwest of this large neck there is an intrusive body less than 300 feet wide and elliptical in plan. The rock of this body is pale purple to reddish gray, fine grained, and very brittle, being closely fractured in a radial pattern. This rock consists mainly of devitrified glass containing less than 2 percent andesine phenocrysts and accessory minerals.

TUFF LENSES

Two thin unexposed tuff lenses that separate flows of dacite vitrophyre east of Cooper Summit are mapped with the dacite vitrophyre. Poorly exposed tuff and tuffaceous sandstone that underlie the dacite vitrophyre flow just west of the townsite of Taylor are also mapped with it because of their mineralogic affinities with the dacite. They contain oxyhornblende and augite but no sphene; the refractive index of the glass is 1.508.

ENVIRONMENT, AGE, AND CORRELATION

The wide distribution of remnants of dacite vitrophyre and the still wider distribution of dacite fragments in the younger gravel units indicate that the dacite flows once covered a large area. After the more explosive phase of eruption, during which latite tuff was ejected, a less silicic magma was brought to the surface through vents that were partly in normal fault fissures. The largest intrusive body may be the root of a large volcano, and the others may be the roots of satellite cones.

Biotite from the upper of the flows east of Cooper Summit has been dated by the potassium-argon radiogenic method as 36 million years (± 10 percent) (S. S. Goldich, written commun., 1964). This age would probably indicate that the rocks are late Eocene, but some may be younger; they are here assigned an Eocene and Oligocene(?) age. The sample was collected along the jeep trail about 200 feet south of the saddle three-quarters of a mile east of Cooper Summit.

Winfrey (1960) reported a radiogenic age of 34 million years from similar volcanics above the Eocene Sheep Pass Formation, and Armstrong (1963) reported additional ages of 30 million years and 36 million years from these rocks. The rock is also similar to some of the flows in the poorly dated Kalamazoo Volcanics of Young (1960b), in the northern part of the Schell Creek Range, and to the dacite flow at the south end of the Snake Range (Drewes, 1958).

FANGLOMERATE

A gray poorly indurated cobble fanglomerate at least 600 feet thick is exposed between Cooper Canyon and U.S. Highway 6–50–93. It lies unconformably upon a dacite vitrophyre flow, the Eocene reddish-gray conglomerate, the Ely Limestone, and the Arcturus Formation. The fanglomerate is inclined almost as steeply as the underlying Eocene conglomerate and is faulted against limestone of Late Cambrian or Early Ordovician age. Slopes on the fanglomerate are gentle to moderately steep and are mostly covered with residual gravel, but the fanglomerate is exposed along some gullies and in a few small ledges on the steeper slopes.

The fanglomerate consists mainly of angular to subangular poorly sorted cobbles and boulders in a sandy matrix, but it includes some interbedded tuffaceous sandstone and pebble conglomerate. It is thick bedded, and in some exposures it shows no bedding. Most of the cobbles in it are derived from sheared Upper Cambrian and Lower Ordovician cherty limestone and from the Middle Cambrian Pole Canyon Limestone, which is slightly metamorphosed as it is in the exposures immediately to the east. To the north, where the fanglomerate is finer grained than it is generally elsewhere, it contains some pebbles of upper Paleozoic limestone.

The poor sorting and rounding of the fragments indicate that they are derived from a nearby source; this source must have been to the east, for elsewhere the rocks from which the fragments are derived are still covered by younger Paleozoic rocks.

In general character, and in its stratigraphic and structural relations to adjacent formations, the fanglomerate is comparable to the North Creek Formation of late Tertiary and Quaternary age of Young (1960b) in the northern Schell Creek Range. A Pliocene(?) age is tentatively assigned to the fanglomerate.

QUATERNARY SYSTEM

Quaternary gravel, sand, and silt underlie extensive areas of Spring Valley and Steptoe Valley and very restricted areas along the larger valleys within the Schell Creek Range. They are divisible into lacustrine deposits and alluvial gravels. The gravels are further divisible into three units, the youngest of which differs only slightly in distribution and character from the most recent deposits along the drainages.

OLDER ALLUVIAL AND FAN GRAVEL

In both Spring and Steptoe Valleys, there are deposits of gravel at least 200 feet thick, which underlie high relatively smooth slopes, incised by widely spaced gullies that have gently sloping sides and narrow bottoms. Most of the larger areas in which this gravel occurs lie close to the fronts of the Schell Creek

Range and extend up into its larger valleys. Except in one area near Cleve Creek, the erosion surfaces on this gravel slope more steeply toward the centers of Spring Valley and Steptoe Valley than the surfaces of the younger gravel deposits do, and at moderate distances from the Schell Creek Range this gravel is commonly overlapped by younger deposits. Close to the range fronts it is also overlapped by a thin apron of gravel that is generally unmapped.

The deposits beneath the mature high piedmont surfaces consist mainly of angular to subangular poorly sorted pebbles and cobbles but contain scattered boulders and bouldery lenses, all in a sandy, silty, and limy matrix. Boulders are most common to the northeast, as they also are in the younger deposits, but throughout the area the older gravel is coarser than the nearest deposits of younger gravel. The difference is most striking at the mouth of Cooper Canyon, where the older gravel contains scattered blocks of Eureka Quartzite as much as 10 by 10 by 15 feet in size, whereas the younger gravel contains few blocks as much as a foot long. The older gravel is only slightly indurated, though remnants of caliche rest upon it in many places .

The older gravel was deposited by streams on fans and narrow pediments at a time when coarse debris was being moved over steep gradients. Either the relief of the area was higher than it is now, or the climate was slightly different. The older gravel may be largely of early or middle Pleistocene age, for it is older than the lacustrine deposits, which are probably of late Pleistocene age.

YOUNGER ALLUVIAL AND FAN GRAVEL

Gravel having moderately mature surface morphology, typified by more closely spaced gullies and rougher surfaces along drainage divides than are common on the older gravel, generally underlies fans and terraces of intermediate height above the present stream courses along the range front in Spring Valley and away from the front in Steptoe Valley. This gravel overlaps and partly buries the older gravel and is at least 100 feet thick. Gravel of uncertain age that appears in isolated bodies within the Schell Creek Range is arbitrarily mapped with this gravel.

The younger gravel is much like the older gravel except that it is finer grained. Southwest of the Taylor district it contains much silt, which apparently has been washed from the older gravel for the younger deposits derived from the mountain slopes above the fans are much coarser even though similar in dip of bedding. Caliche locally caps the younger gravel, especially along and north of Bastian Creek.

The younger alluvial and fan gravel was also deposited by streams on fans and narrow pediments and was deposited at a time when relief or climate conditions were not as different from the present ones as they seem to have been slightly earlier. Some of the youngest part of this gravel overlaps the lacustrine deposits, but most of the gravel is probably contemporaneous with, or older than, the lacustrine deposits. Most of the gravel is probably of late Pleistocene age, but some of it may be slightly older or slightly younger.

LACUSTRINE DEPOSITS

Silt, sand, and gravel of lacustrine origin underlie the lowest parts of Spring Valley, along the east edge of the quadrangle. The upper surface of these deposits is nearly horizontal, but it has a relief of 10–15 feet and is characterized by benches and ridges parallel to the contours and by small hummocks. The most conspicuous ridges and benches lie at altitudes of 5,720 feet, 5,760 feet, and 5,840 feet, but faintly defined ones appear still higher, and at Rattlesnake Knoll a narrow irregular terrace is cut on the bedrock at an altitude of 5,880 feet. The area underlain by the lacustrine deposits is densely covered with tall sagebrush and other shrubs, this vegetation indicating proximity to ground water.

The low ridges are underlain by moderately well sorted and rounded pebble and shingle gravel, which is well exposed in a borrow pit just north of Highway 6–50. Most of the surface material between the ridges consists of silt in small dunes or gravel in lag concentrates. A few miles southeast of the quadrangle, similar gravel is underlain by a clayey silt containing small unidentified pelecypods and gastropods (Drewes, 1954, p. 81).

The geomorphology of these deposits at the edge of the quadrangle and the rounding and sorting of the pebbles contained in them indicate that they were deposited in a lake. This interpretation is strengthened by a comparison with the more extensive deposits and geomorphic features in the bottom of Spring Valley. Clark and Riddell (1920) were the first to show that there had formerly been a pluvial lake in this valley, and I have discussed the history of the southern part of the lake (Drewes, 1954, p. 82–84). Spring Valley was first occupied by a large pluvial lake that reached an altitude of at least 5,880 feet. As the water level dropped, the lake was divided by a gravel bar, now followed by the north highline road a few miles southeast of the quadrangle, into a restricted Spring Valley Lake to the north and a smaller Shoshone Lake to the south. Thereafter, the lakes fluctuated in level many times but continued to recede.

At present, some of the lowest parts of the northern basin contain ponds and both basins contain large marshes.

The higher bars and the wave-cut terrace at Rattlesnake Knoll were thus formed in the earlier, more extensive stage of the pluvial Spring Valley Lake, and the bars at and below an altitude of 5,780 feet are related to the younger, more restricted stage of that lake. The lake is assumed to have formed during a regionally cool moist period, and the high lake stage was probably contemporaneous with the high level of pluvial Lake Bonneville, of late Pleistocene age, which extended into Snake Valley to the east. If that assumption is correct, the lacustrine deposits are mainly of late Pleistocene age, though the youngest may have been deposited in early Recent time.

ALLUVIAL SAND AND GRAVEL

Unconsolidated gravel and minor amounts of sand and silt form low terraces along stream courses and are still being deposited. Because the surface gradient on these deposits is gentler than that on the older gravel deposits, the youngest gravel commonly lies below the upper levels of the older gravel deposits near the canyon mouths at the foot of the mountains but overlaps them in small alluvial fans away from the mountains. The extremely rough surface on the gravel is dissected by a drainage network that tends to follow the abandoned channels of the streams that deposited the gravel. The gravel contains only debris of local origin. The largest boulders are washed out of nearby terraces, rather than washed down from the bedrock exposures. The youngest alluvium is entirely of Recent age.

STRUCTURAL GEOLOGY

In order to attempt to explain the structural environment, origin, and development of structural features within the Connors Pass quadrangle, I must summarize the tectonics of the surrounding region. The Connors Pass quadrangle lies in the eastern part of the Basin and Range province. This province is generally underlain by 20,000–40,000 feet of pre-Tertiary sedimentary rock, which is overlain in many places by a great thickness of Tertiary volcanic and sedimentary rock. Scattered Mesozoic and Tertiary stocks intrude the pre-Tertiary rock. The region contains many high-angle faults trending almost north, some bounding the ranges, and low-angle faults. There are also noteworthy smaller structural features pertinent to the geology of the Connors Pass quadrangle.

The regional distribution of the major groups of sedimentary rocks indicates that a broad dome about 40 miles wide and 80 miles long lies in eastern White Pine County and the adjacent part of Utah (Drewes, 1960b); fig. 11. On the north, west, and south sides of this dome there are numerous complex low-angle faults, many of them near bedding planes, along which the apparent thickness of the strata is commonly reduced rather than increased. The dome and many faults near bedding planes are cut by several major high-angle faults, such as range front faults, not shown on figure 11. The major dome is flanked by six or eight minor domes, each 10–30 miles in diameter. Some low-angle faults lie on or near each small dome, and few such faults have been found away from the domes. The larger areas of the most intense low-angle faulting, where upper Paleozoic rocks are commonly brought down onto Lower Cambrian or even Precambrian rocks, lie closer to the core of the major dome, and the areas of less intense faulting are farther from the core of the major dome. Many of the larger intrusive bodies lie near the centers of the small domes.

The Connors Pass quadrangle lies on the southwest flank of a small dome extending along the east side of the central part of the Schell Creek Range and on the south end of another small dome in the Duck Creek Range, and it is in a zone of numerous low-angle faults. Most of the rocks here are moderately deformed, and in places they are severely deformed, chiefly by low-angle faults and steep normal faults but partly by tear faults and folds. The low-angle faults probably include both thrust faults and glide faults. The term "glide fault" is used here to describe faults formed as a direct result of gravity. At the time of movement, a glide fault plane reached the surface of the earth all around the allochthonous plate. In these respects a glide plate is a large variety of slump block. Low-angle normal faults, though also formed as a direct result of gravity, die out within the crust along their lower edge. Faults beneath slumps and glide plates may begin as low-angle normal faults along their upper edge, but become glide faults once the fault plane has been extended beneath the entire plate. The term "thrust fault" is used here to describe faults formed beneath plates that have been thrust laterally, either from some force within the crust or perhaps indirectly through gravity.

The structural features are of four general ages: Mesozoic(?), middle or late Tertiary, late Tertiary or early Pleistocene, and late Pleistocene or Recent. The scarcity of Mesozoic and early Tertiary rock prevents more precise dating of the structural features and

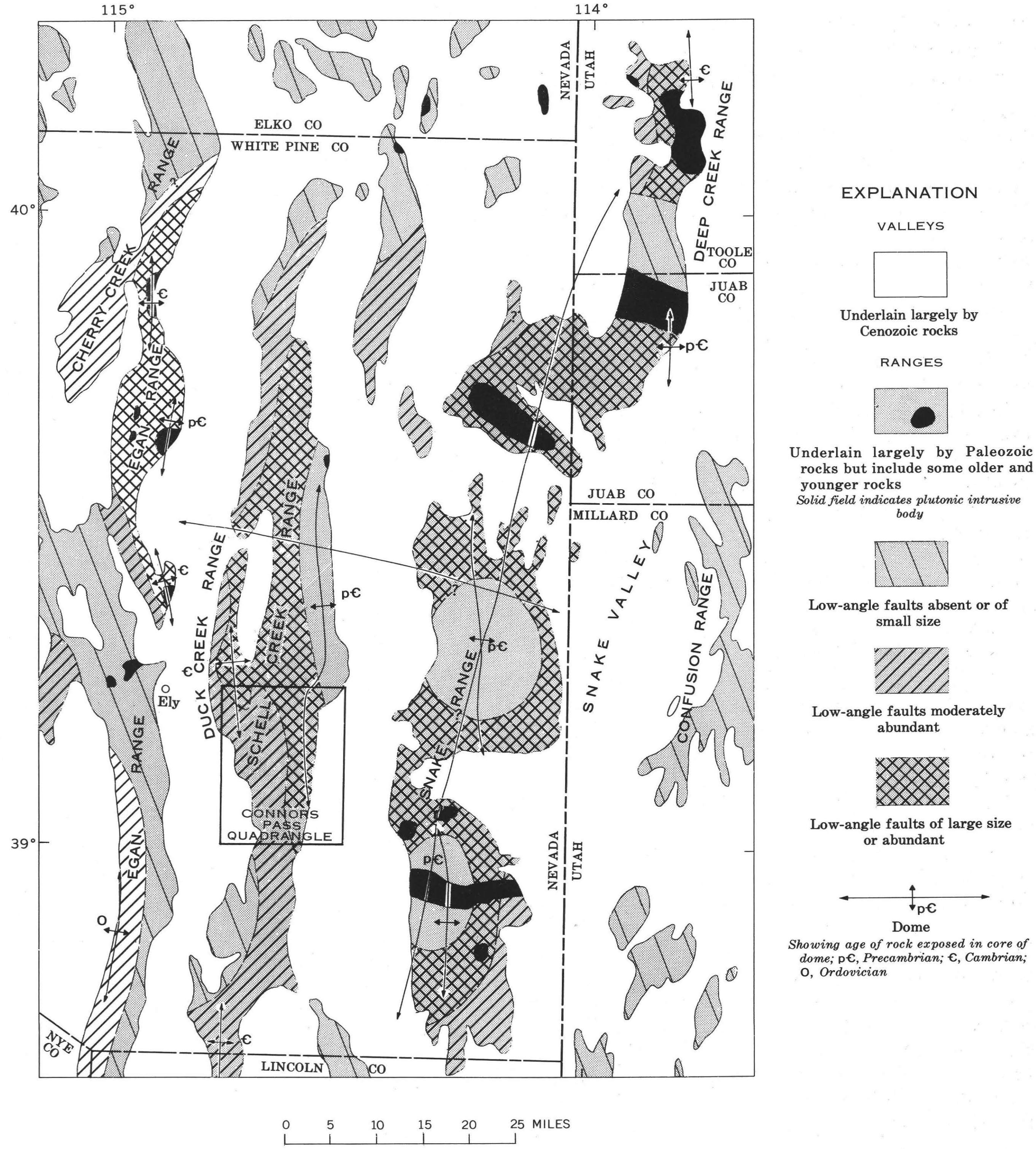

FIGURE 11.—Generalized tectonic map of the region around the Connors Pass quadrangle, showing the distribution of domes and low-angle faults. General stratigraphy adapted from McJannet (1960), Nolan (1935), Bick (1958), and Fritz (1960).

even causes considerable uncertainty regarding the ages assigned to some of them.

STRUCTURAL FEATURES OF MESOZOIC(?) AGE

A group of low-angle faults that deformed Permian rocks but not Eocene rocks is tentatively dated as Mesozoic. Similar faults in nearby areas were dated by Misch (1960, p. 33) as probably of Late Jurassic or Early Cretaceous age. I include in structural features of this age some faults that are geometrically similar to those of Mesozoic(?) age but of even more uncertain age, some broad arches, smaller faults, and folds within the low-angle fault plates. The porphyritic rhyolite dikes believed to be related to the monzonitic plutons of adjacent areas were probably intruded late in this period of deformation.

Among the oldest, and possibly the very oldest, structural features in the area are two broad anticlines, the southern noses of minor domes, striking about north and plunging gently southward, in the northern part of the quadrangle (fig. 11). The axis of the larger anticline runs through Cave Mountain and the crest of the Schell Creek Range, and that of the smaller one passes near Camel Peak between upper Steptoe Creek and Mosier Canyon. The formations as a whole, as well as the minor thrust planes, are arched over the axes, but the bedding planes, except perhaps in the oldest rocks, do not closely conform to the anticlinal structure. This apparent incongruity will be explained in the following section. Howell (1875, p. 240–247) was the first to recognize a quaquaversal structure in the southern Snake Range, and Spurr (1903, p. 44–47) mentioned a large north-trending fold in the Schell Creek Range.

The age relations between the minor domes and other structural features of Mesozoic(?) age are uncertain: It seems as likely, at first glance, that the doming warped the low-angle faults as that the faults followed bedding planes already warped. Inasmuch as the age relations between these features seem to be essential to the interpretation of the structure, the reasons for this uncertainty will be discussed in the section on low-angle faults (p. 74). Suffice it to say here that regional relations preclude my inferring that the faulting has been caused by the doming and hence that these features are contemporaneous.

SCHELL CREEK RANGE THRUST FAULT

The main low-angle fault in the Connors Pass quadrangle will here be referred to as the Schell Creek Range thrust fault. Its trace extends diagonally northwestward across the quadrangle from near Majors Place to Mosier Canyon (pls. 1, 2). A low-angle fault, whose structural relations are similar to those of the Schell Creek Range thrust fault and which may actually be an extension of it, appears 15 miles to the north of the quadrangle (Young, 1960a, b). The trace of the Schell Creek Range thrust fault, or a very similar one, recrosses the range 3–8 miles south of Majors Place. Spurr (1903, p. 44–47) recognized a large fault between Carboniferous and Cambrian rocks east of Connors Pass, and faults in this area were discussed by Misch and Easton (1954), by Drewes (1960a), and by Misch and Hazzard (1962, p. 319).

Within the quadrangle the Schell Creek Range thrust fault is crossed by some faults and is confluent with others. Between Majors Place and Cooper Canyon, a block of fanglomerate and volcanic rock younger than the Schell Creek Range thrust fault is faulted into its upper plate, and for about 3 miles the normal fault bounding this block is superposed on the trace of the thrust fault, and thereby interrupts the thrust fault. At Steptoe Creek, again, the trace of the Schell Creek Range thrust fault is interrupted for more than a mile by a graben, by which that thrust fault has been offset for several miles to the north, beyond the edge of the quadrangle.

The Schell Creek Range thrust fault is confined to a single well-defined shear surface where the rocks of the upper plate are relatively imcompetent, but where they are competent it is distributed among several subparallel shear surfaces. Stripped low-angle fault surfaces are exposed just north of Cooper Canyon and northwest of Grasshopper Canyon, and the fault is also well exposed along U.S. Highway 6–50–93. At the surface the fault dips about 25° SW.; this dip can be projected downward for 200–1,000 feet where there is considerable relief. Northeast of its trace, a restored surface of the main fault would presumably arch over the crest of the ridge, and about 1 mile southeast of the Kolcheck Basin dip downward beneath the small klippe or tectonic outlier of Chainman Shale and Joana Limestone. South of the mouth of the canyon of Cleve Creek, the fault probably also underlies a small block of Ely Limestone and Chainman Shale that has been further dropped along range-front normal faults.

The youngest rock in the lower plate of the Schell Creek Range thrust fault is generally the lower part of the limestone of Cambrian and Ordovician age. Near the fault it is slightly sheared, and in places it is mylonitized, recrystallized, and warped. The bedding of this rock is nearly parallel to the fault surface except for a few small folds, whose axial planes strike northwest and generally dip southwest. The rock at the base of the upper plate is most commonly the

Chainman Shale, but in places the Arcturus Formation and Ely Limestone lie directly on the lower plate, and locally there are a few slices of Guilmette Formation and even older rock above the fault. Farther from the trace of the fault, rocks older than the Guilmette Formation appear in the upper plate, and less rock appears to be missing along the fault. The rocks of the upper plate commonly dip into the fault plane, and their structural habit differs from that of the rocks immediately beneath the fault plane.

STRUCTURE OF THE LOWER MAJOR THRUST PLATE

In the lower major plate, low-angle faults of various kinds are more plentiful than steep normal faults, and they become less common away from the major fault. Northeast of Cave Creek most of the low-angle faults separate formations or separate beds of different character within formations. In general they are almost parallel with bedding, and the rocks in the minor fault plates are mostly in their normal stratigraphic order. Locally, however, parts of the formations, or even entire formations, are cut out along minor low-angle faults; in other places, beds in adjacent parts of minor plates are parallel to each other but dip more steeply than the intervening fault. On the 10,300-foot peak 3 miles north of Cave Mountain, for example, the fault contacts between formations are nearly horizontal, but the beds of each formation dip 30° or more to the east. The steeper dips occur on the east side of the peak, where the section is least complete. A stratigraphic section normally about 5,000 feet thick is here represented by fault slices only 1,000 feet in total thickness.

On plate 2 the larger of the minor low-angle faults are lettered consecutively in ascending order from A to M. A few faults that bifurcate are designated by numerical subscripts. Fault C, for example, splits to form faults C_1 and C_2, C_1 being the lower. Although the minor faults along some horizons are discontinuous, their separate segments are given one letter where the faults seem to be potentially continuous (had the displacement along the fault been greater) or to join up or down the dip. For instance, fault A, which is within a thin shaly unit between thick massive units, cannot be continuously traced, but I suspect that its separate segments are joined beneath the surface or were once joined in the rocks now eroded away. I also describe some of the minor faults as smaller or larger than others and thereby loosely compare their prominence. "Smaller" and "larger" need not be synonymous with either amount of displacement or length of the fault, but may be a combination of both parameters.

The minor low-angle faults differ greatly in extent, and so do the thicknesses of the rocks that are cut out by them. These variations can be illustrated by examples arranged in a gradational series. The most extensive minor low-angle fault in the lower major plate, fault C, follows the base of the limestone of Cambrian and Ordovician age. Generally very little of the Lincoln Peak Formation is missing beneath this fault, but near Bastian Spring the entire formation is faulted out. Along the upper part of Steptoe Creek, northwest of Grasshopper Canyon, and south of Cooper Canyon, fault C follows the same horizon above shaly rocks and beneath relatively massive limestone, but north of Cooper Canyon and in the Cleve Creek area it splits into two branches (C_1 and C_2), rarely more than 200 feet apart. The rocks between C_1 and C_2 are alternating shaly limestone and moderately thick bedded limestone, intermediate in competence between the shale and the limestone adjacent to fault C. Fault C does not split further south, perhaps because rocks of intermediate competence are there absent along this horizon, owing to a facies change. Faults C_1 and C_2 can be recognized where they are separated by rocks of this transitional character that dip more steeply than either of the faults. The bedding in the shaly rocks below C_1 and that in the relatively massive limestone above C_2 also converge with these faults. Fault C, or C_2, is exposed in many places; fault C_1 is covered by colluvium, but there are abrupt changes of attitude and lithologic character in the rocks along its course. Faults C_1 and C_2 appear to be connected at several places by minor low-angle faults, and presumably other cross faults have been overlooked.

Faults D and E are as long as fault C, but inasmuch as the thicknesses of the rocks missing along them are less, they seem to be somewhat smaller. Fault E follows the base of the Fish Haven Dolomite where the Eureka Quartzite is missing, and in places it splits into E_1 and E_2, which are separated by thin sheared lenses of the quartzite. This fault brings rocks that are probably low in the Fish Haven Dolomite against rocks as low in the normal stratigraphic sequence as the limestone beneath the olive-gray clay-shale member of the limestone of Cambrian and Ordovician age. To the south, fault D converges gradually with fault E, but to the north, near the Kolcheck mine, fault E cuts fault D at a fairly large angle, this relation indicating that movement on fault D had ceased before movement along fault E began. Other very local minor bedding-plane faults may follow the base of the limestone beds above the olive-gray clay-shale member.

Faults F and B are much less conspicuous than those described above. Fault F extends for almost 2 miles between Cave and Cleve Creeks, where it brings rocks low in the Simonson Dolomite against the Sevy Dolomite and in places against the Laketown Dolomite. Along fault B, which is more extensive than fault F, only a thin sequence of rocks is missing. Near the mouth of Cooper Canyon, rocks probably low in the Lincoln Peak Formation are faulted across the upper several hundred feet of the Pole Canyon Limestone. The lowest shaly beds along the fault are sheared; they weather to a reddish-brown soil.

Misch and Hazzard (1962) considered fault B a décollement fault in the vicinity of Majors Place. (See also Misch and Easton, 1954; Drewes, 1960a; Misch, 1960.) Fault B, however, is only one of many minor thrust faults almost parallel with bedding planes. It is neither the lowest nor the largest, nor does it lie at the base of shingled thrust plates. Fault B, therefore, seems no different from the many other similar faults in the area. Considering our present knowledge of the area, I suspect that the application of the term "décollement fault" to any fault exposed in the Connors Pass quadrangle would detract from the appeal that décollement faulting has as a regional hypothesis.

The lowest and least extensive mapped minor bedding-plane fault, A, cuts out all of the Pioche Shale about a mile north of Majors Place, but only a part of it 3 miles north of Majors Place and also just south of Cleve Creek. On the south wall of Mosier Canyon, in the northwest corner of the quadrangle, vertical and slightly overturned beds of Prospect Mountain Quartzite and Pioche Shale indicate local movement on this fault along the shale and perhaps also movement along a lower horizon within the quartzite. About 2 miles north of the mouth of Cooper Canyon, structural details along marker bed a of the Lincoln Peak Formation show very minor bedding-plane faults within the formation in addition to the mapped faults. In this place, shale beds adjacent to the mapped limestone marker beds have been disharmonically folded along the contacts with the limestone beds.

The lower major plate is broken by one group of high-angle tear faults about a mile north of Cave Mountain and also along the main ridge about a mile south of the north edge of the quadrangle. These faults strike northeastward, and downdip they join minor low-angle faults above the Lincoln Peak Formation (pl. 3). Rocks of Middle Ordovician to Devonian age, which are generally missing along the trace of the Schell Creek Range thrust fault, are present between the tear faults. Perhaps they were preserved beneath the Schell Creek Range thrust fault because within the lower major plate they occupy a structurally low position between the tear faults.

Most of the high-angle faults that cut the rocks of the lower major plate are normal faults that strike north to northeast, and many end upward, downward, or in both directions against minor low-angle faults. The normal faults along the upper part of Cleve Creek are truncated upward by the lowest of the minor low-angle faults present there, fault C. With the possible exception of these, the normal faults appear to have relieved stresses between blocks within the minor low-angle fault plates. Many other minor normal faults probably cut the rocks, but because of the present lack of detailed stratigraphic control it has been impossible to map them. The difficulty of recognizing and mapping subdivisions of the Cambrian and Ordovician limestone is due to the many small normal faults within it. The widespread convergence of the strata with the minor low-angle faults is probably caused in part by the rotation of segments of minor thrust plates between these small normal faults.

The Lincoln Peak Formation and the limestone of Cambrian and Ordovician age have been deformed in places by folds a mile or more in length that have gently dipping limbs and north- to northwest-trending horizontal axes. These folds are disharmonic features, in that no one of them affects more than one formation. Smaller folds, which are asymmetric and rarely more than 30 feet in amplitude are common in the Lincoln Peak Formation north of U.S. Highway 6–50–93. Their axes trend northwest and are horizontal or plunge gently either way, and their axial planes dip 10°–60° northeastward. They are disharmonic with respect to the fault contact at the top of the Lincoln Peak Formation. Other small folds occur throughout the lower major plate.

STRUCTURE OF THE UPPER MAJOR THRUST PLATE

In the upper major plate, above the Schell Creek Range thrust fault, steep normal faults are more numerous than minor low-angle faults, and folds are relatively scarce. The normal faults differ more widely in strike than those in the lower major plate, and only locally do they form regular patterns. For instance, a roughly arcuate pattern is formed by widely spaced normal faults high in the upper major plate southeast of Connors Pass. Several small northeast-trending faults cut the Joana Limestone near the junction of Cave Creek and Steptoe Creek and die out in the overlying Chainman Shale and the underlying Pilot Shale. The Joana is cut by small faults in many other places, apparently in response to stress adjustments between the incompetent shale formations adjacent to the limestone.

Most of the minor low-angle faults in the upper major plate are near the Schell Creek Range thrust fault. Fault H follows the base of the Guilmette Formation on both sides of lower Steptoe Creek. Rocks in the lower part of the Guilmette Formation there lie either on the Sevy Dolomite or on the lower part of the Simonson Dolomite. On the west side of Tamberlaine Canyon, the Simonson Dolomite immediately beneath fault H dips steeply and is truncated by a thin limestone unit of member a of the Guilmette Formation. The relations are less obvious, however, on the large hill a mile and a half south of the mouth of the canyon of Steptoe Creek, where rocks of member a of the Guilmette Formation lie immediately above, and parallel to, beds high in the Sevy Dolomite. This contact has been mapped as a fault rather than an unconformity, which it also resembles, because no regional unconformity is known at this horizon and because local bedding-plane faults are abundant.

The Pilot Shale is also thinned in several places by minor faults almost parallel to bedding. Just south of Tamberlaine Canyon a group of three such faults, the middle and probably the most extensive being fault I, brings a thin unit of tilted Joana Limestone close to beds in member a of the Guilmette Formation. These rocks are similarly faulted together along the lower part of Steptoe Creek, but there the low-angle faults are displaced by many normal faults. Farther south the deformation along fault I gradually becomes less marked, as Pilot Shale is brought onto different members of the Guilmette Formation.

Along some of the minor low-angle faults in the upper major plate, beds have been repeated rather than cut out. In the large hill 3 miles south of lower Steptoe Creek and west of Cooper Wash, dolomite of the Guilmette Formation has been moved on fault J over other apparently younger dolomite in that formation and over the Chainman Shale (pl. 1, section *M–M'*). The small plate above this fault is bounded on the northwest by a northeast-trending tear fault. Structural details along the south end of this plate are obscured by strong local dolomitization and consequent blurring of stratigraphic details within the Guilmette Formation.

Higher in the upper major plate a low-angle fault, K, extends along, or just beneath, the contact of the Chainman Shale with the Ely Limestone, which has been so much eroded that only scattered outliers remain. The fault plane is nowhere exposed, for the shale is largely covered by limestone debris, and the shale is commonly slumped. Faulting is inferred because limestone of various faunal zones lies directly on the shale, and because over large areas the limestone dips into the upper part of the shale. In some places as much as 1,500 feet of the lower part of the limestone is missing along fault K, but elsewhere little or none of it is missing. This relation is well shown at the outcrop of limestone on a ridge 1 mile north of The Narrows on the lower Steptoe Creek (pls. 1, 3). On the southwest side of the hill the contact seems to be normal, for rocks characteristic of the basal part of the Ely Limestone lie conformably on rocks presumably at the top of the Chainman Shale. The rocks dip 30°–45° northeastward, the steeper dips being in the northeast. *Chaetetes*-bearing beds of Pennsylvanian age and the silty limestone containing fusulines of Permian age lie in their normal stratigraphic positions, about 1,400 feet and 2,000 feet, respectively, above the base of the Ely. The ridge crest thus consists of Permian rocks, but only 200 feet topographically below and northeast of it, downdip from some of the Permian rocks, the Chainman Shale reappears. Around the northwest and southeast ends of the ridge the beds converge northeastward with the nearly horizontal fault contact at the top of the shale. On three sides of the hill the fault movement is concentrated at the base of the limestone, as shown by fault K. On the southwest side of the hill, however, most of the movement lies at a slightly lower horizon within the shale, fault K2, and a little movement may also have followed fault K1 without leaving any conspicuous trace.

In a block of Ely Limestone near the Aspen Spring in the center of the quadrangle, the beds dip as steeply as 65° NE. (not shown on map) and over large areas the same beds dip 40° NE., where the contact with the Chainman Shale dips about 5° SE. A similar structural relation exists around the flanks of the limestone forming the peak at Taylor bench mark (fig. 8).

Other small masses of Ely Limestone lie on or within the Chainman Shale. Those not associated with slump topography and not in positions that could have been part of a more extensive thrust plate or glide plate are regarded as blocks that moved along unmapped faults within the shale. Some of these blocks may even have been engulfed in the highly plastic shale at times during which the rocks were under great stress, or they may have been slump blocks of Tertiary age. In several places the shale has been plastically deformed near its contact with the limestone. Just south of Cooper Canyon and near the center of the drainage basin, slightly overturned Chainman Shale lies on a large body of Ely Limestone. Along the highway a mile southwest of Connors Pass, Chainman Shale has been squeezed up into the limestone. Such plastic deformation is greatly facilitated by the gypsum and high clay content of the shale.

In the southwestern part of the quadrangle, signs of structural deformation along the contact between the Chainman Shale and Ely Limestone are fewer and less impressive or are masked by high-angle faults. Two miles southwest of Connors Pass, however, just north of the highway, some beds of limestone and conglomerate in the Chainman Shale are upended beneath the base of the Ely Limestone.

On the south wall of Cooper Canyon just west of the trace of the Schell Creek Range thrust fault, the Arcturus Formation seems to be thrust along fault L over the Ely Limestone. The Ely Limestone has fairly steep to vertical dips, and the Arcturus Formation gentle to moderate dips, whereas the contact between the formations seems to be nearly horizontal. East of this locality the Ely Limestone wedges out between fault L and the Schell Creek Range thrust fault.

In a downfaulted block high in the upper major plate, the Ely Limestone is thrust over the Rib Hill Sandstone and the Arcturus Formation on fault M, which is gently arched from northeast to northwest, just north of Connors Pass. Fault M is not exposed because the rocks immediately beneath it are incompetent and the slopes are covered with debris of the Ely Limestone. About a mile northeast of the pass, Pennsylvanian *Chaetetes*-bearing rocks in the middle of the Ely Limestone lie a short distance above Permian rocks, some of which may be as young as the gypsiferous Lower Permian rocks in the Arcturus Formation.

Fossil determinations made after the completion of the fieldwork indicate that some beds low in the Permian part of the Ely Formation were mistakenly assigned to the Arcturus Formation in the Connors Pass area but, regardless of this possible miscorrelation, known Pennsylvanian rocks there have moved over Permian rocks. To the west, fault M is cut out by a high-angle normal fault that brings rock high in the Ely Limestone overlying fault M in contact with the middle part of the Ely Limestone overlying fault K. Thus, ½–1 mile west of Connors Pass the Chainman Shale is exposed in part of a fenster through fault K, and the Rib Hill Sandstone adjacent to the shale is exposed in part of a fenster through fault M (pl. 1, section *O–O′*). The Chain Shale and Rib Hill Sandstone are here separated by narrow slices of brecciated Ely Limestone.

In the southwestern part of the quadrangle, three broad gentle folds extend northeastward and are 1–2 miles long. Smaller folds occur at a few places in the Chainman Shale, their axes striking northwest and their axial planes vertical or dipping steeply to the southwest.

In both of the major plates, the small folds are probably drag folds formed during the time or times of movement along the low-angle faults. The axes of the drag folds almost all strike north to northwest and are horizontal or gently inclined, but the direction of dip of the axial planes of the folds is less systematic. In the lower major plate, somewhat more than half of the axial planes dip southwestward, and the acute angle between the fault plane and the axial planes of the drag folds points northeastward, this relation indicating that the upper plate moved relatively northeastward. The axial planes of many other drag folds are at right angles to the fault planes, so the direction of movement is equivocal, and the axial planes of a few drag folds in the lower plate dip northeastward. In the upper major plate, however, the axial planes all dip southwestward, this dip indicating that the upper plate on each of the minor thrusts in that plate moved relatively northeastward. Evidence presented by these structural features, as well as by the northeast-striking tear faults, shows clearly that the upper plate moved relatively northeastward or southwestward on the main fault, but the actual direction of movement is somewhat uncertain. Perhaps there were even two different directions of movement at different times.

STRUCTURAL FEATURES OF MIDDLE OR LATE TERTIARY AGE

A second group of features comprises normal faults and glide faults that cut rocks as young as those of Oligocene(?) age and that also cut the Schell Creek Range thrust fault. Normal faults of this group trend north or alternately north and northwest, and some of them are paired to form the sides of grabens. These faults are less numerous, but of larger displacement, in the northern part of the quadrangle than in the southern part. Where normal faults of large displacement join subparallel thrust faults without deflecting them, as they do east of Cooper Summit and south of Cooper Canyon, the normal fault surfaces are believed to have followed thrust-fault surfaces downdip for a moderate distance.

The largest graben, Duck Creek graben, lies between the Grasshopper Canyon fault and the Steptoe Creek fault (pl. 2) and is at the south end of a topographic and structural basin, about 25 miles long and 5 miles wide, between the Duck Creek Range and the main part of the Schell Creek Range. The geology at the north and south ends of the basin is probably more indicative of the rocks and structure beneath the alluvium in the basin than is the geology on the slopes to the east or west. South of this basin the Grasshopper Canyon fault dips steeply eastward, and toward the south its stratigraphic displacement decreases and even reverses. Also south of the basin, the Steptoe Creek fault consists of several closely spaced west-dipping faults, most of which have only moderately steep dips. The south end of this fault zone turns eastward along Cave Creek and follows a segment of the older Schell Creek Range thrust fault. Although rocks as young as Permian are dropped in the graben between rocks as old as Late Cambrian, the stratigraphic displacement along the faults adjacent to the graben is not as large as it might seem, for much of the stratigraphic section is cut out by the older low-angle faults. Inasmuch as the Duck Creek graben topographically resembles many of the larger valleys of the region, the structure of the rocks within it may be similar to that of the rocks under Spring Valley and Steptoe Valley; therefore, the features at the ends of the Duck Creek graben might give us some clues to the structure underlying those valleys.

South of the Duck Creek graben are several smaller grabens that contain sedimentary or volcanic rock of Tertiary age. The fault bounding the east side of a small graben between Cooper Canyon and U.S. Highway 6–50–93 has been superposed on the trace of the Schell Creek Range thrust fault for about 3 miles. It probably truncates this fault, as projections of the faults downdip indicate (pl. 1, section *O–O′*), but a part of the normal fault probably coincides with a steeply dipping part of the thrust fault. Some of the faults along Cooper Wash and near the Taylor mining district are tentatively grouped with these Tertiary normal faults because of their association with small grabens.

Between Cooper Canyon and the central part of Steptoe Creek, low-angle faults underlie two plates of Tertiary conglomerate and volcanic rock. The remnant of the fault plate south of Cave Creek covers about 8 square miles and consists of at least 5,000 feet of conglomerate, lava flows, tuff, and welded tuff. The one north of Cave Creek is much smaller, consists of a thinner sequence of similar rock, and is largely covered by the older alluvial and fan gravel. Both plates rest mainly on the Chainman Shale, but small parts of them rest on or against blocks of Ely Limestone, and part of the east edge of the large fault plate rests on Cambrian and Ordovician limestone.

Although the contacts between the Tertiary rock and the older rock are covered and the rocks near the contact indicate little of the local structure, these contacts are inferred to be fault contacts. There must be a fault contact between the eastern half of the larger body of Tertiary rock and the Paleozoic rock because the overlying Tertiary rock dips eastward toward the adjacent exposures of older rock. The trace of the contact around the eastern half of this body is deflected across spurs and around ridges in such a manner as to require a fault surface that dips gently inward like the edges of a saucer. The contrast in style of deformation between the Tertiary and the Paleozoic rocks also indicates the nature of the faulting. The rock above the fault is gently warped into a broad syncline, which plunges eastward toward the underlying fault surface, but the Tertiary rock itself is cut by relatively few faults. The Paleozoic rock, however, is intricately faulted and is sheared along faults near bedding planes. These contrasts indicate that the deformation of the Tertiary rock and that of the Paleozoic rock reflect wide differences of stress and environment. The plates of Tertiary rock resemble those that moved along the Turtleback faults in the Death Valley region of California (Drewes, 1959), except, of course, that their lower surfaces are concave rather than convex upward. The plates of Tertiary rock also resemble large slump blocks in that they are rotated about a north-trending horizontal axis and dip eastward and in that they appear to rest in part on a saucer-shaped surface. These comparisons and contrasts suggest that the fault beneath the plate of Tertiary rock was formed near the surface, as a glide fault. This interpretation is strengthened by the fact that the Tertiary rock most commonly overlies the plastic Chainman Shale, which must always have underlain topographically low areas and which would be an ideal lubricant for either a large slump block or a glide plate, two structural features differing only in size.

If this interpretation is accepted, the low-angle fault must be inferred to continue between the Chainman Shale and the main body of Tertiary rock to the west, for the rotation of the eastern part of the plate could only be effected by a shift of the western part, where the Tertiary rock is more nearly conformable with the contact beneath it.

One minor complication around the larger plate of Tertiary rock must be considered here. Along the road in the upper part of Cave Creek Canyon, conglomerate in this larger plate overlies a small block of Ely Limestone without signs of faulting. The block of Ely Limestone possibly was dragged along the base of the plate of Tertiary rock on the shale immediately beneath the limestone.

Some of the other Tertiary rock within the Connors Pass quadrangle also seems to rest on faulted surfaces, but other Tertiary rock merely rests on unconformities. The body of Tertiary rock north of Cave Creek resembles the larger one in every respect except size and is therefore also inferred to be the remnant of a glide plate. Inasmuch as the Tertiary rock south of Cooper Canyon is bounded by high-angle faults and does not dip into underlying Paleozoic rock, it is not inferred to rest on a low-angle fault, even though the conglomerate there is gently warped like that in the bodies regarded as remnants of a glide plate. In the northern part of the Schell Creek Range, Young (1960b, p. 167–168) reported low-angle faults resembling the one here described.

STRUCTURAL FEATURES OF LATE TERTIARY OR EARLY PLEISTOCENE AGE

Structural features of middle or late Tertiary age were cut by normal faults, apparently before the deposition of upper Pleistocene gravel. Several normal faults that lie wholly within the Schell Creek Range cut the low-angle fault and the overlying plate of Tertiary rock south of Cave Creek. The upper part of the largest of these faults probably was active during movement of the glide plate of Tertiary rock, for it offsets the rocks above the glide fault more than it does the glide fault itself. Renewed normal faulting occurred along the existing fault plane within the glide plate and extended below the glide plate, so at present the rocks within the glide plate are offset more than those beneath the glide plate. Near Cave Creek the later movement along this largest normal fault probably extended along the Steptoe Creek fault, which had already been active during early or middle Tertiary time. In other words, there was recurrent movement along some, at least, of the segments of the normal fault along upper Steptoe Creek, Cave Creek, and Cooper Canyon. It is noteworthy that the low-angle glide fault beneath the Tertiary rock occurred between times of other large crustal movements.

The only normal fault along either front of the range is in the northeast corner of the quadrangle. Near the mouth of the canyon of Cleve Creek this fault splays out toward the south, and branches of it extend into Spring Valley as well as into the range. North of the mouth of the canyon the fault is buried but probably lies close to the foot of a prominent scarp, which continues at least 25 miles northward beyond the quadrangle. The latest movement along this fault probably occurred in late Quaternary time, for the younger alluvial and fan gravels are crossed by dim fault scars.

A normal fault—the Taylor fault (pl. 2)—trends alternately northward and northwestward close to the west front of the range. Rocks that are mainly of late Paleozoic age but include the older alluvial and fan gravel of Pleistocene age are faulted down to the west along the northwest-trending segments of this fault, which seems to truncate some of the older grabens. The north-trending segments probably followed some of the older normal faults. According to A. L. Brokaw (oral commun., 1962), another large young normal fault, downthrown to the east, trends northwestward along the east flank of the Egan Range and cuts obliquely across part of it near the copper pit at Ruth. The northwestward trend of a part of Steptoe Valley—roughly between the latitudes of Connors Pass and Ely—seems to have been controlled or influenced by a segment of a major graben lying between the large fault in the Egan Range and the Taylor fault.

Normal faults probably underlie the low scarplets in the fan gravel several miles west of the Schell Creek Range in the southwestern part of the quadrangle. Because these scarplets follow the contours closely, they superficially resemble the little wave-cut cliffs in Spring Valley, but the scarplets in Steptoe Valley are not associated with lacustrine gravel. A preliminary gravity survey across Steptoe Valley west of the Taylor mining district by D. R. Mabey (written commun., 1962) showed a locally steep gravity gradient, suggestive of a fault, alined with the scarplets.

In Grasshopper Canyon a large body of shattered or brecciated, but not strongly rotated or mixed, Ely Limestone and Chainman Shale rests partly on more of the Chainman Shale and Ely Limestone and partly on Cambrian rocks. Where the bedding is preserved, it dips 30°–55° eastward—at least twice as steeply as similar rocks in the adjacent ridge. This body of shale and limestone is believed to be a large slump block, or possibly a landslide, and it differs from the numerous recent slump blocks in its relation to adjacent topography. It deflected the course of the wash in Grasshopper Canyon about 500 feet to the west, and thereafter a canyon about 100 feet deep was cut into the distal end of the body of shale and limestone. This amount of canyon cutting indicates that the body of shale and limestone probably moved into its present position at least as long ago as Pleistocene time.

STRUCTURAL FEATURES OF LATE PLEISTOCENE OR RECENT AGE

Slump blocks are given special attention here as structural features resting on low-angle faults, rather than as topographic or weathering features characteristic of shaly formation, because the area contains many somewhat similar but larger structure features. The term "slump block" is used here to describe a variety of landslide in which a mass of rock has moved relatively slowly as a discrete body over a fault surface. Material in a debris slide, another variety of landslide, is shattered, generally because it has moved more rapidly. Some landslide masses have features intermediate between those of slump blocks and debris slides.

Most of the slump blocks are underlain by Chainman Shale, but some are underlain by the Lincoln Peak Formation. Areas marked by prominent landslide topography total about 2 square miles; they are outlined on plate 1 by a fault contact and are distinguished from other low-angle fault blocks by an overprint pattern. In areas of slight local relief, the slump blocks are characterized by an irregular hummocky topography, but where the local relief is greater they make a rudely terraced topography, in which many of the terrace surfaces slope gently in a direction opposite to the general slope. Springs issued from the lower edges of several of the areas of slump blocks. The slumped material consists mainly of shale similar to that over which the slump block has moved, but in places it also contains fragments of the more competent rocks overlying the shale. For instance, blocks of Ely Limestone as much as several hundred feet in length seem to be rafted on internally disturbed and weathered Chainman Shale. The slump blocks are believed to be of Pleistocene or Recent age, because they are closely related to the present topography in areas of moderate relief. That some downslope movement is currently in progress is indicated by locally steep frontal lobes, by small depressions, and by disrupted vegetation.

ORIGIN OF STRUCTURE

The stress pattern responsible for the structural features in the Connors Pass quadrangle is not fully known, especially for those that extend beyond the borders of the quadrangle, for too little is known of the surrounding region—particularly of the rocks under the floor of the alluviated basins—to form a complete picture of the geometry of these major structural features; very few descriptions of the geology in this region recorded the observations on which conclusions regarding directions of movement on faults were based. The available evidence within the quadrangle is not sufficiently diagnostic for me to determine the origin of some of the low-angle faults, and that evidence is distorted by recurrent and diverse movement along some of the faults.

The main structural problem in the eastern Great Basin is the origin of the low-angle faults. Thrust faults are reported in most mountain ranges, but glide faults also are recognized in some places. Ideally, thrust faults are formed as the result of tangential compressive stresses relatively deep in the earth, whereas glide faults are the product of gravitative forces near the surface of the earth. In practice, the distinction between glide faults and thrust faults has never been a simple one to make, because in many places the rocks involved were probably subjected to both gravitational and laterally transmitted compressive stresses, which varied widely in relative importance. Where irregular glide surfaces introduce local compressive stresses or where glide faults form downslope in front of thrust faults, field evidence may be apparently conflicting. Local field evidence may not be diagnostic, and even in more thoroughly mapped areas than the eastern Great Basin it may be largely indicative rather than conclusive.

Even in areas where field evidence is not ambiguous, listing of rigid criteria for distinguishing the various kinds of low-angle faults is difficult. General evidence more strongly indicative of thrust faulting than of glide faulting includes persistent and widely distributed signs of compression well within the fault plates. Just how uniform, how strongly developed, and how extensive the compressional features must be is not clear; it apparently depends on environment—such factors as type of sediment, depth of burial, rate of deformation, and hydraulic conditions. Recognition of a root area is accepted as evidence of thrust faulting, but there is no general agreement about the criteria for recognizing a root area. A fault that plunges downward in the direction from which the upper plate moved is commonly regarded as a thrust fault, but because the configuration of many glide surfaces is irregular, the downward plunge must persist for a long distance (another vague quantity) if it is to be regarded as evidence of thrust faulting.

The presence of a gentle slope down which a plate has moved may be regarded as evidence, though perhaps only permissive evidence, of glide faulting. Apparently the larger the glide plate is, the gentler the slope can be, for the slope under a large plate must be extensive, and thus the total drop across the slope must be considerable (still another vague quantity). Perhaps, also, great incoherence in the faulted rocks and conflicting evidence of direction of movement indicate glide faulting rather than thrust faulting. This brief review of some salient features of glide faults and thrust faults shows how difficult it often is to distinguish between them. To reconstruct the conditions under which the low-angle faults in the Great Basin were formed—to identify slopes of regional extent for glide faults or root zones for thrust faults—is exceptionally difficult, because the low-angle faults are cut by younger high-angle faults and because relatively narrow mountain ranges are separated by broad belts of surficial deposits. Parts of all the low-angle faults dip downward in one direction or another, but thus far no long low-angle fault in the eastern part of the Basin and Range province has been so adequately mapped throughout its length as to show how extensive a particular dip is or to reveal clearly the fault's age relation to the folds associated with it. For these reasons and others, correlation of faults between ranges has not been conclusive.

I believe that the Connors Pass area contains some thrust faults and some glide faults and that it has undergone regional uplift and normal faulting. Some of the low-angle faults may have begun as thrust faults and later may have become glide faults.

GLIDE FAULTS

Some flat-flying faults are clearly glide faults. These faults are overlain by rock masses that range in size from small slump blocks to glide plates many square miles in extent. The gravity origin of slump blocks is widely known from studies elsewhere; many are well exposed or have been drilled, and many have been studied during movement. In the slumped areas mentioned on page 73, the material overlying the small glide planes is broken into discrete blocks, which are tilted upslope, as indicated by the terraces and depressed areas. Such movement requires a concave local glide surface. Slump blocks are derived from areas a relatively short distance upslope from the present site of the blocks, for many blocks lie downslope from topographic embayments in rocks like those in the slump blocks. Some slump blocks are moving at the present time, others moved in the recent past, and still others, such as the slump or debris slide of shattered shale and limestone along Grasshopper Canyon and possibly some of the blocks of Ely Limestone included in Chainman Shale, moved during Pleistocene or early Recent time. Tectonic forces may have set the stage for the movement of the slump blocks by increasing the local relief and by enabling streams to undercut certain blocks and thus leave them inadequately supported, but the chief moving force has been gravity.

The plates of Tertiary conglomerate and volcanic rock adjacent to Cave Creek have some structural features in common with the slump blocks, these similarities indicating that their origin was similar to that of the slump blocks. Large discrete plates of these rocks were rotated heel down (pl. 1, section *M–M′*; pl. 3, fig. 12), and, although of several orders of magnitude larger than the slump blocks in the area, these plates are of relatively local extent; they are not klippen of more extensive thrust or glide plates. The lower surface of the larger plate at least is probably saucer shaped and concave upward. The plates lie at the foot of structurally and topographically high areas, which were uplifted before, as well as after, the plates came to their present position. They are much less broken than the typical slump blocks, perhaps because of their greater size and strength. The combined evidence then, though only permissive, suggests that these large plates were glide faulted from a position slightly higher than the summit of Cave Mountain. The rocks of the eastern part of the large plate were rotated about 40° as they moved downward and westward, thereby pushing the rocks of the western part, with little or no rotation, over a surface eroded on the Chainman Shale. If one assumes that the axis of rotation of the eastern part of the block was not shifted laterally and that the small pipe of quartz latite vitrophyre was intruded at right angles to the bedding of the host rocks, the pipe may be projected beneath the glide plate (pl. 3, section *E–E′*). The projected position of the pipe would then coincide with a north-trending normal fault, extending from NW¼ sec. 25, T. 15 N., R. 65 E., to SE¼ sec. 12, T. 14 N., R. 65 E., along which other igneous bodies were intruded. If, as seems probable, the axis of rotation was nearly stable, the distance of movement along the glide fault was about 3,500 feet.

Certain features in the large blocks of Ely Limestone widely scattered on the Chainman Shale indicate that they too are glide plates, though this evidence is not as convincing as is that in the plates of conglomerate and volcanic rock. The blocks of Ely Limestone considered as possible glide plates are those 1–2 miles north of The Narrows along Steptoe Creek, at Taylor triangulation mark, and near the Aspen Spring in the middle of the quadrangle, but not the smaller blocks already mentioned as constituents of slump blocks or those that extend along small high-angle faults. In these large blocks, the bedding makes moderately high angles with the low-angle faults beneath the blocks and the blocks are internally broken by faults, along which the beds have been tilted. This structure is possibly also illustrated at a lower horizon by the much-repeated Joana Limestone along Steptoe Creek (fig. 12); inasmuch as the Joana in these blocks generally dips to the east or northeast and the surface beneath the blocks is flat rather than saucer shaped, the blocks could be the base of a large glide plate derived from a single area to the east or northeast. The fact that there was more rotation of the beds in the blocks of Ely Limestone north of the peak at Taylor triangulation mark (near the glide plates of conglomerate and volcanic rock) than in those farther south also suggests a source area nearby to the northeast. These blocks may be remnants of one or several large glide plates that moved southwestward from a structurally high area, underlain by Cambrian and Precambrian rocks, about 10 miles to the northeast, but there also is some indication that they are remnants of the upper plate on a thrust fault.

THRUST FAULTS

Evidence bearing on the origin of the low-angle faults within the Paleozoic rocks is available within the Connors Pass quadrangle, as well as in the surrounding region, but neither approach leads to a conclusive interpretation of how they originated.

Two somewhat related lines of reasoning, based on observations within the quadrangle, have a bearing on whether the faults are thrust or glide faults, or possibly both. First, the distribution of structural features resulting from compressive stress and the uniformity of the direction of that stress indicate the minimum extent of the stress field associated with the faulting. Second, the differences in styles of deformation of the rocks above and below the Schell Creek Range fault suggest differences in the environment of deformation, which could lead to a hypothesis of thrusting on some low-angle faults and gliding on others.

In the Connors Pass quadrangle there is more evidence of compressive deformation in the rocks beneath the Schell Creek Range thrust than in those above it. For instance, between lower Cooper Canyon and U.S. Highway 6–50–93, compressive stress of at least local extent is indicated by the imbrication of small slices of the Lincoln Peak Formation and of the lowest limestone in the overlying Upper Cambrian and Lower Ordovician limestone (pl. 1, fig. 12). More extensive compression is indicated by widespread gentle folding throughout the Lincoln Peak Formation, and the consistent northward strike of these fold axes suggests that the compressive stress was fairly uniform. The tighter smaller folds close beneath the thrust fault were formed independently of the larger open folds; they are drag folds formed in response to very local and relatively intense shearing stresses. Only one suggestion of compressive deformation can be cited in rocks above the thrust fault. On the south side of the klippe at Taylor triangulation mark, part of the Ely Limestone (pl. 1, section N–N') has been repeated along two subsidiary low-angle faults. On the basis of compressive deformation within the quadrangle, then, the small low-angle faults

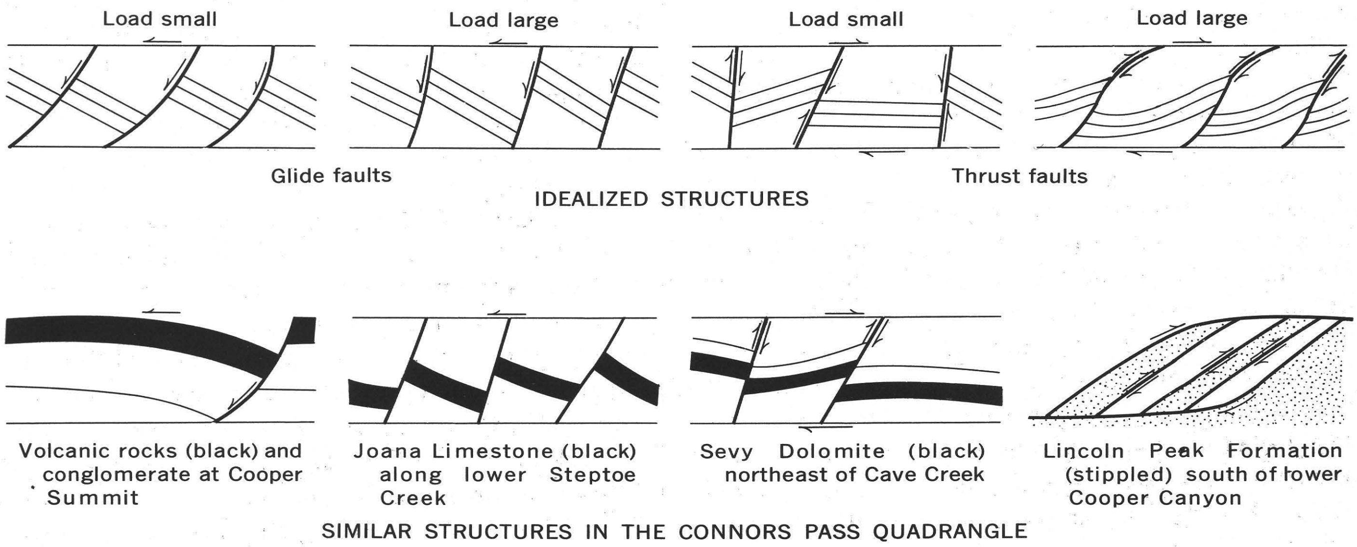

FIGURE 12.—Drawings showing comparison between idealized and actual patterns of fault deformation in rocks subjected to stress by overriding glide or thrust plate. Horizontal scale almost equal to map scale.

above the Schell Creek Range fault may be glide faults but the ones beneath it could be thrust faults.

The contrast in local styles of deformation between the rocks above and below the Schell Creek Range thrust may be due to differences of environment. The rock above the fault is not very strongly deformed and is broken chiefly by normal faults, many of which are younger than the thrust. Local evidence permits some of the low-angle faults that occur within the upper plate to be interpreted as glide faults, though some small plates may first have moved on thrust faults. The rock beneath the Schell Creek Range thrust fault is more strongly sheared and is locally drag folded and recrystallized. This difference in style of deformation cannot be explained by differences in lithology alone, for in some places the fault brings rocks of similar lithology but different degree of deformation in contact, and there are several bodies of incompetent rock in the upper plate as well as in the lower. However, the difference in style of deformation can be explained if the rocks beneath the Schell Creek Range thrust, and perhaps that fault itself, were deformed at much greater depth than that at which the rocks above the fault were deformed. These two environments of deformation may, but need not, indicate that the low-angle faults include both thrust and glide faults. When combined with the line of reasoning based on the extent of compressive stresses in the older rocks, the idea that there are both thrust and glide faults in the quadrangle has considerable merit, but convincing evidence is unavailable. Therefore, on the basis of local evidence alone I am sympathetic with DeSitter's (1956, p. 291) point of view that "* * * it is very much a question of taste how much one believes should be accounted for by thrusting and how much by gliding."

Regional evidence indicates even more clearly that stresses were uniform and widespread. The deformation is all of the same general style, and the direction of movement seems to have been east-northeastward in practically all areas. The uncertainty about the direction of movement implied by the words "seems to have been" is admittedly deliberate, for neither the abundance, the persistence, nor the location of the evidence with respect to particular low-angle faults is usually reported in the literature. In many nearby ranges in which there are abundant low-angle faults, the largest fault is not structurally the lowest fault, imbricated fault slices are usually absent or small, and younger rocks have typically been faulted over older ones. Although low-angle faults in adjacent ranges follow approximately the same stratigraphic horizons along incompetent rocks, correlation of specific faults from range to range has still to be demonstrated. The relation between the low-angle faults and the domes gives a further indication of the widespread distribution of the stress field associated with the faults.

The coincidence of much of the low-angle faulting with the top and flanks of the major dome in the northern Snake Range and the minor domes like those just north of the Connors Pass quadrangle (fig. 11) suggests a close genetic relation between domes and faults. The faults are not glide faults formed separately about the domes, for there is no radial direction of transport of the fault plates, and the most intense faulting seems to lie on the tops of the domes rather than on their flanks. Alternately, then, the genetic relation between the low-angle faults and the domes must involve a plate or plates of regional extent. It is unlikely that the domes formed after the low-angle faults, for the coincidence in their distribution can not be explained then. However, it does seem likely that the domes were formed before or during the time in which the regional fault plate moved, the domes being an obstacle beneath the plate above which the deformation was severest. Although it is interesting to point out here why the domes are among the oldest structured features of the region, the essential point is that the low-angle faults are regional and were propelled by sufficient energy to move over a number of fairly large obstacles.

A conventional approach to explaining the origin of low-angle faults of regional extent involves the identification of either a root area from which lateral stresses were transmitted, or a slope down which a glide plate could move, but both are conspicuously missing. In central Nevada there is no large upthrust or intrusive mass that might pass for a root area; likewise, there is no record of a structurally high area of sufficient magnitude, even accepting the gentlest of slopes, and of appropriate age to have been the source of a glide plate of regional extent. These unresolved difficulties indicate that perhaps some less conventional approach must be explored.

The difficulty in resolving the question of where the root area of the thrust faults is located may indicate a weakness in our concept of what a root area is like. The root of a thrust plate is commonly pictured as a simple piston of rock, as may be exemplified by an expanding magma chamber that pushes out against its host rock. Perhaps a root may also consist of a body of rock in which the pore pressure is equal to the lithostatic pressure, in the general sense used by Hubbert and Rubey (1959), and in which local lateral and vertical variations in pressure help to produce lateral movement of the rock as an indirect result of the load itself. Local variations in the pressure balance arise from variations in the porosity and other properties of the rock and in the amount of pore fluid present. This balance would also be affected by the rates of deposition or erosion of overlying sediments, as well as by gentle regional tilting or a push from a root area. Most of these following factors have varied throughout the region: lateral facies changes and metamorphic changes that would change the porosity of the rocks, rates of deposition during the late Paleozoic, and the rate of unloading of sediments.

In accordance with Misch's (1960, p. 33) view that the low-angle faults are Late Jurassic or Early Cretaceous, they must have formed under extremely high lithostatic pressure; for at the end of Early Jurassic time, the thickness of rock above the Precambrian was about 6.5 miles and that above the Cambrian about 5 miles. At such depths the lithostatic pressure and pore fluid pressure may well have been equal. Perhaps, then, some of the low-angle faults in the Schell Creek Range, which generally cut out beds and rarely repeated them, were formed in a root area. Conceivably, too, the zone in which pore fluid pressure balanced lithostatic pressure may have shifted upward as sedimentation continued. The rocks may thus have been squeezed out laterally at successively higher horizons, and the time during which they were being squeezed out could have been much longer than the time that would have been available if the faults were formed during a single orogenic period. If this hypothesis is correct for the rocks in the Schell Creek Range, or at least for the rocks of the lower plate of the Schell Creek Range thrust, the lithostatic pressure would never have been as great as the weight of the entire sedimentary column. Where relative ages are known in the Schell Creek and Snake Ranges, the low-angle faults at the higher stratigraphic levels are consistently younger than those at the lower stratigraphic levels. They cannot all be assigned a Late Jurassic or Early Cretaceous age unless they are assumed to have been formed during one orogenic period, but in accordance with the hypothesis of lateral squeezing only the youngest fault need be of that age. Although these unconventional ideas do not fully explain the origin of the low-angle faults, they do point to some approaches that may have fewer shortcomings than the usual hypotheses regarding thrusts and glide faults of this region; thus far our facts are few and our assumptions many.

REGIONAL UPLIFT AND NORMAL FAULTS

The regional uplift or uplifts recorded, in the Schell Creek Range and in most other parts of the eastern Great Basin, by the major angular unconformity between Paleozoic marine rocks and Cretaceous or Tertiary continental rocks caused an enormous change in the regional structural environment, from the great depths at which the thrust faults appear to have formed to near the surface. The cumulative uplift amounted to many thousands of feet—in places to tens of thousands of feet—the thickness of the Paleozoic and Mesozoic rocks that have been removed by erosion. No evidence of the cause of this major uplift has been found in this quadrangle, but a relaxation or cessation of the forces involved may have produced the normal faults of Tertiary and Quaternary age and permitted the intrusion and extrusion of magma. Inasmuch as the normal faulting and volcanic activity were recurrent, the forces producing tension in the crust must have been discontinuous or most have varied in intensity. Some of the low-angle faults, at least those beneath the Tertiary rocks, are believed to have been caused by gravity in areas of high relief adjacent to Tertiary normal faults.

ECONOMIC GEOLOGY

Scattered mining districts in the region around the Connors Pass quadrangle have produced chiefly silver, tungsten, gold, lead, zinc, manganese, and beryllium, but few mines are now being operated in the region other than the one in the large porphyry copper body at Ruth, in the Egan Range (fig. 1). Much of the mineralization lies in or near granodiorite or quartz monzonite stocks and related dikes, though not all of these intrusive bodies are associated with mineralization.

Several small areas within the quadrangle are mineralized, but aside from the Taylor mining district the duration of mining activity in them was brief. Silver was the chief metal produced, but small quantities of gold, copper, lead, zinc, antimony, and tungsten were also extracted, and barite and fluorite were locally prospected. At the time of the field mapping, attempts were being made to recover antimony, and several areas were being explored for other metals by drilling and by sinking a shaft. Currently some silver ores are being shipped from the Taylor district, and further exploration activity is concentrated along the eastern edge of the district.

TAYLOR MINING DISTRICT

The Taylor mining district occupies several square miles low on the west flank of the Schell Creek Range and 3 miles northwest of Connors Pass. Silver was discovered in the district in July 1872 (Whitehill, 1873, p. 77), and according to Hill (1916, p. 200–202), who briefly described the geology and ore deposits of the area at a time when a little high-grade ore was still available, the mines in the area were most active from 1872 to 1878. Available production data are summarized in table 13.

Although Lower Devonian to Recent rocks are exposed in the district, only some of the Middle Devonian to Lower Mississippian rocks are mineralized. These rocks include much limestone and shale and some dolomite that are capped by a small amount of dacite lava and tuff and cut by numerous dikes of porphyritic rhyolite (fig. 13). Most of the mineralized rocks dip gently to moderately steeply eastward, except along the west side of the district, where both the direction of dip and the inclination are more varied. These rocks are cut by many normal faults, most of which trend north to northeast and dip 65°–70° SE. The Taylor fault, however, which is the largest, trends northwestward and dips steeply westward. One fault in the district is probably a low-angle fault. The stratigraphic displacement on the Taylor fault exceeds 1,000 feet where it brings unmineralized rocks against the mineralized rocks, but on the other faults it is in the order of hundreds of feet and diminishes northward.

The limestone of the Guilmette Formation and the Joana Limestone contains stringers, sheets, pods, and larger irregular bodies of siliceous rock, which form dark-brown to dark-gray ledges and ribs that project above the level of the adjacent limestone. Tabular bodies of siliceous rocks are alined along some of the north- to northeast-trending faults and along some north-striking stratigraphic units. Many of the larger silicified bodies grade into the limestone host rocks through zones of weakly silicified rock, in which quartz stringers are common; the separation of limestone fragments by a siliceous matrix indicates a replacement of limestone by silica. The silica has preferentially replaced thin-bedded shaly limestone of member c of the Guilmette Formation, especially its lower part, and it has also replaced, to a lesser extent, the thin-bedded limestone at the top of the Joana Limestone. Some of the contacts, however, along the smaller tabular bodies are strongly sheared and relatively abrupt, and some of the silicified bodies are sheared and brecciated. Near the largest open pit north of Taylor townsite, the shear planes commonly strike N. 25° W.–N. 40° E. and dip vertically to 50° W. Near the large fault that runs northeastward through the district, the attitudes of the shear planes approach that of the fault, this relation indicating that the shears feather out away from the fault. Carbonate and quartz veinlets cut the silicified rock and commonly fill the interstices in the brecciated rock.

The district contains many small strongly altered but unmineralized porphyritic rhyolite dikes. Where they cut silicified rock the dikes are subparallel to the nearby shear planes.

Most of the ore has been mined from opencuts and from a maze of shallow drifts extending along the main silicified horizons in the central part of the district. About 20 shafts and inclines are more than 50 feet deep, and some may be more than 200 feet deep. None of the deeper workings were entered and, although the pits and shallow passages are accessible, their floors are commonly covered with waste material. During part of the time in which the area was being mapped, a small stibnite prospect was being mined east of the townsite.

Ore minerals are scarce and inconspicuous in most of the workings. Some bluish-green mineral, possibly chrysocolla, occurs in small veinlets and fracture coatings in the central and northern part of the district. Azurite and malachite are less common, and some rock fragments are coated with a dark-gray sooty material, probably either a copper or a manganese mineral. Limonite and other secondary iron minerals, some having conspicuous boxwork structures, are common in the prospects in the southern part of the district. A barite vein roughly follows a locally faulted contact between the Guilmette Formation and Pilot Shale about 1,000 feet northwest of the unnamed 9,200-foot peak in the northeast corner of the district. Stibnite is abundant in pods of silicified Joana Limestone east and northeast of the central part of the district and is most abundant at the prospect in an area of much faulted Chainman Shale and silicified and mineralized Joana Limestone. The silicified limestone is fractured and brecciated, and the fragments are cemented with stibnite, calcite, silica, and a bluish-green copper(?) mineral. Nearby are some porphyritic rhyolite dikes which appear to be younger than the faults that cut the silicified limestone.

Spectrographic analyses of 40 samples from six of the larger workings were made to supplement the scanty mineralogical data and to show the distribution of silver in the silicified rock (table 14). Three of the workings extend along the east side of the central part of the district, and the others, which lie west of the central part of the district, are separated from these by a normal fault. The workings from which collections 17–22 were made are identified on figure 13. The samples were obtained either from rock in place or from waste in the bottom of the pit and contain fragments typical of the rocks exposed

TABLE 13.—*Production record of the Taylor mining district*

[Only those years listed for which production was recorded. Ore production: includes ore from reworked tailings]

Compiled by Couch and Carpenter (1943)

Year	Number of mines producing	Ore produced (short tons)	Metals recovered	Mines or mining companies known	Total value
1883		1, 875	Combined Ag, Au, Cu, and Pb		$66, 187
1884		5, 545	do	Monitor, Sunrise	150, 617
1885		8, 935	do	do	258, 914
1886		6, 746	do	do	193, 020
1887		6, 644	do	Monitor, Sunrise, Argus.	161, 975
1888		4, 428	do	do	111, 181
1889		999	do	do	43, 641
1890		907	do	do	21, 414
1891		1, 277	do	do	25, 596
1892		463	do	do	7, 994
1936		3, 127	do	Oxborrow and Mckenzie.	22, 121
1940		18, 865	do	Oxborrow and Mckenzie, and Ely Gold Mining Co.	118, 965
					$1, 181, 625

Compiled by U.S. Bureau of Mines (1937–61)

Year	Number of mines producing	Ore produced (short tons)	Au (Fine ounces)	Ag (Fine ounces)	Cu (Pounds)	Pb (Pounds)	Zn (Pounds)	Sb (Pounds)	Mines or mining companies known	Total value
1935	2	15, 442	288	104, 656						$85, 285
1936	3	29, 265	887	176, 938						168, 093
1937	3	1, 271	18	8, 002						6, 820
1938	1	243	4	1, 433						1, 066
1939	2	1, 320	28	8, 365						6, 658
1940	9	12, 932	385	107, 029					Ely Mining Co	89, 585
1941	6	14, 244	418	199, 900					Gore Mineral Farm, Monitor, Sunrise.	156, 781
1942	5	9, 084	208	90, 544					Mineral Farm	71, 667
1946	3	3, 284	37	12, 892	500	10, 600				12, 948
1947	3	2, 957	74	15, 532					Goodman mine (Nevada Sunshine Mining Co.).	16, 646
1948	3	102		897	1, 300	17, 100	7, 200			5, 113
1949	1	13	2	1, 134						1, 096
1950	2	411	6	6, 046						5, 682
1951	1	190	4	1, 411						1, 417
1952									Mineral Farm, Garder Mining Co.	
1953–54									Taylor mine, Imperial Nevada Co.	
										$628, 857

Estimated by Drewes

Year	Number of mines producing	Ore produced (short tons)	Au	Ag	Cu	Pb	Zn	Sb	Mines or mining companies known	Total value
1959									Merrimac claims	Several thousand dollars (for Sb).

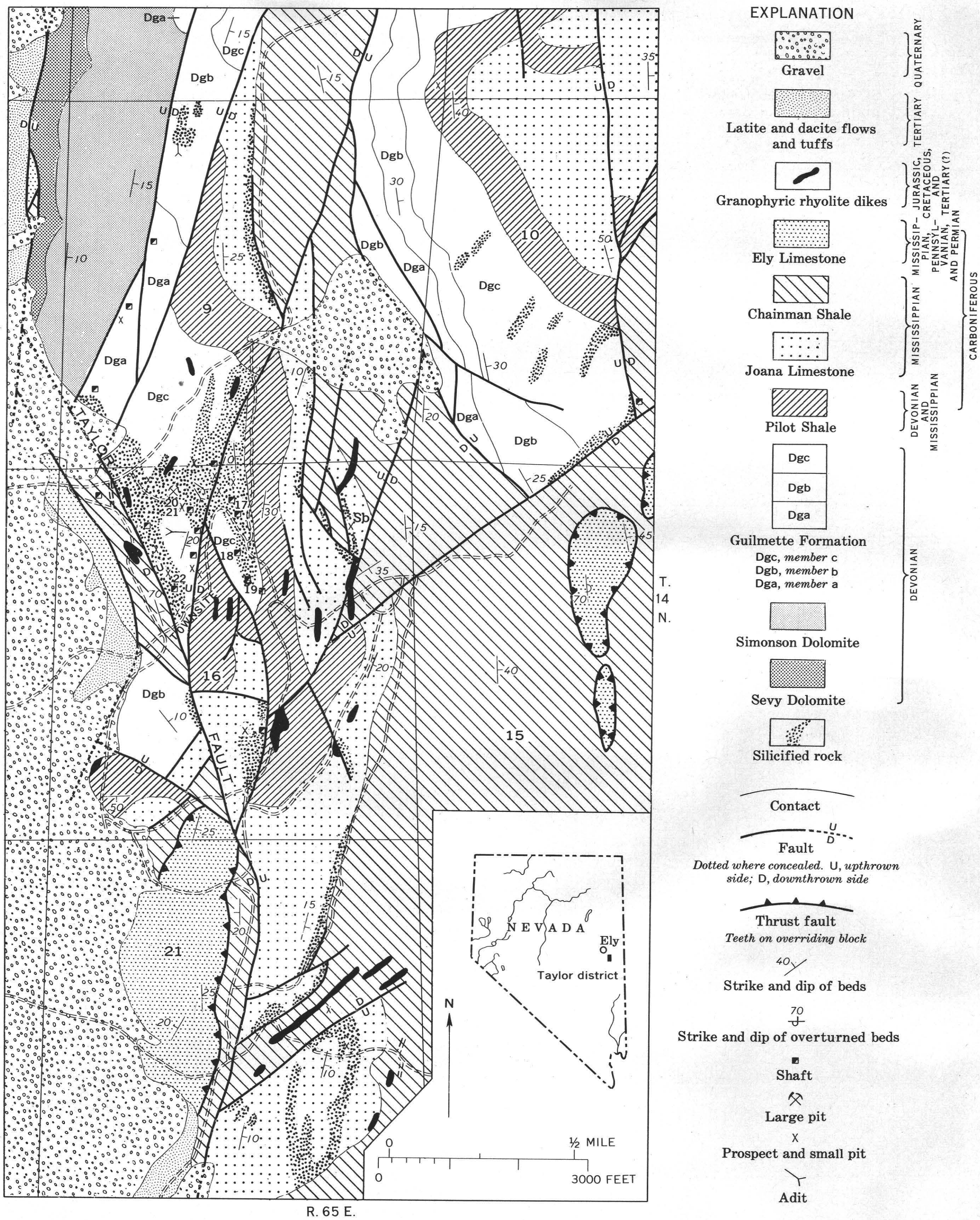

FIGURE 13.—Geologic map of the Taylor mining district, in the western part of the Connors Pass quadrangle, modified from Drewes (1962). Silver ore occurs in some silicified bodies. Numbers 17–22 at workings indicate sample localities.

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TABLE 14.—*Semiquantitative spectrographic analyses of rocks from the Taylor mining district*

[Analyses, in parts per million, by E. F. Cooley. Be <1, Ge <20, Ga <20, Sn <10, Sc <10, Bi <10, In <50, Ti <100, W <50, Nd <50, Ta <50]

Sample loc. (fig. 13)	Lab. No. 61–	Rock type	Visible mineralization	Fe	Mg	Si	Cu	Pb	Zn	Sb	Ag	Mn	As	Ti	Cd	La	B	Cr	Ba	Sr	Zr	Ni	Co	V	Mo	Y
17a	242S	Silica breccia	--------	15, 000	5, 000	500, 000	500	1, 500	700	5, 000	1, 000	300	<1, 000	1, 500	<50	<50	50	50	150	70	200	7	<10	70	20	10
b	243S	Silica	Cu	7, 000	1, 500	300, 000	5, 000	>10, 000	10, 000	7, 000	150	500	1, 000	1, 500	200	<50	20	30	100	200	500	10	<10	30	<5	15
c	244S	Calcite vein in silica	--------	1, 500	2, 000	70, 000	200	300	<200	200	200	>10, 000	<1, 000	200	<50	<50	<10	10	20	1, 000	10	5	<10	30	<5	15
d	245S	Silicified limestone	--------	1, 500	30, 000	200, 000	500	1, 000	1, 000	200	200	5, 000	<1, 000	200	<50	<50	10	10	50	200	150	10	<10	100	<5	10
e	246S	Silica breccia	--------	5, 000	2, 000	500, 000	500	700	1, 000	1, 000	2, 000	200	<1, 000	700	<50	<50	50	20	200	50	300	10	<10	50	<5	10
f	247S	Vuggy silica	--------	5, 000	1, 500	500, 000	500	1, 000	500	3, 000	3, 000	100	<1, 000	700	100	<50	50	20	100	50	200	7	<10	20	<5	10
g	248S	Silica breccia and quartz vein.	--------	5, 000	1, 500	500, 000	200	700	500	5, 000	700	150	<1, 000	500	<50	<50	30	15	100	10	150	5	<10	20	<5	10
18a	249S	Altered limestone	--------	5, 000	2, 000	300, 000	1, 500	3, 000	5, 000	5, 000	1, 000	200	1, 000	700	100	<50	20	20	100	500	200	5	<10	50	<5	10
b	250S	Silica	Cu	2, 000	5, 000	100, 000	10, 000	>10, 000	>10, 000	10, 000	1, 000	700	1, 500	500	1, 500	<50	20	15	100	<10	100	50	<10	15	<5	15
c	251S	Vuggy silica	--------	3, 000	1, 000	500, 000	1, 000	2, 000	10, 000	3, 000	150	50	<1, 000	500	150	<50	20	15	150	20	200	5	<10	30	<5	10
d	252S	Calcite vein and silica.	--------	5, 000	3, 000	100, 000	700	1, 500	5, 000	2, 000	150	1, 500	<1, 000	1, 000	150	<50	15	20	50	700	200	5	<10	50	<5	20
e	253S	Silica	Cu	20, 000	1, 500	300, 000	>10, 000	>10, 000	>10, 000	>10, 000	5, 000	1, 500	5, 000	2, 000	5, 000	<50	50	30	300	70	200	20	<10	50	20	20
19a	254S	Dolomite breccia	Cu	10, 000	1, 500	300, 000	5, 000	>10, 000	>10, 000	7, 000	2, 000	500	<1, 000	1, 500	300	<50	30	20	1, 500	200	300	10	<10	30	500	10
b	255S	Silicified limestone	Cu	10, 000	1, 500	500, 000	3, 000	>10, 000	2, 000	7, 000	2, 000	200	<1, 000	1, 500	100	<50	50	30	200	100	200	10	<10	30	10	20
c	256S	Silica breccia	--------	5, 000	1, 500	500, 000	2, 000	10, 000	>10, 000	3, 000	1, 500	700	<1, 000	1, 000	200	<50	30	30	1, 500	200	100	20	<10	30	1, 500	20
d	257S	Silica	Cu	10, 000	500	300, 000	>10, 000	>10, 000	>10, 000	>10, 000	5, 000	150	7, 000	500	3, 000	<50	30	20	150	20	150	10	<10	20	20	20
e	258S	Limestone replaced by quartz and calcite.	--------	1, 000	5, 000	15, 000	1, 500	3, 000	2, 000	500	150	500	<1, 000	300	50	<50	<10	<10	1, 000	300	10	<5	<10	30	20	20
f	259S	Silicified limestone	--------	5, 000	1, 500	500, 000	1, 000	3, 000	2, 000	1, 000	1, 000	50	<1, 000	1, 000	50	<50	50	20	200	50	500	5	<10	20	<5	20
20a	273S	Silica breccia	Sb	1, 000	500	500, 000	200	20	500	>10, 000	150	50	1, 000	200	<50	<50	20	20	500	50	30	20	<10	15	<5	<10
b	274S	Calcite vein and silicia.	Cu	5, 000	1, 500	100, 000	10, 000	3, 000	10, 000	5, 000	1, 500	2, 000	1, 000	100	150	<50	<10	10	50	200	30	<5	<10	10	7	10
c	275S	Silica breccia	Cu	1, 500	500	300, 000	>10, 000	1, 000	10, 000	500	150	500	<1, 000	200	<50	<50	20	10	150	50	200	10	<10	10	<5	10
d	276S	Silicified limestone	Sb	1, 500	1, 000	300, 000	1, 000	2, 000	500	7, 000	150	10	<1, 000	1, 500	50	<50	20	15	700	150	500	<5	<10	20	100	10
e	277S	Silica	Sb	1, 500	500	500, 000	700	200	500	>10, 000	700	10	1, 000	150	<50	<50	10	20	1, 500	100	10	50	<10	10	5	<10
f	278S	Silicified limestone	Cu	20, 000	1, 000	200, 000	>10, 000	>10, 000	>10, 000	>10, 000	1, 500	3, 000	1, 000	200	200	50	20	10	700	500	20	10	<10	15	10	20
g	279S	Silica and stibnite	Sb	1, 000	500	150, 000	1, 000	150	1, 000	>10, 000	3, 000	50	2, 000	70	<50	<50	<10	10	>10, 000	100	<10	20	<10	10	<5	10
h	280S	Silica	Cu	5, 000	500	300, 000	>10, 000	>10, 000	>10, 000	>10, 000	150	3, 000	<1, 000	200	<50	100	50	10	700	20	150	10	10	15	10	15
i	281S	Vein calcite	--------	300	1, 500	5, 000	500	200	1, 000	5, 000	15	700	<1, 000	<10	<50	<50	<10	10	200	500	<10	<10	<10	10	<5	15
21a	260S	Silica breccia	--------	2, 000	1, 500	500, 000	500	500	2, 000	200	150	200	<1, 000	1, 000	<50	<50	50	10	150	500	300	5	<10	20	7	10
b	261S	Vuggy silica	--------	1, 500	1, 000	500, 000	700	1, 000	2, 000	1, 000	1, 500	20	<1, 000	700	50	<50	30	10	100	10	300	<5	<10	10	<5	10
c	262S	Silicified limestone	--------	1, 000	1, 000	500, 000	300	300	3, 000	300	1, 000	700	<1, 000	150	50	<50	20	15	150	100	70	<5	<10	20	<5	<10
d	263S	Silica	--------	1, 500	700	500, 000	300	300	3, 000	200	150	100	<1, 000	300	<50	<50	20	10	100	20	200	5	<10	10	<5	10
e	264S	Vuggy silica	--------	500	500	500, 000	150	70	<200	<200	70	100	<1, 000	100	<50	<50	20	10	70	50	100	<5	<10	<10	<5	10
f	265S	Silcified limestone breccia.	--------	1, 500	1, 000	500, 000	150	200	300	<200	150	300	<1, 000	200	<50	<50	30	10	100	100	200	5	<10	20	<5	10
22a	266S	Silica and calcite veins.	--------	700	1, 000	200, 000	150	100	200	200	200	2, 000	<1, 000	150	<50	<50	20	10	50	200	50	<5	<10	10	<5	10
b	267S	-----do	--------	2, 000	1, 000	500, 000	500	1, 000	10, 000	3, 000	500	300	<1, 000	700	100	<50	30	10	70	150	300	10	<10	15	<5	10
c	268S	-----do	--------	1, 000	2, 000	200, 000	200	2, 000	1, 000	2, 000	2, 000	2, 000	<1, 000	150	100	<50	10	10	20	700	20	<5	<10	20	5	10
d	269S	Vuggy silica	--------	2, 000	700	500, 000	300	700	1, 000	1, 000	2, 000	150	<1, 000	300	<50	<50	30	10	100	50	100	7	<10	10	<5	10
e	270S	-----do	--------	1, 500	1, 000	300, 000	150	500	1, 500	500	700	700	<1, 000	200	50	<50	20	10	50	100	70	5	<10	10	<5	10
f	271S	Dolomitic limestone.	--------	500	2, 000	20, 000	70	70	500	<200	70	300	<1, 000	50	<50	<50	<10	<10	30	1, 000	10	<5	<10	20	<5	10
g	272S	Silica	--------	1, 000	1, 000	300, 000	70	100	1, 500	200	100	200	<1, 000	200	<50	<50	20	10	50	100	30	7	<10	15	<5	<10

in the walls nearby. Both field observations and analytical results showed that the mineralization is strongest in a zone extending from the large pit, locality 20, on the northwest side of the central part of the district to locality 19, slightly southeast of the center of the district.

Bluish-green copper(?) minerals can be recognized in any sample that contains as much as 300ppm of copper. Stibnite or a secondary antimony mineral is recognizable in any sample containing as much as 10,000 ppm of antimony, unless the sample also contains large quantities of secondary minerals of other base metals, which mask secondary minerals derived from antimony. No lead, zinc, or silver minerals could be identified, even with the aid of the analyses. Rocks having a large content of silver or base metals most commonly consist of brecciated silica, less commonly of nonbrecciated silica, and least commonly of silicified limestone; much of the silicified rock, however, contains only small amounts of silver or base metals.

Lead and zinc minerals are most closely associated, copper is somewhat more widely distributed than lead and zinc, and antimony is not always associated with the other base metals. Silver tends slightly to be associated with antimony rather than with other base metals.

The rocks were probably mineralized in late Mesozoic or early Tertiary time, apparently in the early part of the period during which the region was being subjected to tensional stresses. Siliceous solutions probably followed the main faults and selectively replaced the impure limestones beneath the more refractory shales. The silicified rock was fractured by further adjustments along faults, and ore-bearing solutions also apparently followed the main faults and deposited base metals and silver in the zones of severest brecciation near the center of the district. Solutions of antimony and perhaps of silver retained mobility longer than those of other metals and thus ascended a little higher. Shortly after the mineralization, porphyritic rhyolite was intruded along or near the main fractures. The siliceous solutions and the magma may both have been derived from a single stock that lies not far beneath the surface. Further movement along the faults trending north and northeast was succeeded by movement along those trending northwest and also, presumably, by the extrusion of the Tertiary latite and dacite. Still later the mineral deposits were altered and secondary minerals were concentrated near the surface.

Guides to prospecting in the Taylor district (Drewes, 1962) place the chief emphasis on downdip exploration of the two mineralized horizons along which additional silica pods enriched in silver may be anticipated. Extensions of these horizons across some of the faults along the eastern edge of the district may also be found by deep drilling.

CLEVE CREEK DISTRICT

There are two groups of prospects and mines on quartz veins in the Cleve Creek district, but they were developed only a short time. One group of shafts and adits is scattered along the lower part of Cleve Creek 1½ miles west of the Cleve Creek campground and extends to the crest of the hill south of the creek. These prospects are all in the Pioche Shale or the upper part of the Prospect Mountain Quartzite. Vein quartz and dolomite are abundant in the rubble around the shafts near the creek; they contain traces of a black mineral, perhaps hematite or hübnerite, and of a bluish-green copper(?) mineral. South of Cleve Creek a shallow incline follows a quartz vein containing pyrite, bornite, and malachite in a limy bed in the Pioche Shale. In this area the alluvium and colluvium have a slightly anomalous base-metal content that extends east of the area of prospects (fig. 14).

The Kolcheck mine and surrounding prospects lie high above the west side of the Kolcheck Basin, along the upper part of Cleve Creek. The workings are in the middle of the complexly faulted Pole Canyon Limestone. According to the U.S. Bureau of Mines (1953, v. 3, p. 672), 32 tons of ore, containing 3.15 percent WO_3, was shipped from the mine in 1953. Scheelite is probably the chief or only tungsten mineral.

The prospects containing quartz, pyrite, malachite, and possibly hübnerite in quartzite and shale are similar to the deposits at and north of the Hub Mine Basin in the southern Snake Range (Drewes, 1954, p. 114), and the setting of the Kolcheck mine resembles that of the Minerva district, 10 miles south of the Hub Mine Basin. In the Hub Mine Basin and the Minerva district the mineralization is closely associated with a granodiorite stock or with porphyritic rhyolite dikes, this relation indicating that hydrothermal solutions may have emanated from a magma chamber to form a manganese tungstate in the quartzite and a calcium tungstate in the limestone. The mineralization in the Cleve Creek area may have originated in a similar manner to that in the southern Snake Range but was less intensive.

MINERALIZATION NEAR MAJORS PLACE

Several shafts, adits, and smaller prospects lie high on the east flank of the ridge about 1½ miles northwest of Majors Place, and a few prospects lie farther north on the ridge, about a mile south of Cooper Canyon. Some development work was done near Majors Place a few years before the area was mapped, but no production was recorded. Most of these prospects were dug in sheared shale of the lower part of the Lincoln Peak Formation, but several were extended a short distance into the top of the Pole Canyon Limestone, and those to the north were dug in the middle of the Pole Canyon. Rubble of vein quartz is widely scattered throughout this area, and the quartz veins at the prospects contain traces of a bluish-green copper(?) mineral.

Fluorite appears in two small prospects in Rattlesnake Knoll, just south of U.S. Highway 6–50 and about 4 miles east-northeast of Majors Place. Here the host rock is fragmental dacite vitrophyre, perhaps a flow agglomerate; fluorite fills some of the voids between the angular fragments, and some of the fragments are cut by fluorite veins. Most of the fluorite is light gray, but some of it is greenish or pinkish gray. This mineralization is the youngest in the area for it affects the dacite volcanics of Eocene and Oligocene(?) age, and it may be related to that volcanism.

OTHER MINERALIZATION

Several adits and small prospects lie in Tamberlaine Canyon, about 3 miles south of the northwest corner of the quadrangle. A small amount of silver(?) ore was probably shipped from the largest adits, but production records are not available. The workings are in pods of silicified limestone in members b and c of the Guilmette Formation, along a zone that extends westward from the major normal fault trending northwestward across the canyon and continues beyond the west boundary of the quadrangle. Quartz replaces limestone along northeast-trending fractures within the main east-trending zone. Some of the quartz is brecciated and sheared and contains later silica and coatings of brownish- and yellowish-gray material, and the mineralization is associated with the brecciated silicified limestone. No ore minerals were recognized, but silver was probably mined here as well as in the Taylor district.

Other small silicified bodies are scattered throughout the quadrangle. Although I do not know whether they are similarly mineralized or not, they closely resemble the larger silicified and mineralized bodies and are probably of the same age. One group of silicified bodies is along a gently inclined fault in the Ely Limestone at the southeast end of the klippe west of Grasshopper Canyon. Another silicified body, of pipelike form, is at the junction of several normal faults along Cooper Wash, SW¼ sec. 33, T. 15 N., R. 65 E., and still another silica pod appears on the north wall of Cooper Canyon three-quarters of a mile east of the center of the basin, SW¼ sec. 36, T. 15 N., R. 65 E. Several other silicified pods extend along the fault between the Arcturus Formation and the dacite vitrophyre volcanic rocks south of Cooper Canyon, in the projected position of SW¼ sec. 8, T. 14 N., R. 66 E. The bodies along this fault are mapped as silicified Arcturus Formation, but they could be altered Ely Limestone or Guilmette Formation that has been faulted, for a small block of Chainman Shale also extends along the fault. Silicified rock along a fault in Chainman Shale at the head of Steptoe Creek, NW¼ sec. 20, T. 16 N., R. 65 E., may likewise be infaulted silicified Ely Limestone, Joana Limestone, or limestone of the Guilmette Formation, all of which are exposed nearby.

A small exploratory shaft was sunk in the Joana Limestone on a bench west of Steptoe Creek slightly more than a mile south of the north edge of the quadrangle. No mineralization was seen in the debris on the dump, but slightly anomalous lead contents were found in some samples of alluvium collected a short distance downstream. Inasmuch as the shaft is less than 3 miles south of the Success mine, from which small amounts of lead, silver, and gold were reported to have been produced between 1907 and 1910 (Hill, 1916, p. 198), it was perhaps sunk in search of a similar deposit.

GENERAL GEOCHEMISTRY

A general survey was made of trace-element content in alluvium and black shale (tables 15 and 16). Alluvium was sampled close to the range fronts and near the mouths of the large tributaries of the major watercourses. Samples of silt and sand were collected from at least 2 inches below the surface of the youngest deposits in the bottoms of dry washes or in sand bars next to running water. Most of the samples represent material derived from drainage areas no more than a few square miles in extent, and all but a few drainage areas of any considerable size were sampled. Alluvium in unmineralized areas contained, on the average, about 20 ppm copper, 25 ppm lead, and 25 ppm zinc, which may be considered background values of those metals, but the tungsten content is below the threshold value of detection. Zinc background values increased somewhat erratically in areas underlain by rocks of late Paleozoic age, particularly in areas underlain by black shale.

TABLE 15.—*Semiquantitative analyses of lead, copper, and zinc content of alluvium*

[Analyses, in parts per million, from field-method wet tests in the laboratory by H. H. Mehnert, W. W. Janes, and K. W. Leong. Tungsten content <20 parts per million except for sample 69, which contains 20 ppm]

Sample loc. (fig. 14)	Lab. No.	Element			Sample loc. (fig. 14)	Lab. No.	Element			Sample loc. (fig. 14)	Lab. No.	Element		
		Pb	Cu	Zn			Pb	Cu	Zn			Pb	Cu	Zn
49	59-1645S	50	50	50	81	2938S	25	20	50	113	2919S	25	30	50
50	1646S	<25	40	50	82	2935S	25	10	50	114	2918S	25	10	25
51	1647S	<25	40	50	83	2934S	25	30	50	115	2917S	25	10	50
52	1648S	25	40	50	84	2933S	25	10	25	116	2916S	25	10	50
53	1652S	<25	40	50	85	2932S	25	10	25	117	2915S	25	20	50
54	1657S	<25	20	25	86	2931S	25	20	50	118	2902S	25	20	25
55	1658S	<25	20	25	87	2940S	25	10	25	119	2943S	25	10	25
56	1672S	<25	20	25	88	2939S	25	10	25	120	59-1677S	<25	20	25
57	1673S	25	40	25	89	2944S	25	20	25	121	1671S	<25	20	25
58	1674S	<25	30	50	90	2942S	25	10	25	122	1670S	<25	20	50
59	1675S	<25	20	25	91	2930S	25	10	50	123	1679S	<25	10	25
60	1676S	<25	20	25	92	2929S	25	10	25	124	1678S	<25	20	50
61	1680S	<25	20	25	93	2928S	25	10	25	125	61-2911S	25	10	50
62	1681S	<25	20	25	94	2941S	25	10	50	126	2912S	25	10	75
63	60-2906S	25	10	25	95	61-2916S	25	40	75	127	2914S	25	30	100
64	2907S	25	20	25	96	60-2946S	25	20	50	128	2913S	25	10	75
65	2905S	25	20	25	97	2945S	25	20	50	129	59-1669S	<25	20	<25
66	59-1644S	75	75	75	98	2947S	25	20	50	130	1668S	<25	20	25
67	60-2913S	25	20	50	99	2914S	25	10	25	131	1666S	<25	30	25
68	2912S	25	30	25	100	2926S	25	10	25	132	1667S	<25	20	25
69	2911S	25	30	75	101	2927S	25	10	50	133	1665S	<25	30	25
70	2910S	25	20	25	102	2925S	25	10	25	134	1664S	<25	20	25
71	2909S	25	20	<25	103	2924S	25	10	50	135	1663S	<25	20	25
72	2908S	50	20	50	104	61-2907S	50	40	75	136	1662S	<25	20	25
73	59-1649S	25	75	50	105	60-2923S	25	10	25	137	1661S	<25	20	25
74	1651S	<25	40	50	106	61-2908S	50	40	75	138	1656S	<25	20	25
75	1650S	<25	40	25	107	2909S	50	30	75	139	1660S	<25	20	25
76	1655S	<25	30	50	108	60-2922S	25	10	50	140 [1]	1659S	<25	20	25
77	1653S	<25	40	25	109	61-2910S	50	30	75	141	60-2903S	25	10	<25
78	1654S	25	40	25	110	2915S	25	20	75	142	2904S	25	10	<25
79	60-2937S	50	10	50	111	60-2921S	25	30	50					
80	2936S	25	20	50	112	2920S	25	20	50					

[1] Sample from terrace 18 inches above stream and sample from locality 61.

TABLE 16.—*Semiquantitative spectrographic analyses of alluvium*

Analyses, in percent, by E. F. Cooley. Elements looked for, but not found, in ppm: Ag <1, As <1000, Bi <10, Cd <50, Ga <20, Ge <20, In <50, La <50, Nb <50, Sb <200 Sc <10, Zn <200]

Sample loc. (fig. 14) / Lab. No.	49 59-1645	50 59-1646	51 59-1647	52 59-1648	53 59-1652	54 59-1657	55 59-1658	56 59-1672	57 59-1673	58 59-1674	59 59-1675	60 59-1676	61 59-1680	62 59-1681
Si	20	10	20	30	15	30	20	20	10	10	7	30	10	20
Fe	2	2	3	2	2	1.5	1.5	1.5	3	2	1	3	2	3
Mg	2	2	2	1.5	2	.3	1	1.5	2	2	2	1.5	1.5	1
Ti	.3	.15	.3	.3	.1	.15	.15	.1	.15	.15	.1	.3	.2	.15
Mn	.07	.1	.07	.07	.15	.02	.05	.07	.07	.07	.05	.07	.07	.07
B	.005	.005	.01	.01	.003	.002	.005	.002	.003	.003	.001	.007	.003	.003
Ba	.05	.03	.07	.05	.05	.05	.05	.03	.03	.03	.03	.05	.05	.07
Be	<.0001	.0001	.0001	<.0001	<.0001	<.0001	<.0001	<.0001	.0001	.0001	<.0001	.0001	.0001	<.0001
Co	<.001	<.001	.001	.001	.001	.001	<.001	<.001	.0015	.001	<.001	.001	.001	<.001
Cr	.005	.003	.007	.005	.005	.005	.003	.007	.007	.002	.003	.005	.007	.007
Cu	.003	.003	.003	.003	.002	.0007	.002	.0015	.005	.005	.002	.005	.003	.002
Ni	.0015	.0015	.002	.0015	.001	.0015	.001	.0015	.003	.002	.001	.002	.002	.002
Pb	.002	.003	.002	.002	.0015	<.001	.002	.001	.005	.005	.002	.003	.002	.002
Sr	.03	.05	.02	.01	.03	.005	.02	.03	.07	.02	.03	.02	.05	.05
V	.005	.005	.01	.01	.005	.005	.005	.005	.01	.003	.003	.01	.007	.005
Y	.003	.003	.005	.002	.002	.003	.003	.003	.002	.002	.002	.005	.002	.003
Zr	.03	.007	.02	.02	.007	.015	.01	.015	.005	.01	.005	.02	.01	.01

The areas that show minor anomalies in the content of copper, lead, and zinc correspond closely with known mineralized areas, as illustrated on figure 14. Very weak anomalies, having roughly two to three times the background value of a single metal, are near small mineralized areas, and somewhat more pronounced anomalies, involving several metals or greater values for one metal, are near two of the larger mineralized areas. Lead seems to be a more sensitive indicator of moderately intense mineralization than either copper or zinc. The extent and intensity of the anomalies in mineralized areas is not proportional to the extent and intensity of the known mineralization in those areas. The anomaly along lower Cleve Creek, for instance, is of the same order of magnitude as that near the Taylor district, although there appears to have been much less mineralization along Cleve Creek than in the Taylor district. Perhaps there are minalized rocks along Cleve Creek that have not yet been discovered, or possibly the quartzite terrane along Cleve Creek facilitates the retention of the metals in forms readily found by the tests employed, whereas in the limestone-rich terrane of the Taylor district the metals are washed out or are not found by those

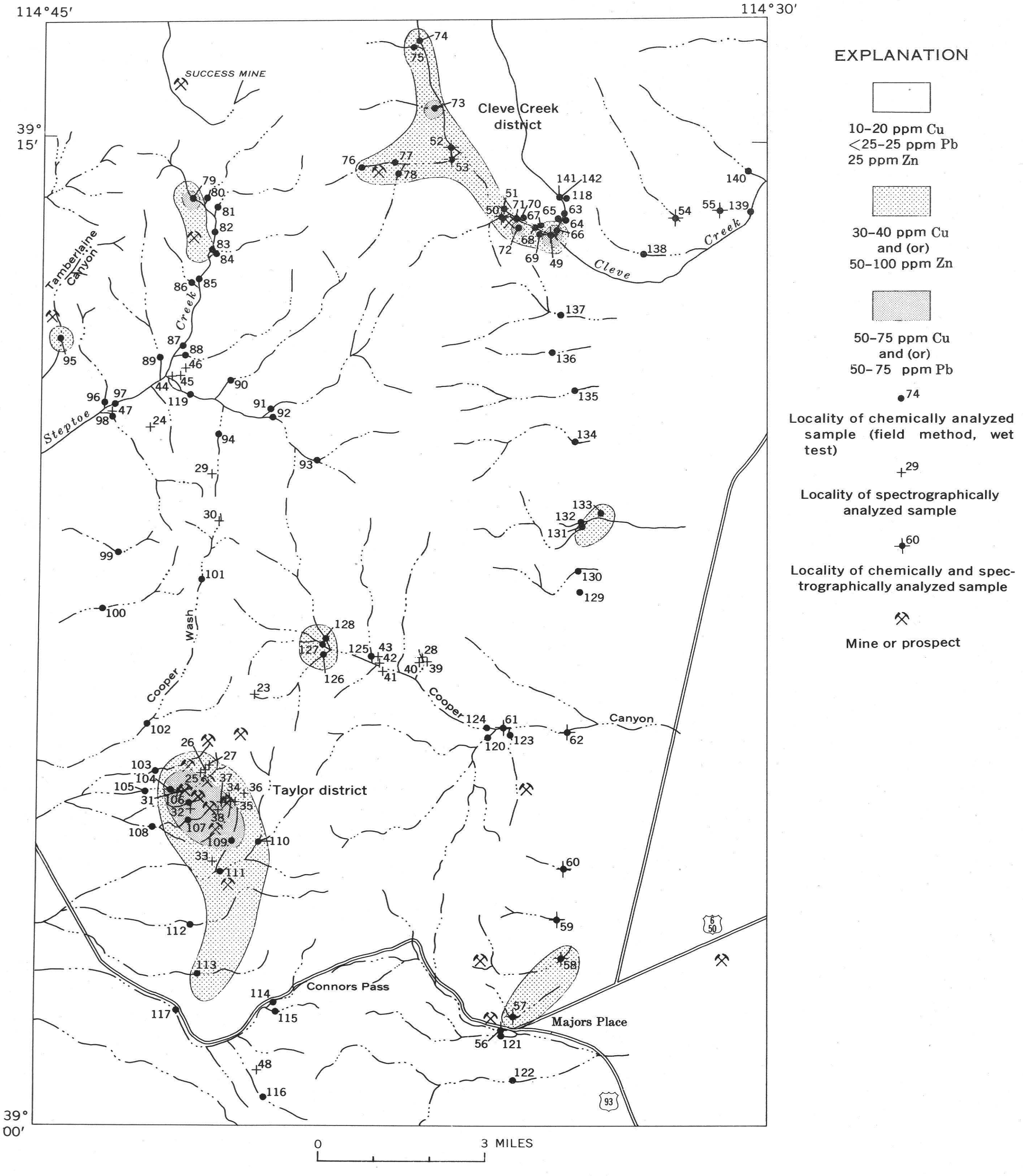

FIGURE 14.—Distribution of lead, copper, and zinc in alluvium in the Connors Pass quadrangle.

tests. This idea seems to be supported by the abrupt decrease of metal content in the washes that drain the western edge of the Taylor district, as opposed to gradual decrease of metal content in the washes to the east and south. When the same analytic methods are used and the same type of mineralization is being checked, the sampling interval in a limestone terrane ought not to be more than 1 mile, but in a quartzite terrane may be slightly larger.

Although strong prevailing winds frequently carry fumes from the copper smelter at McGill southward along the west front of the range for about 30 miles, the samples taken in the Taylor district do not record any base-metal anomalies attributable to contamination by these fumes. Within the Connors Pass quadrangle, a minimum of 15 miles from the smelter, the depth at which samples were collected apparently was adequate to safeguard against contamination by airborne particles.

Semiquantitative spectrographic analyses were made of many alluvium samples taken in areas underlain by Pioche Shale to check specifically for beryllium, which occurs in the Pioche southeast across Spring Valley (Stager, 1960; Whitebread and Lee, 1961). The contents of copper and lead measured by these analyses generally agreed with those measured by the wet tests (table 15).

Semiquantitative spectrographic analyses of samples from the Pilot Shale and the Chainman Shale showed anomalously high contents of several elements (numbers in italic type on table 5). One sample contained 0.2 percent zinc, 0.1 percent each of chromium and vanadium, and smaller but significant amounts of copper, silver, and other elements. The high values of copper, silver, and zinc are of special interest because these metals have been economically important in the region. However, a further cursory study showed no obvious relations of high values of some metals in these shales to either areal or stratigraphic variations, to distance from the large thrust fault or from an intrusive body, or to the mineralization in the Taylor district. No explanation has yet been found for the local concentration of some metals, but it may be revealed by more detailed study of the sedimentation and environment during, and shortly after, the deposition of the black shales.

GEOLOGIC HISTORY

The following section gives the best interpretation I am now able to make of the geologic history. The interpretation is necessarily incomplete, because a great part of what happened in the area since the end of Precambrian time has left no geologic record.

Near the end of late Precambrian time, the sea covered the region and currents brought much sand and some mud into the area. Marine deposition continued into Cambrian time without interruption and without change in the type of sediments.

During Paleozoic time, the sea floor seems never to have been deeply submerged, but it subsided almost continuously, though not at a uniform rate, to receive sediments aggregating about 5 miles in thickness. The tectonic conditions during this era changed markedly twice, these changes causing both local and regional changes in the character of the sediments and in the rate of sedimentation. On the basis of these changes, the Paleozoic Era can be divided into three parts, each of which lasted about 120 million years.

During early Paleozoic time, about 12,000 feet of sediment was deposited at a fairly uniform rate of about 100 feet per million years. In Early Cambrian and part of Middle Cambrian time, sand and mud similar to those deposited in late Precambrian time continued to be brought into the area from the east, but later in the Cambrian Period the shoreline moved progressively eastward and carbonate sediments were deposited. Early in the Late Cambrian, during Lincoln Peak time, deposition of carbonates alternated with deposition of mud that came from the west; then the sea regressed and the source of the clastics shifted back to the east. Toward the end of Cambrian time and in much of Early Ordovician time, less clastic material was brought into the area; thereafter it increased sporadically and carbonate sand, flakes and chips of limy mudstone, and bioclastic material were reworked. The sea was evidently shallow at this time, and it seems to have become increasingly shallow in the interval preceding the major change in the seaway system that followed.

The sea floor subsided much more slowly in middle Paleozoic time, and less than 4,000 feet of sediments accumulated at an average rate of 40–60 feet per million years. This phase began during Middle Ordovician time with the deposition of the well-sorted and well-rounded clean quartz sand that formed the Eureka Quartzite. The sand was deposited from a magnesium-charged sea. During much of middle Paleozoic time, the calcium carbonate that was precipitated was diagenetically altered to dolomite. In Late Ordovician time, much organic material was included with the sediments that formed the Fish Haven Dolomite. Perhaps in Early Silurian time, and certainly in Late Silurian, the area was nearly at sea level. Some of it may occasionally have risen briefly above sea level, but probably did not undergo much erosion. In Early Devonian time, immediately after the second emergence, the sea appears to have contained so much magnesium that primary dolomite was deposited to form the Sevy Dolomite. Late in Sevy time more clean quartz sand came into the area. Throughout Simonson time and the early part of Guilmette time, magnesium carbonate from sea water may have diagenetically changed the limy deposits to dolomite; however the supply of magnesium diminished gradually toward the end of Middle Devonian time, so the sediments of the Guilmette Formation are not completely dolomitized and show many replacement textures. Clean quartz sand was deposited in the area during middle Guilmette time, perhaps from a relatively distant source, for the last time; subsequently the clastic material was derived more locally as a result of the changing tectonic conditions that distingiushed late from middle Paleozoic time.

The sea floor again subsided more rapidly in late Paleozoic time and the area received abundant sediments at about half the regional average rate of about 200 feet per million years, this rate indicating that the Connors Pass area lay in a relatively stable part of a sea that was at least partly surrounded by tectonically active areas. In early Late Devonian time and only shortly after the deposition of the last clean sand, dirty poorly sorted and poorly rounded sand was intercalated with the calcareous deposits to record the earliest movements of the tectonically active areas (an area 100–150 miles to the northeast). Quieter conditions returned briefly and, at the end of Guilmette time, reef structures formed. Near the end of Devonian time the area seems to have been briefly uplifted, and the upper few hundred feet of the Guilmette Formation was largely removed by erosion before the Pilot Shale was deposited. The sea returned at the end of Devonian time and remained into Early Mississippian time, when the area was covered with the dirty silt and mud of the Pilot Shale, these deposits reflecting further tectonic activity nearby (to the northwest). During Joana time, calcium carbonate was again deposited and minor reef structures were formed in a shallow sea. During Late Mississippian time, however, clastics that formed the Chainman Shale were deposited in the area. Some patches of the black mud accumulated anomalously large amounts of chromium, vanadium, copper, silver, and zinc. Toward the end of Chainman time some sand and conglomerate were deposited in channels on mud that formed the Chainman; the area may have been a tidal flat or the floor of a lagoon. During Early Pennsylvanian time, a deeper sea returned to the area, and limy mud and small reefs of the Ely Limestone were deposited. Late in Pennsylvanian time, however, the area was once more uplifted, sufficiently at least for sediments from other areas to bypass it, and possibly sufficiently for sediments to be removed from it, and all were deposited in the rapidly subsiding Oquirrh Basin 60 miles to the northeast. Shortly after the beginning of the Permian, abundant dirty sand and silt were again brought in from the northwest to form the Rib Hill Sandstone; the influx of these sediments diminished somewhat in Arcturus time, when small reef structures formed. Late in this period, probably in Leonard time, some gypsum was deposited here.

Mesozoic sedimentary rocks are largely absent in the Connors Pass area because of orogeny that probably began in Late Jurassic time. In nearby areas about 5,000 feeet of sediments we deposited during Triassic and Early Jurassic time, largely in a marine environment. Near the end of the Triassic Period, the sea left the region for the last time, and at least the youngest of the deposits, correlative with the Navajo Sandstone of the Colorado Plateau region to the east, was deposited in a terrestrial environment.

During Late Jurassic time, the base of the Cambrian rock lay at depths as great as 7 miles, and because of the great weight of the overlying column, the Precambrian rock and some of the Cambrian rock were mildly metamorphosed to very low grade argillite, phyllite, slate, and marble. During Late Jurassic or Early Cretaceous time, the rock in the area was warped into widely scattered broad domes, and the Paleozoic rock was thrust northeastward for an unknown distance along multiple fault planes that lay subparallel to bedding. Subsidiary tear faults and normal faults were formed within the plates, and rock is generally missing rather than repeated along the faults that are almost parallel to the bedding. In some places the rocks along the thrust faults were dynamically metamorphosed. The greatest movement was concentrated along the Schell Creek Range thrust fault lying near the middle of the group of multiple low-angle faults. Some of the faults above the Schell Creek Range thrust fault may be glide faults, and they may even be glide faults along older thrust fault surfaces. By Late Cretaceous or early Paleocene time, the area had already been strongly unlifted and deeply eroded, and at least one basin had been formed in which clastic rocks accumulated. Some volcanism is also recorded by the volcanic ash intercalated with the sediments.

Epeirogenic and possibly also orogenic movements continued during Tertiary time, when the rocks were broken by normal faults and were again deeply eroded. It was probably during the Tertiary that the granophyric dikes were intruded and that siliceous fluids rose along the existing faults. In the Taylor district these solutions selectively replaced parts of the Guilmette Formation and the Joana Limestone with silica and ores of silver, antimony, and other base metals.

During Eocene time, the area lay along the western flank of a mountainous region and became partly covered with piedmont gravel. Volcanism became widespread during the Eocene and Oligocene(?) Epochs and was closely associated with later normal faulting. Basaltic andesite was the first intrusion, followed successively by the extrusion of a latite lava, eruption of abundant latite tuff (partly welded) and several extrusions of dacite lava interlayered with some dacite tuff.

Further movement along normal faults during middle or late Tertiary time caused north-trending grabens and block-faulted ranges that resemble the present ones in the region but that do not coincide with the present Schell Creek Range. Two large glide plates of Tertiary rock moved westward from a raised block into a valley cut in the Chainman Shale; the larger moved about 3,500 feet. While they were gliding, the plates were gently warped into open eastward-plunging synclines.

During late Tertiary and Quaternary time, volcanism ceased, but normal faulting and accumulation of piedmont gravel and basin deposits continued. Several faults cut the glide plates of middle or late Tertiary age and locally lie along older normal-fault surfaces. Probably in Pliocene time, gravel was deposited at the foot of some of the larger fault-block mountains and was itself faulted. The Schell Creek Range was probably formed in about its present position early in Quaternary time, and pediment and fan gravels were deposited along both of its flanks. Owing to the climatic changes during Pleistocene time and to renewed movement along normal faults extending along some of the mountain fronts, the older of these gravels was incised and then the younger and successively finer grained gravel was deposited. During late Pleistocene time, at least, pluvial Spring Valley Lake overlapped the eastern edge of the area. During the desiccation of the lake, bars were formed at progressively lower elevations, small dunes accumulated behind them, and they became partly covered by younger gravel.

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