Regional Geology of the Steamboat Springs Area Washoe County, Nevada

GEOLOGICAL SURVEY PROFESSIONAL PAPER 458-A

Thompson and White—REGIONAL GEOLOGY OF THE STEAMBOAT SPRINGS AREA, WASHOE COUNTY, NEVADA—Geological Survey Professional Paper 458-A

Regional Geology of the Steamboat Springs Area Washoe County, Nevada

By G. A. THOMPSON *and* D. E. WHITE

GEOLOGY AND GEOCHEMISTRY OF THE STEAMBOAT SPRINGS AREA, NEVADA

GEOLOGICAL SURVEY PROFESSIONAL PAPER 458-A

Volcanic geology, structure, and mineral deposits of the Mount Rose quadrangle and additional data from the Virginia City and nearby quadrangles

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1964

UNITED STATES DEPARTMENT OF THE INTERIOR

STEWART L. UDALL, *Secretary*

GEOLOGICAL SURVEY

Thomas B. Nolan, *Director*

For sale by the Superintendent of Documents, U.S. Government Printing Office
Washington, D.C. 20402

CONTENTS

ILLUSTRATIONS

[Plates are in pocket]

TABLES

REGIONAL GEOLOGY OF THE STEAMBOAT SPRINGS AREA, WASHOE COUNTY, NEVADA

By G. A. Thompson and D. E. White

ABSTRACT

The Mount Rose quadrangle, in southern Washoe County along the western border of Nevada, lies astride part of the bounding structures between the Sierra Nevada and the Basin Ranges. Together with the Virginia City quandrangle, which lies immediately to the east and which was studied at the same time, the area includes nearly 500 square miles and contains the formerly great silver-gold mines of the Comstock Lode and the well-known thermal area at Steamboat Springs. This report is focused on the Mount Rose quadrangle but also draws freely on data from the Virginia City quadrangle and to some extent from surrounding areas.

The region is one of climatic contrasts between the forested mountains of the Carson Range and the semiarid, generally treeless region to the east. The relief is 6,400 feet, and within this range of altitude is included both an upper timberline determined by winter cold and a lower timberline fixed by summer drought. The Carson Range was glaciated, and even at present snowfields in some places last through a summer into the following winter, but no certain signs of past glaciation have ever been found in the Virginia Range.

Geologic studies in the region began soon after the discovery of the Comstock Lode in the 1850's. Most of the previous geologic work consisted of local studies of the mineral-bearing districts or of regional reconnaissance, and there was little or no mapping outside the mining districts. Indeed, adequate base maps for quadrangle mapping were unavailable before the present work. The successive studies of the Comstock Lode and Steamboat Springs reflect to a remarkable degree the historical progress of the geologic sciences. Each new study brought new understanding based upon improved theory and new tools, but in the complexity of the volcanism, structural deformation, and hydrothermal activity are concealed many unanswered questions for the future.

The rocks consist of a deeply eroded basement overlain by voluminous Cenozoic volcanic rocks and sedimentary deposits of lakes and streams. The basement rocks are regionally and thermally metamorphosed volcanic and sedimentary rocks, probably Mesozoic in age, which are intruded by granitic rocks of Cretaceous age. Following the deep erosion that exposed the granitic rocks, sporadic Cenozoic volcanism over a long time produced in the two quadrangles more than a 100 cubic miles of lava and tuff-breccia, most of which is andesitic in petrographic character but in average chemical composition is probably not far from granodiorite. The earliest eruptions were of Hartford Hill Rhyolite Tuff, not present in the Mount Rose quadrangle, and this was followed by andesite flows and tuff-breccias of the Alta Formation, present only locally in the Mount Rose quadrangle. A plutonic rock, the Davidson Granodiorite, invaded the Alta Formation in the Virginia City quadrangle. Next, the voluminous and widespread andesitic flows and tuff-breccias of the Kate Peak Formation of Miocene or Pliocene age were erupted from many vents. During and after these eruptions, sandstone, diatomaceous shale and other fresh-water sedimentary rocks of the Truckee Formation of Pliocene age accumulated in structural basins that were forming between volcanic hills; basaltic lava and rhyolitic pumice were also erupted and interfinger with other deposits of the Truckee Formation. After considerable deformation of the Truckee Formation, including tilting to 30° or more in many places and deep erosion, basaltic andesites of the Lousetown Formation were erupted from many vents in late Pliocene or early Pleistocene time. A few domes of pumiceous rhyolite were also extruded at about the same time. Still later Pleistocene eruptions of andesite and olivine basalt occurred in the Virginia City quadrangle.

Repeated glaciations, at least four in number, occurred in the larger valleys in the Carson Range. Alluvium and landslide debris were also deposited repeatedly over a span of time comparable with the glaciations. The same is true of hot-spring sinter near the northeastern margin of Steamboat Hills.

Rock alteration, generally consisting of some combination of hydrothermal epidote, albite, chlorite, clays, zeolites, calcite, and pyrite, was most widespread and intense prior to deposition of the Truckee Formation, but alteration of different types is still going on at Steamboat Springs. Where the altered rocks contain pyrite, sulfuric acid is formed during weathering, and these rocks are thoroughly bleached.

Intense pre-Cenozoic folding and faulting left the metamorphic rocks dipping 45°–90° in most places and commonly striking between northeast and northwest. Cenozoic block faulting and warping, along with erosion, shaped the present mountainous topography; these processes were underway before Pliocene time and have continued through Pleistocene. The northern and northeastern part of the Carson Range was principally arched up while farther south at Slide Mountain the range rose as a normal fault block. Many normal faults along the eastern front of the Carson Range and western front of the Virginia Range are antithetic, that is, the fault scarps face the ranges rather than the basins. The trough between the Carson and Virginia Ranges contains a chain of basins which are oval in general plan and separated by transverse ridges. The major Cenozoic deformation of which we have a record took place after deposition of the Truckee Formation and prior to eruption of the Lousetown Formation, but the Lousetown also has several hundred feet of structural relief. Erosion has left a subdued topography in the uplands, precipitous slopes on the mountain fronts and, locally, broad pediments around the base of the mountains; deposition continues in some of the basins.

Metallic mineral deposits of pre-Tertiary age include the lead-silver deposits of the Union lead mine and scattered prospects of gold or tungsten. Cenozoic deposits include mercury and gold. In addition there are large reserves of nonmetallic materials including pumice, clayey and siliceous altered rock, basaltic cinder, and alluvial sand and gravel. The thermal waters and heated rocks at Steamboat and elsewhere are an energy resource that has been utilized slightly in the past and may be developed in the future for geothermal power.

INTRODUCTION

LOCATION AND SURFACE FEATURES

The Mount Rose quadrangle, comprising an area of 230 square miles between Reno and Lake Tahoe, is in the southern part of Washoe County at the western border of Nevada (fig. 1). Although the western part of the quadrangle is closely related both topographically and geologically to the Sierra Nevada, the area is entirely within the Great Basin and drains through tributaries of the Truckee River to Pyramid Lake, a shrinking relic of the Pleistocene Lake Lahontan.

From the lowest area, about 4,400 feet in the northeastern part of the quadrangle, the altitude increases to 10,778 feet at the summit of Mount Rose in the Carson Range. This range, which is sometimes considered a part of the Sierra Nevada, occupies more than half of the quadrangle on the western side; it is a northward offshoot of the main mass of the Sierra Nevada, which trends northwestward. West of the quadrangle a deep trough, the southern part of which holds Lake Tahoe, separates the two mountain masses. East of the Carson Range a chain of basins, including Washoe, Pleasant, and Steamboat Valleys, and Truckee Meadows, separates the Carson Range from the semiarid Virginia Range. Steamboat Hills and Huffaker Hills rise within the chain of basins.

The climatic contrast at the eastern base of the Carson Range is abrupt. To the west, yellow pines flourish at intermediate altitude but are succeeded by lodgepole pines at high altitude, up to timberline at 9,000–10,000 feet. To the east, where annual rainfall is generally less than 10 inches, sagebrush and pinyon pine are the normal vegetation. In a few areas, as for example in the northeastern part of Steamboat Hills, the yellow pines, favored by some competitive advantage when growing on acid-altered rock, invade the semiarid climatic zone (Billings, 1950). There are several similar stands growing on altered rock to the east in the Virginia City quadrangle.

Most parts of the quadrangle are readily accessible from Reno, which lies partly within the northern boundary, via Routes 40 and 395 and the Mount Rose Road, State Route 27. The Virginia and Truckee railroad, which connected the Southern Pacific at Reno with Carson City to the south along Route 395, and which once carried supplies and ore for the colorful mining town of Virginia City, was abandoned in 1950. The western and northern areas of the quadrangle are exceptionally rugged and some parts are accessible only on foot.

PURPOSE AND SCOPE

An intensive investigation of the geology and geochemistry of the Steamboat Springs area, which straddles the boundary between the Mount Rose and Virginia City quadrangles (fig. 1), was already in progress in 1948 when Thompson began a study of the two quadrangles to provide the structural and volcanological setting as an essential basis for understanding the history and mechanism of these metal-bearing springs. Two further objectives of the quadrangle mapping were to work out the structure of the basins and ranges near a boundary between the Basin and Range province and the Sierra Nevada and to determine insofar as possible any relations that might exist between igneous activity and structural deformation.

Geologic study of both quadrangles was carried out at the same time, although the Virginia City quadrangle was completed first. A geologic map of the Virginia City quadrangle (Thompson, 1956) was published with a brief text to explain the rocks and structure appearing on the map, but most of the general geologic interpretation was reserved until the map of the Mount Rose quadrangle could also be published. The present report therefore refers frequently to the Virginia City quadrangle, although emphasis is given to relations within the Mount Rose quadrangle.

This report sets forth the results of the regional geologic study and constitutes the general framework for a comprehensive study of the geology and geochemistry of Steamboat Springs.

FIELDWORK AND ACKNOWLEDGMENTS

Fieldwork in the Mount Rose and Virginia City quadrangles was carried on during the field seasons from 1948 to 1952. Most of the mapping has been done by Thompson and his assistants; White mapped Steamboat Hills and the surrounding alluviated areas, including most of Truckee Meadows. J. G. Moore participated effectively in the fieldwork in 1950, 1951, and 1952, and R. A. Pomeroy in 1951. Robert Horton assisted for a time in 1949, W. J. Carr and Frank Campbell for part of the 1950 field season, and R. F. Clark for part of 1952. Others, including R. G. Reeves and William Ebert, lent aid for shorter periods.

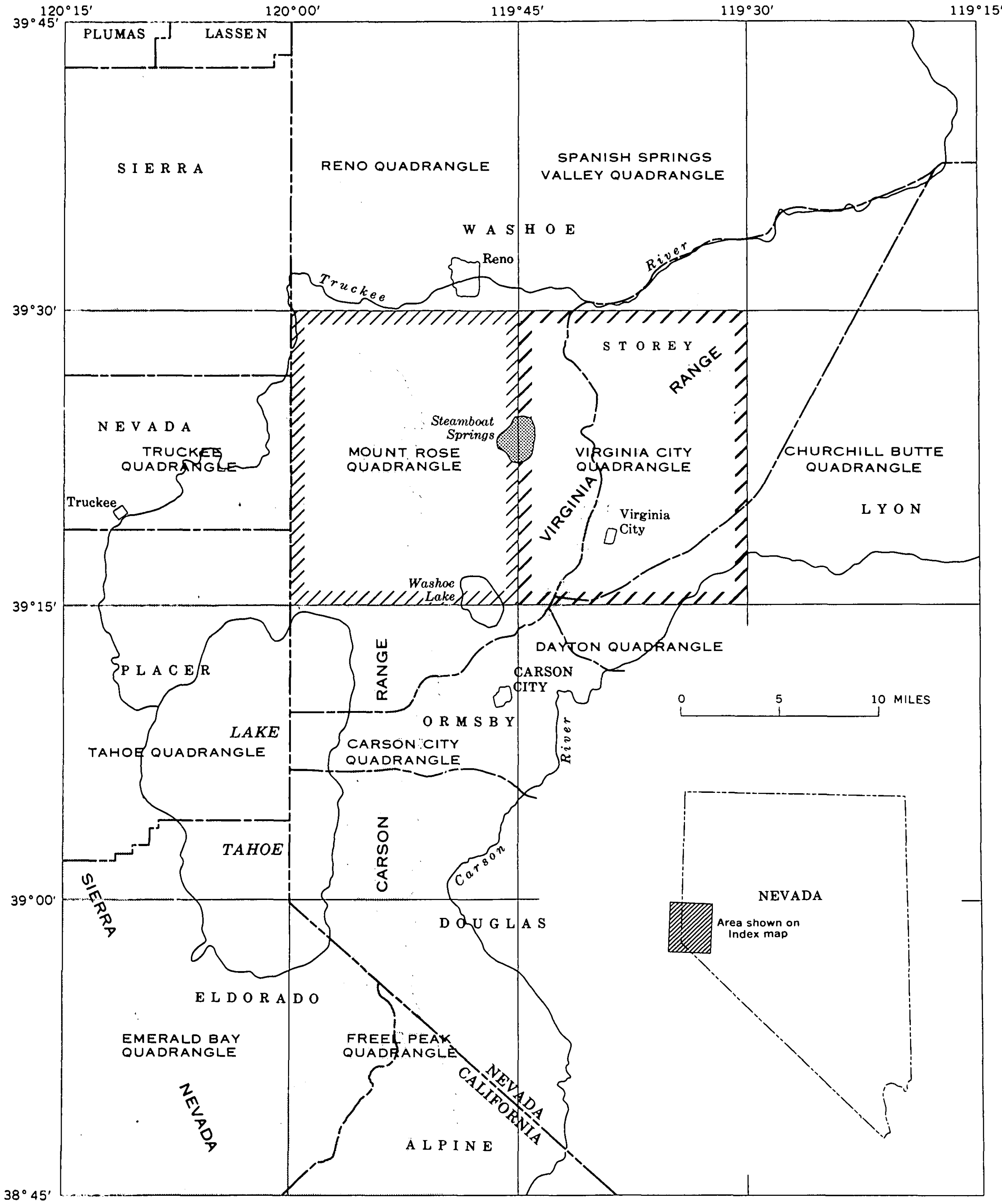

FIGURE 1.—Index map showing location of Mount Rose and Virginia City quadrangles. Shaded area represents mountain areas with altitude greater than 7,000 feet.

Field conferences with F. C. Calkins, who had mapped the Comstock Lode district, aided immeasurably in the early stages of the investigation. V. P. Gianella, who had mapped the Silver City district, gave generously of his wide knowledge of the region. S. W. Muller examined fossils from pre-Tertiary rocks in and immediately south of the Virginia City quadrangle and contributed comparative information on the Mesozoic rocks elsewhere in Nevada. H. G. Ferguson, in a field conference and in many stimulating discussions, focused the writers' attention on fundamental structural problems. Finally, C. H. Sandberg, by means of geophysical studies, helped to clarify structural questions that could not be answered without information from below the surface.

PREVIOUS WORK

Scientists were attracted to the region in the latter half of the 19th century by the Comstock Lode and Steamboat Springs. Summaries of the early work by Gianella (1936), Calkins (1944), and Brannock, Fix, Gianella, and White (1948) make unnecessary any extensive review here, except to note briefly the parallel progress of fundamental geologic knowledge and of geologic understanding of this region.

Von Richtofen (1865, quoted by Becker, 1882) supposed that certain altered rocks in the Comstock Lode district represented a distinct kind of volcanic rock, which he called propylite, meaning gateway, because he thought these rocks were at the bottom or gateway of the Tertiary section. It was a major contribution of Becker's (1882) to discover that propylite was merely an alteration product principally of andesite and had no stratigraphic significance.

Early workers also had great difficulty in distinguishing the Davidson Granodiorite of the Comstock Lode district from the metamorphosed and altered rocks that surround it. The supposed area of the intrusive rock changed greatly from King's map (1870) to Becker's map (1882), and when the contacts were fully understood, the granodiorite appeared on the map by Calkins and Thayer (1945) as an area only a fraction of its former size. One peculiar result of this difficulty is that a rock analysis that King, Becker, and later geologists supposed represented the Davidson Granodiorite is actually an analysis of andesite of the Alta Formation (Thompson, 1956, p. 53). The early geologists correlated the Davidson Granodiorite with the Mesozoic granitic rocks of the Sierra Nevada, and the intrusive relation of the Davidson to Tertiary volcanic rocks was first described by Smith (1912).

Faults on the east side of the Carson Range were described by Russell (1885). Later investigators, including Lindgren (1897, 1911) and Louderback (1903, 1907), disagree on the age of the major displacements, Lindgren regarding the major movements as earlier than the andesite of the Kate Peak Formation and Louderback considering most of the block faulting even younger than post-andesite basaltic rocks of the Lousetown Formation. Reid (1911) carried the hypothesis of young faulting to an extreme by considering most of the topography as formed of small fault blocks. Only Anderson (1909) recognized faults with downthrow to the west (anthithetic faults) along the east front of the Carson Range, and in the same area southwest of Reno he recognized the importance of differential erosion in forming the abrupt transition between gentle ridges and steep mountains. The evidence on which these diverse views of the structure are based can be reconciled only if the deformation was continuous or sporadic over a long period during the latter part of the Cenozoic Era, and this conclusion is one of the principal findings of the present study.

The most important modern geologic studies are those of Gianella (1936) and Calkins and Thayer (1945; also Calkins, 1944). The major aspects of the volcanic stratigraphy were at last deciphered by these men and a foundation was laid for regional mapping. Other important contributions to the geology are referred to in connection with specific topics.

STRATIGRAPHY AND PETROLOGY

Pre-Tertiary granitic and metamorphic rocks crop out in less than a fourth of the quadrangle and of this area only about a tenth is metamorphic rocks (pl. 1). For comparison, only about 2 percent of the Virginia City quadrangle contains pre-Tertiary rocks at the surface, and these are about equally divided between granitic and metamorphic rocks (pl. 2). These old rocks are deeply buried in most places beneath Cenozoic volcanic rocks, which in turn are overlain by Cenozoic sedimentary deposits in about a third of the Mount Rose quadrangle.

PRE-TERTIARY ROCKS

The plutonic rocks, which are mostly granodiorite and quartz monzonite, intrude older sedimentary and volcanic rocks that are deformed, regionally metamorphosed, and further metamorphosed near intrusive contacts.

METAMORPHOSED SEDIMENTARY ROCKS

By far the largest area of the old sedimentary rocks is in Steamboat Hills and in the western salient of the

Virginia Range south of Steamboat Hills. Hornfels, schist, slate, argillite, conglomerate, quartzite, metagraywacke, and limestone are all present. In the Virginia Range east of Little Washoe Lake a partial section, which dips eastward and is unusually well exposed, includes conglomerate, limestone, and dusty black carbonaceous slate and phyllite in ascending order (fig. 2). A few dikes of metamorphosed andesite porphry cut the metamorphosed sedimentary rocks but are not shown separately on the map because of their very small size.

The composition and texture of some of the sandstone indicate that it is tuffaceous (see analyses 1 and 3, tables 1 and 2), and the conglomerate contains abundant pebbles of andesite. Although a large amount of the clastic material is certainly volcanic, the rocks are largely if not entirely water laid, as shown by the rounding of pebbles, the local occurrence of well-bedded sandstone, and the lenses of limestone.

FIGURE 2.—Metamorphosed volcanic conglomerate or tuff-breccia with schistosity diagonal to photograph. Contact with intrusive granodiorite is 20 feet away. Weathering here accentuates the original texture, which is commonly obscure. Southeast of Little Washoe Lake.

TABLE 1.—*Chemical analyses of rocks, Mount Rose, Virginia City, and adjacent quadrangles, Nevada*

[Specific locations of samples are given in table 2. All analyses by rapid methods unless noted]

	Metamorphic rock			Granitic rock				Hartford Hill Rhyolite Tuff			
	Metavolcanic (1)	Inclusion (2)	Metatuff (3)	Granodiorite (4)	Quartz monzonite (5)	Granodiorite (6)	Granodiorite (7)	Tuff (8)	Tuff (9)	Welded tuff (10)	Welded tuff (11)
No.	(35)	(W420-b)	(W421)	(A-17)	(138)	(W419)	(W420-a)	(A-11)	(13)	(267)	(311)
Lab. No.	[1] 52-857	[2] 51-1447	[2] 51-1448	[2] 51-1466	[1] 52-869	[2] 51-1445	[2] 51-1446	[2] 51-1460	[1] 58-855	[1] 52-873	[1] 52-876
Location	CW Comstock	CW Steamboat	CW. Pleasant valley	CW Southwest of Washoe City [5]	CW Comstock	CW Steamboat [6]	CW Steamboat [7]	Comstock Forman shaft	CW Washoe Valley	CW Sutro Springs	CW Carson Airport
SiO_2	51.7	57.5	61.1	65.4	69.2	65.6	65.9	70.8	72.0	73.6	73.8
Al_2O_3	18.3	17.6	21.1	16.2	16.1	16.5	16.7	13.2	14.0	12.4	12.4
Fe_2O_3	.44	2.8	2.1	1.8	1.5	2.2	.60	.84	2.3	.43	.35
FeO	7.7	3.9	3.2	2.4	1.0	2.1	2.2	.98	.06	.37	.30
MgO	5.2	4.0	1.5	1.6	.66	1.7	1.5	.03	.12	.12	.07
CaO	6.7	5.0	1.6	5.2	2.6	3.2	3.8	2.5	.40	.86	.41
Na_2O	4.4	4.3	3.2	3.7	4.5	3.6	3.6	2.1	3.6	3.4	2.8
K_2O	2.2	1.6	2.3	3.5	3.0	2.9	2.8	4.4	5.8	4.7	5.0
TiO_2	.71	.76	.72	.58	.36	.55	.52	.54	.34	.16	.05
P_2O_5	.27	.20	.16	.14	.10	.15	.15	.07	.04	.00	.02
MnO	.12	.18	.07	.07	.06	.06	.16	.01	.02	.02	.04
H_2O	[11] 2.2	[11] .93	[11] 2.9	[5] .44	[11] 1.4	[6] .55	[11] .67	[11] 1.9	[11] 1.4	[11] 3.3	[11] 4.2
FeS_2 [12]	------	------	------	.00	------	------	------	.13	------	------	------
SO_3	------	------	------	------	------	.00	------	------	------	------	------
CO_2	------	------	------	.00	------	.09	------	1.2	------	------	------
Total	100	99	100	101	100	99	99	99	100	99	99
Powder gravity	2.85	2.78	2.73	2.70	2.66	2.69	2.68	2.66	2.50	2.40	2.36
Bulk gravity	2.70	2.66	2.66	2.62	2.60	2.62	2.62	2.50	1.58	2.22	2.06

See footnotes at end of table.

Table 1.—*Chemical analyses of rocks, Mount Rose, Virginia City, and adjacent quadrangles, Nevada*—Continued

	Alta Formation										Davidson Granodiorite	
	Tuff-breccia (12)	Shale (13)	Pyroxene andesite (14)	Hornblende-pyroxene andesite (15)	Andesite (16)	Andesite (17)	Andesite (18)	Andesite (19)	Soda trachyte (20)	Soda trachyte (21)	Granodiorite (22)	Granodiorite porphyry (23)
No	(A-10)	(A-12)	(A-13)	(124)	(C19)	(C159)	(383)	(48)	(W355)	(W422a)	(314)	(360)
Lab. No	[2] 51-1459	[2] 51-1461	[2] 51-1462	[1] 52-867CW	[1] 52-895CW	[1] 52-898CW	[1] 52-880CW	[1] 52-860CW	[2] 51-1432	[2] 51-1449	[3] 53-626C	[1] 52-878CW
Location	Comstock Forman shaft	Comstock Forman shaft	Comstock Forman shaft	Comstock Gold Hill	Sutro tunnel	Sutro tunnel	South of Castle Peak mine	Steamboat Valley	Steamboat Valley	Steamboat Hills	Bullion Ravine	Bullion Ravine
SiO_2	57.6	71.0	55.8	58.6	57.6	57.7	56.2	57.0	64.6	65.0	63.8	63.2
Al_2O_3	18.7	13.8	18.4	16.8	17.7	17.3	16.7	18.1	18.0	18.6	15.6	15.4
Fe_2O_3	5.9	.26	3.3	3.1	2.6	1.9	3.8	2.7	3.5	1.5	2.2	2.0
FeO	1.1	1.3	3.7	2.4	3.3	3.9	2.7	4.2	.33	2.0	1.8	2.3
MgO	2.4	1.1	4.6	1.8	2.5	2.6	.86	2.2	.15	.74	2.0	1.9
CaO	4.8	2.1	5.4	7.0	6.8	6.4	7.2	4.2	1.3	3.1	3.9	4.6
Na_2O	4.4	2.6	3.6	3.7	3.3	3.3	4.0	3.9	5.5	5.0	4.0	4.0
K_2O	1.2	3.1	1.5	2.3	2.0	2.1	2.4	2.6	3.9	3.2	3.6	3.2
TiO_2	.90	.41	.86	.66	.72	.74	.74	.84	.45	.46	.52	.60
P_2O_5	.32	.13	.22	.20	.20	.16	.48	.48	.35	.21	.12	.16
MnO	.15	.16	.16	.13	.10	.14	.24	.50	.15	.14	.08	.08
H_2O	[11] 2.1	[11] 2.4	[11] 1.6	[11] 3.8	[11] 1.8	[11] 2.7	[11] 4.7	[11] 3.5	[11] 1.4	[11] .25	[11] 1.1	[11] 1.2
FeS_2 [12]	.05	.21	.07	----	.02	.04	----	----	----	----	.06	----
SO_3	----	----	----	----	----	----	----	----	----	----	----	----
CO_2	.62	1.4	.88	----	1.8	1.9	----	----	----	----	----	----
Total	100	100	100	100	100	101	100	100	100	100	99	99
Powder gravity	2.76	2.64	2.74	2.70	2.74	2.70	2.73	2.72	2.66	2.58	2.68	2.73
Bulk gravity	2.59	2.42	2.68	2.44	2.70	2.70	2.47	2.60	2.54	2.52	2.61	2.63

	Kate Peak Formation											
	Lava flow (24)	Lava flow (25)	Flow or intrusion (26)	Lava flow (27)	Lava flow (28)	Lava flow (29)	Lava flow (30)	Lava flow (31)	Andesitic scoria (32)	Breccia block (33)	Tuff-breccia (34)	Breccia fragment (35)
No	(A-1)	(A-6)	(A-9)	(335)	(403)	(P8)	(W236b)	(W445)	(W442b)	(62)	(66)	(W432b)
Lab. No	[2] 51-1450	[2] 51-1455	[2] 51-1458	[1] 52-877CW	[1] 52-885CW	[1] 52-894CW	[3] 53-636C	[1] 52-910CW	[1] 52-909CW	[1] 52-862CW	[1] 52-863CW	[1] 52-908CW
Location	Thomas Creek	Lousetown	South of Lousetown	Mount Rose Relay Station	Northwest of Big Meadows	North of Big Meadows	Steamboat	Steamboat Hills	Steamboat Hills	Southeast of Little Washoe Lake	Washoe Summit	Steamboat Hills
SiO_2	62.6	61.2	60.3	53.9	58.8	58.6	56.0	59.3	54.6	58.8	55.4	66.6
Al_2O_3	18.5	16.4	16.9	18.6	17.8	17.2	16.4	16.8	17.0	19.8	19.1	16.5
Fe_2O_3	3.0	2.7	3.7	5.0	3.6	3.0	4.4	3.6	7.0	2.9	4.8	1.2
FeO	1.3	2.7	1.9	2.5	2.7	3.2	1.6	2.4	1.4	3.8	2.2	1.2
MgO	.67	2.8	2.2	3.6	3.2	3.6	2.2	3.0	1.8	1.8	3.0	.52
CaO	5.5	5.2	5.6	8.8	6.2	6.6	6.8	6.7	5.6	4.8	6.2	2.4
Na_2O	4.4	3.0	3.9	3.5	3.4	3.2	4.0	3.6	4.0	2.8	3.1	4.6
K_2O	1.7	2.4	2.1	1.4	2.3	2.4	3.0	2.2	3.5	2.0	.96	4.3
TiO_2	.38	.72	1.1	1.1	.62	.78	1.4	.62	1.4	.97	.70	.25
P_2O_5	.23	.24	.26	.24	.18	.16	.46	.16	.44	.20	.11	.10
MnO	.08	.01	.02	.10	.14	.14	.08	.20	.06	.14	.10	.14
H_2O	[11] .8	[11] 1.5	[11] 1.2	[11] 1.2	[11] 1.3	[11] 1.2	[11] 2.2	[11] .90	[11] 2.6	[11] 2.0	[11] 4.6	[11] 2.1
FeS_2 [12]	----	----	----	----	----	----	----	----	----	----	----	----
SO_3	----	----	----	----	----	----	----	----	----	----	----	----
CO_2	----	----	----	----	----	----	----	----	----	----	----	----
Total	99	99	99	100	100	100	99	99	99	100	100	100
Powder gravity	2.64	2.65	2.68	2.82	2.76	2.71	2.68	2.76	2.72	2.77	2.58	2.52
Bulk gravity	2.48	2.46	2.45	2.55	2.61	2.58	2.50	2.54	2.38	2.65	1.93	2.41

See footnotes at end of table.

TABLE 1.—*Chemical analysis of rocks, Mount Rose, Virginia City, and adjacent quadrangles, Nevada*—Continued

	Kate Peak Formation—Continued							Truckee Formation			Washington Hill Rhyolite	
	Intrusion (36)	Intrusion (37)	Intrusion (38)	Intrusion (39)	Vitrophyre (40)	Vitrophyre (41)	Intrusive vitrophyre (42)	Vitric Tuff (43)	Basalt member (44)	Basalt member (45)	Vitrophyre (46)	Perlite (47)
No.	(A-14)	(55)	(273)	(302)	(384a)	(391)	(461)	(230)	(406)	(455)	(A-5)	(370)
Lab. No.	[1] 51-1463	[1] 52-861CW	[1] 52-874CW	[1] 52-875CW	[1] 52-881CW	[1] 52-882CW	[3] 5-3-633C	[1] 52-871CW	[3] 53-630C	[3] 53-631C	[2] 51-1454	[1] 52-879CW
Location	Comstock quarry	Comstock, Mount Grosh	East side Virginia Range	South of Bailey Canyon	Northeast of Chalk Hills	Bronco Creek	West of Hunter Creek	West side of Chalk Hills	U.S. 40, east of Verdi	U.S. 40, southwest of Verdi	Washington Hill	Southwest of Washington Hill
SiO_2	63.8	60.8	63.0	59.2	71.4	70.8	70.2	69.6	50.4	50.2	76.7	76.0
Al_2O_3	15.9	16.0	15.4	16.4	14.2	14.8	15.0	11.8	16.2	15.7	13.1	12.4
Fe_2O_3	3.2	4.3	2.9	2.7	1.4	.67	2.1	1.8	4.6	4.1	.30	.45
FeO	1.2	1.2	1.5	2.7	.50	.66	.64	1.1	4.0	4.6	.30	.12
MgO	2.2	3.4	2.2	2.6	.46	.24	.64	.26	7.4	7.8	.40	.04
CaO	4.4	5.6	4.3	6.1	1.8	1.8	2.7	.72	7.8	7.6	.90	.40
Na_2O	3.9	3.8	4.0	3.6	3.0	4.2	4.2	1.3	3.0	3.0	3.9	3.6
K_2O	2.5	2.0	2.4	2.3	5.2	3.0	2.6	5.2	.88	1.1	4.2	4.7
TiO_2	.63	.72	.65	.65	.25	.16	.28	.24	1.0	1.0	.13	.08
P_2O_5	.16	.17	.21	.18	.07	.06	.09	.02	.20	.17	.09	.01
MnO	.03	.10	.08	.08	.11	.12	.02	.02	.12	.10	.01	.11
H_2O	[11] 1.3	[11] 2.1	[11] 2.0	[11] 2.5	[11] 2.9	[11] 3.9	[11] 1.4	[11] 7.5	[11] 3.5	[11] 3.5	[11] .73	[11] 3.1
FeS_2[12]	----	----	----	----	----	----	----	----	----	----	----	----
SO_3	----	----	----	----	----	----	----	----	----	----	----	----
CO_2	----	----	----	----	----	----	----	----	----	----	----	----
Total	99	100	99	99	101	100	100	100	99	99	101	101
Powder gravity	2.62	2.64	2.68	2.64	2.43	2.42	2.56	2.33	2.80	2.80	2.46	2.38
Bulk gravity	2.38	2.27	2.42	2.45	2.33	1.92	2.33	1.29	2.74	2.73	2.06	2.09

	Lousetown Formation											
	Upper flows (48)	Lower flows (49)	Darkest flows (50)	Clark Mtn. flows (51)	Steamboat flows (52)	Vent(?) (53)	McClellan flows (54)	Mt. Rose flows (55)	Mt. Rose flows (56)	Intrusion (57)	Truckee River flows (58)	Truckee River flows (59)
No.	(A-7)	(A-8)	(136)	(A-3)	(128-0)	(W415a)	(20)	(70)	(393)	(396)	(457)	(485)
Lab. No.	[2] 51-1456CW	[2] 51-1457	[1] 52-868CW	[2] 51-1452	[4] Report IWC-9	[2] 51-1443	[1] 52-856CW	[1] 52-864CW	[1] 52-883CW	[1] 52-884CW	[3] 53-632C	[3] 53-634C
Location	Near Lousetown	Near Lousetown	South of Lousetown vent	Clark Mountain	Steamboat Hills [8]	East side Truckee Meadows	Southeast of Mound House	Ridge south of Thomas Creek	Bronco Creek	Gray Creek	West of Fleish	Cinder cone southwest of Boca
SiO_2	51.8	51.9	52.6	53.3	54.38	54.8	55.2	51.6	58.9	55.9	52.0	53.5
Al_2O_3	16.8	18.8	18.4	17.9	17.83	17.1	17.4	17.4	17.9	17.8	15.8	16.6
Fe_2O_3	3.8	6.6	4.2	2.7	2.57	4.4	4.8	3.1	3.2	3.6	3.1	4.0
FeO	4.4	1.9	4.0	5.2	4.97	3.0	2.0	5.0	2.8	3.3	5.2	2.9
MgO	7.3	4.8	5.4	4.3	4.12	2.7	3.4	7.2	3.2	5.4	8.4	5.2
CaO	10.9	8.0	10.2	7.2	6.58	7.2	8.8	8.6	5.6	7.6	8.1	6.0
Na_2O	2.8	3.4	3.1	3.8	4.03	4.8	3.9	3.4	4.2	3.6	3.0	4.4
K_2O	.96	1.1	1.0	1.7	2.50	1.5	1.7	1.2	2.3	1.3	.95	3.0
TiO_2	1.0	.96	.82	1.2	1.89	1.0	.94	1.0	.83	.70	.82	1.6
P_2O_5	.34	.39	.15	.48	.57	.40	.34	.22	.32	.18	.16	.82
MnO	.17	.17	.14	.18	.11	.14	.18	.14	.13	.16	.06	.10
H_2O	[11] .85	[11] 1.0	[11] .66	[11] 2.1	[8] .28	[11] 1.7	[11] 1.8	[11] 2.2	[11] .88	[11] .82	[11] 1.7	[11] .50
FeS_2[12]	----	----	----	----	.00	----	----	----	----	----	----	----
SO_3	----	----	----	----	----	----	----	----	----	----	----	----
CO_2	----	----	----	----	.01	----	----	----	----	----	----	----
Total	101	99	101	100	100	99	100	101	100	100	99	99
Powder gravity	2.87	2.84	2.86	2.77	2.80	2.72	2.76	2.81	2.71	2.88	2.83	2.80
Bulk gravity	2.76	2.30	2.61	2.76	2.59	2.66	2.43	2.68	2.52	2.36	2.75	2.50

TABLE 1.—*Chemical analyses of rocks, Mount Rose, Virginia City, and adjacent quadrangles, Nevada*—Continued

	Steamboat Hills Rhyolite				Mustang Andesite	McClellan Peak Olivine Basalt				Hydrothermally altered rocks	
	Pumiceous rhyolite	Perlitic rhyolite	Pumiceous rhyolite	Pumiceous rhyolite		Long Valley flow	Vent	Long Valley flow	Long Valley flow	Silicified andesite	Altered andesite
	(60)	(61)	(62)	(63)	(64)	(65)	(66)	(67)	(68)	(69)	(70)
No.	(A-15)	(A-16)	(111)	(112)	(A-2)	(A-4)	(155)	(261)	(427)	(431a)	431b)
Lab. No.	[2] 51-1464	[2] 51-1465	[1] 52-865CW	[1] 52-866CW	[2] 51-1451CW	[2] 51-1453CW	[1] 52-870CW	[1] 52-872CW	[1] 52-889CW	[1] 52-890CW	[1] 52-891CW
Location	Dome in Steamboat Hills [9]	Dome, east side of Truckee Meadows [10]	Dome, east side Virginia Range	Satellite body, east Virginia Range	Clark Mountain	Long Valley	Mount McClellan	Tumulus east of Long Valley	North side Truckee River	Upper zone west of Washington Hill	Middle zone west of Washington Hill
SiO_2	75.5	75.8	76.2	74.5	56.2	49.8	48.6	48.4	49.9	97.4	79.8
Al_2O_3	12.6	12.7	13.7	13.7	17.2	16.1	15.2	16.0	16.5	1.3	1.9
Fe_2O_3	.36	.25	.30	.21	4.7	2.6	3.1	2.6	2.1	.16	14.7
FeO	.13	.25	.34	.41	2.1	6.0	6.4	6.5	6.2	.16	.09
MgO	.13	.20	.04	.12	2.8	9.1	10.2	8.5	8.6	.16	.08
CaO	.85	.85	.38	.66	7.0	9.5	9.2	9.6	9.6	.14	.10
Na_2O	3.8	3.8	3.9	3.9	4.0	3.1	3.0	3.2	3.0	.07	.07
K_2O	4.6	4.5	4.6	4.6	2.3	1.8	1.7	1.5	1.7	.06	.06
TiO_2	.13	.11	.07	.04	1.1	1.5	2.0	1.9	1.6	.72	.98
P_2O_5	.01	.09	.00	.00	.46	.47	.50	.48	.44	.03	.11
MnO	.05	.10	.04	.06	.05	.18	.14	.10	.14	.00	.00
H_2O	[9] 2.37	[10] 2.41	[11] 1.7	[11] 2.9	[11] .70	[11] .35	[11] .6	[11] .5	[11] .34	[11] 1.0	[11] 3.0
FeS_2[12]	.00	.00	----------	----------	----------	----------	----------	----------	----------	----------	----------
SO_3	----------	----------	----------	----------	----------	----------	----------	----------	----------	.12	.20
CO_2	.02	.00	----------	----------	----------	----------	----------	----------	----------	<.05	<.05
Total	101	101	101	101	99	100	101	99	100	101	101
Powder gravity	2.38	2.40	2.41	2.30	2.76	2.94	2.94	2.95	2.90	2.60	2.79
Bulk gravity	1.84	2.22	2.24	1.07	2.46	2.76	2.70	1.50	----------	1.92	2.42

	Hydrothermally altered rocks—Continued										
	Altered andesite	Altered andesite	Silicified andesite	Altered andesite	Altered andesite	Altered andesite	Alerted hornblende andesite	Altered hornblende andesite	Slightly altered	Pyritized andesite	Argillized andesite
	(71)	(72)	(73)	(74)	(75)	(76)	(77)	(78)	(79)	(80)	(81)
No.	(M51)	(424)	(433b)	(W384)	(W387-225)	(W387-242)	(W387-586)	(W387-622)	(W387-656)	(C179)	(C185)
Lab. No.	[1]52-893CW	[1]52-888CW	[1]52-892CW	[2]51-1436CW	[2]51-1437CW	[2]51-1438CW	[2]51-1439CW	[2]51-1441CW	[2]51-1441CW	[1]52-900CW	[1]52-901CW
Location	Lower zone west of Washington Hill	SW. Washington Hill	North part Long Valley	Pit, Young hole, north of Virginia City	Young hole, 225 ft deep	Young hole, 242 ft deep	Young hole, 586 ft deep	Young hole, 622 ft deep	Young hole, 656 ft deep	Sutro tunnel, 14,734 ft	Sutro tunnel, 15,330 ft
SiO_2	78.8	52.1	99.2	72.5	57.2	61.5	56.6	59.4	57.7	73.5	76.0
Al_2O_3	2.5	24.2	.46	18.6	17.4	12.9	16.6	17.4	16.3	14.7	13.2
Fe_2O_3	10.3	4.4	.05	.02	.0	.3	3.1	3.1	3.0	.28	.16
FeO	.16	.14	.08	.13	1.6	2.3	3.2	2.8	2.8	.56	.48
MgO	.04	.20	.02	.72	3.5	3.8	3.1	2.9	3.3	.58	.89
CaO	.00	.80	.14	.25	2.0	2.2	6.6	5.4	6.7	.89	1.3
Na_2O	.16	.59	.07	.24	2.1	.82	3.2	3.6	3.8	3.8	.13
K_2O	.06	1.0	.07	.10	1.7	3.1	2.1	1.8	1.8	4.4	3.8
TiO_2	5.4	.88	.52	.92	.64	.67	.80	.65	.80	.24	.24
P_2O_5	.18	.16	.06	.22	.19	.13	.27	.22	.25	.04	.05
MnO	.03	.00	.00	.06	.14	.16	.22	.18	.18	.08	.02
H_2O	[11] 2.0	[11] 15.2	[11] .65	[11] 4.7	[11] 9.9	[11] 8.6	[11] 2.6	[11] 1.7	[11] 1.9	[11] 1.9	[11] 3.47
FeS_2[12]	----------	----------	----------	----------	6.4	3.4	.07	.06	.11	.45	.77
SO_3	.38	.02	.03	.18	----------	----------	----------	----------	----------	<.1	<.1
CO_2	<.05	<.05	<.05	----------	<.05	<.05	.32	.07	<.05	<.05	.40
Total	100	100	101	99	103	100	99	99	99	102	101
Powder gravity	2.38	2.24	2.61	2.70	2.56	2.54	2.68	2.70	2.75	2.65	2.65
Bulk gravity	----------	1.52	2.57	1.70	2.34	2.38	2.66	2.66	2.66	2.54	2.25

[1] Analysts: S. M. Berthold, H. F. Phillips, C. C. Alexander, and E. A. Nygaard, U.S. Geol. Survey.

[2] Analysts: Leonard Shapiro, S. M. Berthold, and E. A. Nygaard, U.S. Geol. Survey.

[3] Analysts: H. F. Phillips, Leonard Shapiro, and K. E. White, U.S. Geol. Survey.

[4] Analyst: W. W. Brannock, U.S. Geol. Survey; by standard methods.

[5] Includes, in percent, H_2O^+, 0.40; H_2O^-, 0.04; also determined: Cl, 0.04; F, 0.05; by quantitative spectrographic analysis: B, 0.002; Cu, 0.002; Pb, 0.002; Co, 0.001; Ni, 0.0007; Ga, 0.001; Cr, 0.002; V, 0.01; Sc, 0.0006; Y, 0.003; Zr, 0.01; Be, 0.0008; Sr, 0.01; Ba, 0.1; Li, 0.006; Rb, 0.03; Cs,[13]0.002.

[6] Includes, in percent, H_2O^+, 0.50; H_2O^-, 0.05; also determined: Cl, 0.02; F, 0.03; by quantitative spectrographic analysis: B, 0.002; Cu, 0.003; Hg, 0.00002(?); Pb, 0.001; Co, 0.001; Ni, 0.0009; Ga, 0.001; Cr, 0.002; V, 0.01; Sc, 0.0006; Y, 0.002; Zr, 0.006; Sr, 0.02; Ba, 0.2; Li, 0.0007; Rb, [13]0.02.

[7] By quantitative spectrographic analysis, in percent: Cu, 0.002; Pb, 0.001; Co, 0.001; Ni, 0.001; Ga, 0.0009; Cr, 0.002; V, 0.01; Sc, 0.0004; Y, 0.002; Zr, 0.008; Sr, 0.01; Ba, 0.2; B, 0.002; Li, 0.002; Rb, 0.03; Cs, [13]0.0009.

[8] Includes, in percent, H_2O^+, 0.16; H_2O^-, 0.12; also determined: Cl, 0.01; F, 0.06; by quantitative spectrographic analysis: B, 0.001; Cu, 0.004; Pb, 0.002; Co, 0.002; Ni, 0.005; Ga, 0.001; Cr, 0.004, V, 0.02; Sc, 0.001; Y, 0.004; La, 0.01; Zr, 0.008; Sr, 0.01; Ba, 0.2; Li, 0.002; Rb, 0.03; Cs [13]0.0009.

[9] Includes, in percent, H_2O^+, 2.31; H_2O^-, 0.06; also determined: Cl, 0.03; F, 0.06; by quantitative spectrographic analysis: B, 0.006; Cu, 0.0005; Pb, 0.005; Ga, 0.001; Cr, 0.0003; Sc, 0.0002; Y, 0.003; Zr, 0.01; Be, 0.002; Ba, 0.004; Li, 0.008; Rb, 0.03, Cs, [13]0.0009.

[10] Includes, in percent, H_2O^+, 2.36; H_2O^-, 0.05; also determined: Cl, 0.02; F, 0.05; by quantitative spectrographic analysis: B, 0.006; Cu, 0.0003; Pb, 0.002; Ga, 0.001; Cr, 0.0003; Sc, 0.0002; Y, 0.003; Zr, 0.01; Be, 0.002; Ba, 0.002; Li, 0.008; Rb, 0.04; Cs, [13]0.001.

[11] Reported ignition loss corrected for gain due to oxidation of FeO; also corrected for CO_2 and SO_3 if reported separately. See footnote 13.

[12] Sulfur reported as FeS_2 and iron corrected. See footnote 13.

[13] H_2O^+, H_2O^-, CO_2, Dl, F, and S, by L. M. Kehl; B, by P. R. Barnett; other spectrographic analyses by H. J. Rose.

TABLE 2.—*Location of analyzed rocks, Mount Rose, Virginia City, and adjacent quadrangles, Nevada*

No.	Quadrangle	Quarter	Section	Township (North)	Range (East)
Metamorphic rocks					
1 (35)	Virginia City	SE.	8	16	21
2 (W420-b)	do	SW.	33	18	20
3 (W421)	Mount Rose	SE.	5	17	20
Granitic rocks					
4 (A-17)	Mount Rose	NW.	23	17	19
5 (138)	Virginia City	SW.	8	16	21
6 (W419)	do	SE.	33	18	20
7 (W420-a)	do	SW.	33	18	20
Hartford Hill Rhyolite Tuff					
8 (A-11)	Virginia City	NE.	5	16	21
9 (13)	Carson City	SE.	19	20	16
10 (267)	Virginia City	NE.	4	17	22
11 (311)	Dayton		33	16	20
Alta Formation					
12 (A-10)	Virginia City	NE.	5	16	21
13 (A-12)	do	NE.	5	16	21
14 (A-13)	do	NE.	5	16	21
15 (124)	do	NE.	6	16	21
16 (C-19)	do	[1]			
17 (C159)	do	[2]			
18 (383)	do	SE.	30	18	21
19 (48)	do	SW.	34	18	20
20 (W355)	do	SW.	34	18	20
21 (W422-a)	Mount Rose		36	18	19
Davidson Granodiorite					
22 (314)	Virginia City	NE.	25	17	20
23 (360)	do	NW.	32	17	21
Kate Peak Formation					
24 (A-1)	Mount Rose	SW.	29	18	19
25 (A-6)	Virginia City	NE.	16	18	21
26 (A-9)	do	NE.	9	17	21
27 (335)	Mount Rose	NE.	22	17	18
28 (403)	do	NE.	9	18	18
29 (P8)	do	SW.	3	18	18
30 (W436-b)	do	NE.	33	18	20
31 (W445)	do	SW.	12	17	19
32 (W442-b)	do	NE.	1	17	19
33 (62)	do	NW.	32	17	20
34 (66)	do	NE.	24	17	19
35 (W432-b)	do	SE.	1	17	19
36 (A-14)	Virginia City	NE.	33	17	21
37 (55)	do	NE.	4	16	21
38 (273)	do	NW.	24	17	21
39 (302)	do	NE.	11	17	20
40 (384-a)	do	NW.	29	18	22
41 (391)	Mount Rose	SW.	33	18	18
42 (461)	do	SE.	27	19	18
Truckee Formation					
43 (230)	Virginia City	NW.	25	18	21
44 (406)	Reno	SE.	9	19	18
45 (455)	Mount Rose	NE.	30	19	18
Washington Hill Rhyolite					
46 (A-5)	Virginia City	SW.	34	19	21
47 (370)	do	NW.	4	18	21
Lousetown Formation					
48 (A-7)	Virginia City	NE.	28	18	21
49 (A-8)	do	SW.	33	18	21
50 (136)	do	NE.	27	18	21
51 (A-3)	do	NW.	26	19	21
52 (W128-O)	Mount Rose	SE.	32	18	20
53 (W415-a)	Virginia City	NE.	27	18	20
54 (20)	Dayton	NW.	1	15	20
55 (70)	Mount Rose	NE.	36	18	18
56 (393)	do	SE.	30	18	18
57 (396)	do	SE.	33	17	18
58 (457)	Truckee	SE.	30	19	18
59 (485)	do	NE.	32	18	17

[1] Sutro tunnel, 1,668 ft from portal.
[2] Sutro tunnel, 13,500 ft from portal.

TABLE 2.—*Location of analyzed rocks, Mount Rose, Virginia City, and adjacent quadrangles, Nevada*—Continued

No.	Quadrangle	Quarter	Section	Township (North)	Range (East)
Steamboat Hills Rhyolite					
60 (A-15)	Mount Rose	SE.	1	17	19
61 (A-16)	Virginia City	NW.	23	18	20
62 (111)	do	NW.	16	17	22
63 (112)	do	SE.	8	17	22
Mustang Andesite					
64 (A-2)	Virginia City	NE.	26	19	21
McClellan Peak Olivine Basalt					
65 (A-4)	Virginia City	NW.	27	19	21
66 (155)	do	SW.	12	16	20
67 (261)	do	NE.	17	18	22
68 (427)	Spanish Springs Valley.	NW.	16	19	21
Hydrothermally altered rocks					
69 (431-a)	Virginia City	SW.	28	19	21
70 (431-b)	do	SW.	28	19	21
71 (M51)	do	SW.	28	19	21
72 (424)	do	SE.	5	18	21
73 (433-b)	do	SE.	20	19	21
74 (W384)	do	NE.	6	17	21
75 (W387-225)	do	NE.	6	17	21
76 (W387-242)	do	NE.	6	17	21
77 (W387-586)	do	NE.	6	17	21
78 (W387-622)	do	NE.	6	17	21
79 (W387-656)	do	NE.	6	17	21
80 (C179)	do	[3]			
81 (C185)	do	[4]			

[3] Sutro tunnel, 14,734 ft from portal.
[4] Sutro tunnel, 15,330 ft from portal.

The most intense metamorphism is near granitic contacts, such as the intrusive contact forming the southern margin of the metamorphic rocks, and has produced coarse hornfels, contact schist, marble, and tactite. The original texture of even the conglomerate has been almost obliterated. Andalusite crystals several millimeters in length are commonly visible, and some of the rocks also contain abundant sillimanite. The smaller patches of metamorphic rocks east and northeast of Mount Rose consist of coarse hornfels, biotite schist, garnet-biotite schist, and quartzite. The small inclusion or roof pendant on the west side of Slide Mountain contains marble and biotite schist. The exposures shown near the northern edge of plate 1, midway between the Truckee River and Hunter Creek, contain metamorphosed conglomerate. Near granitic contacts the metamorphic rocks are generally cut by a profusion of quartz veins and granitic, pegmatitic, and aplitic dikes. Within a few inches to a few feet of the contacts, porphyroblasts of hornblende and plagioclase, identical in size and appearance to crystals of the same minerals in the granitic rocks, are commonly abundant.

Similar porphyroblasts are also common in inclusions of metamorphic rock, which are scattered through the granitic rock (see analysis 2, table 1). The size of these inclusions is commonly a fraction of an inch to a foot, though larger inclusions are not rare. Where foliation can be seen in the inclusions, it is not parallel to that in the nearby metamorphic rocks nor to that in other inclusions; rotation of the fragments within the granitic magma is indicated.

Some of the schist and hornfels is pyritic, and weathering of the pyrite has produced local acid alteration and bleaching. In upper Whites Creek, spring water from an oxidized pyritic schist has locally cemented the alluvium with iron oxide and silica.

The strong deformation and obscure stratigraphic relations make an accurate measure of thickness of the old sedimentary rocks impossible, but a thickness of several thousand feet is probable.

No direct evidence of the age range is available, except that they are older than the granitic rocks and are therefore presumably pre-Cretaceous. The limestone and carbonaceous slate are similar to limestone and carbonaceous slate south of American Flat in the Virginia City quadrangle. The carbonaceous slate also resembles that in Triassic rocks south of the Virginia City quadrangle. Some of the metamorphosed sedimentary rocks are therefore probably Triassic but part of them may be as old as Paleozoic. Lindgren (1897) found Jurassic and Triassic ammonites in metamorphosed sedimentary rocks in the Truckee quadrangle, immediately west of the Mount Rose quadrangle.

METAVOLCANIC ROCKS

Metavolcanic rocks, which crop out more extensively in and north of the Virginia City quadrangle, are exposed near the northwestern corner of the Mount Rose quadrangle. Elsewhere, as in the canyon of Galena Creek through Steamboat Hills and in the narrows between Steamboat and Pleasant Valleys, small amounts of metavolcanic rocks are included with the metamorphosed sedimentary rocks. In the canyon of the Truckee River in the northwestern corner of the quadrangle, the bold outcrops and very-dark-greenish-gray color, which are characteristic of the metavolcanic rocks, are spectacularly shown. The original rocks were basalt and andesite. Amygdules (west of Truckee River on south side of outcrop area) and faint traces of stratification clearly indicate that the volcanic rocks were extrusive. A relict porphyritic appearance of some of the material suggests the texture of andesite, and part of this material appears to be pyroclastic breccia. A minor amount of metamorphosed sedimentary rock appears to be intercalated with the metavolcanic rocks.

Thorough recrystallization converted the rocks to hornfels composed chiefly of sodic plagioclase, epidote, mica, chlorite, and amphibole.

The age of the metavolcanic rocks, like that of the old sedimentary rocks, is not known except that they are older than the granitic rocks that intrude them and are therefore Cretaceous or older. The close association of volcanic and sedimentary rocks in the Virginia City and Mount Rose quadrangles and the contribution of much volcanic debris to the sedimentary rocks suggest approximate contemporaneity. Calkins (1944, p. 9) noted andesite interbedded with the sedimentary rocks in the Comstock Lode district. The metamorphic rocks north of the Virginia City quadrangle also include both volcanic and sedimentary material. Lindgren (1897) also found evidence that pre-granitic volcanic rocks in the Truckee quadrangle, immediately west of the Mount Rose quadrangle, were interstratified with metamorphosed sedimentary rocks. The evidence thus suggests at least partial contemporaneity and an intertonguing relation between the metamorphosed sedimentary and volcanic rocks.

GRANODIORITE AND RELATED ROCKS

Plutonic rocks that range from granodiorite to quartz monzonite crop out on the west side of the Virginia Range, in Steamboat Hills, in several places in the northwestern part of the Mount Rose quadrangle, and in a large area in the core of the Carson Range. The most abundant rock type is biotite-hornblende granodiorite containing about twice as much plagioclase as potassium feldspar (analyses 4, 6, 7, table 1), but quartz monzonite containing plagioclase only slightly in excess of potassium feldspar is also plentiful (analysis 5, table 1). In the area east of Slide Mountain and north of Ophir Creek, for example, about 60 percent of the feldspar is plagioclase (analysis 4, table 1). Gradational changes in mineral composition are the rule, rather than sharp boundaries.

Biotite and hornblende range widely in their proportion, although both are generally present. Sphene in bright yellow to brown grains is a common accessory mineral. In some areas deuteric epidote is conspicuous on fracture planes

The batholith contains veins of quartz and dikes of granitic rock, aplite, and pegmatite. In the pegmatite the most conspicuous mineral other than quartz and feldspar is generally black tourmaline.

Along the contact with metamorphic rocks in upper Whites Creek the granitic rock is unusually mafic. This diorite and quartz diorite mass is finer grained than the normal granodiorite into which it grades. It forms a border zone with a maximum width of about a quarter of a mile. The gradational change from normal granodiorite to diorite and the association of the latter with metamorphic rocks suggest that the border zone was formed by contamination or reaction of granodiorite with metamorphic rocks.

In the northwestern part of the quadrangle, southwest of Puny Dip Canyon, a large andesite intrusion includes many small and several enormous blocks of granitic rock. Most of these inclusions (the largest of which might alternatively be regarded as roof pendants) are of the normal granodiorite or quartz monzonite, but the elongate mass shown on the map is composed of a finer grained granitic rock that resembles the Tertiary Davidson Granodiorite of the Virginia City quadrangle. The rock contains euhedral plagioclase crystals, quartz that is small in amount and interstitial to the feldspar, and amphibole; biotite is lacking. The contacts are poorly exposed. A block of the same rock, a few hundred feet to the north and too small to be shown on the map, contains inclusions of normal granodiorite. This evidence shows that the finer grained granitic rock is younger than the normal granodiorite, but whether it merely represents a satellite of the main pluton or a much later Tertiary intrusion is not certain.

The granitic rocks rarely show prominent foliation or lineation and the faint evidence of such internal structure is commonly concealed by the granular disintegration of the rock. Consequently no systematic study of these structures was attempted in the field mapping, but individual observations of foliation were recorded on the map. The orientation of joint systems was recorded only where especially well developed or of unusual significance (as adjacent to the landslide scar on Slide Mountain). The observations of foliation indicate steep dips and a preferred strike between northeast and northwest but with considerable scatter.

The age of the granitic rocks of the Sierra Nevada has long been thought to be Jurassic or Cretaceous. Radioactive age determinations in recent years have indicated that the plutonic rocks of the central and eastern Sierra Nevada are Cretaceous (Larsen and others, 1954; Curtis and others, 1958).

TERTIARY SYSTEM

Rhyolitic tuffs (analyses 8 through 11, table 1) lie at the base of the Tertiary rocks in parts of the quadrangles to the west, south, and east but are absent in the Mount Rose quadrangle and are not discussed further. Lying on and intertonguing with the rhyolitic tuffs in the Virginia City quadrangle is the enormous volcanic complex of the Alta Formation, but the Alta is represented in the Mount Rose quadrangle only by a soda trachyte in Steamboat Hills. Elsewhere in the quadrangle if present at all, it is small in amount and so similar to the Kate Peak Formation that distinguishing the two is not practical nor particularly useful.

The Kate Peak is the most voluminous and extensive of the volcanic formations, and in most places it lies directly on the pre-Tertiary granitic and metamorphic rocks. The Truckee Formation, which intertongues with and overlies the Kate Peak Formation, is composed in large part of sedimentary debris derived from the Kate Peak.

DUAL CLASSIFICATION OF VOLCANIC ROCKS

Two classifications of volcanic rocks are in current use. One uses the mineralogic composition of the phenocrysts and the character of the groundmass to distinguish basalts, andesites, and rhyolites petrographically; the other uses the chemical composition and the norm calculated from it to define the classes chemically. The latter method makes basalt the exact chemical equivalent of gabbro, andesite that of diorite, rhyodacite that of granodiorite and so forth. Traditionally rocks in the region under discussion have been classified by the petrographic method, and on that basis common rocks like those of the Kate Peak Formation are predominantly andesite (Lindgren, 1897; Gianella, 1936; Calkins, 1944). Chemically, however, the same rocks are rhyodacites. Nockolds (1954, p. 1009) points out that, contrary to the generally accepted opinion, volcanic rocks corresponding in composition to granodiorite (the rhyodacites) are common. In the present study, where names are used that are based on chemical composition, Nockolds' classification (1954, p. 1008) is followed.

A petrographic classification may well be the most desirable for field use because such a classification best expresses subdivisions that can be observed megascopically; in this area, for example, dark olivine-bearing rocks, which form cinder cones and relatively thin extensive flows with vesicular tops, range in chemical composition from basalt to alkalic andesite but form a natural basaltic group petrographically. Rocks that are readily divided petrographically into pyroxene andesite and biotite andesite, each with a very distinctive appearance, may all fall chemically into the rhyodacite class. On the one hand, petrographic names based primarily on composition of the phenocrysts may be most useful for stratigraphic and structural studies in the field, and on the other hand, names based on chemical composition may be most useful in studies of the genesis of volcanic rocks.

No attempt to settle problems of the dual classification is intended here. Where names based on chemical composition are used, they are indicated by modifiers (that is, chemically a rhyodacite, or rhyolite in chemical composition); rock names not so designated are based on petrographic character, following the established practice.

The important point is that the petrographic classification might imply misleading correlations with plutonic rocks. Rocks that have long and consistently been designated andesite in this region have an average composition much closer to granodiorite than to diorite.

ALTA FORMATION

The Alta Formation derives its name from the Alta shaft on the Comstock Lode (Gianella, 1936, p. 53). In the type area a thick section of andesitic lavas and pyroclastics is exposed (analyses 12 to 17, table 1), but the volcanic pile thins to the west and contains fewer varieties of rock. One of the rock types on the western flank of the Virginia Range, near the old mining district of Jumbo in the Virginia City quadrangle, is a soda trachyte (Thompson, 1956, p. 51), and near Steamboat Valley similar soda trachyte is the principal rock type included in the Alta Formation (analysis 20, table 1). To the west, in Steamboat Hills, the same soda trachyte lies at the base of the Tertiary section and is the only rock in the Mount Rose quadrangle included in the Alta Formation; in places it is overlain uncomformably by the Kate Peak Formation.

The fresh soda trachyte is black and semivitreous (analysis 21, table 1) but in most places it is devitrified to a light or medium purple-gray rock stained with iron oxide on fractures. Small tabular phenocrysts of oligoclase, with or without a few needles of hornblende and small flakes of biotite, are set in a fine-grained groundmass consisting dominantly of tiny oligoclase laths in a glassy or devitrified matrix. The phenocrysts and elongated plagioclase crystals of the ground mass generally show a striking alinement intersected at moderate angles by an intimate system of fractures that give the rock a platy parting. Realinement of groundmass plagioclase adjacent to the fractures indicates they were formed by shearing during the late stages of movement of the lava flows. The chemical composition of the soda trachyte (analyses 20 and 21, table 1) is that of a soda rhyolite or soda rhyodacite, because quartz in the norm is 15–16 percent.

Although the Alta Formation has been separately mapped only in Steamboat Hills, it is possible that some Alta is present elsewhere in the Mount Rose quadrangle in areas mapped as Kate Peak Formation. In the Virginia City quadrangle the range of rock types in the two formations overlaps; structural discordance, though marked locally, is slight or nonexistent on a regional basis; and the degree of alteration, relied on by previous workers, varies more from area to area than from one formation to the other. Evidence that the two formations may intertongue was found in the Virginia City quadrangle (Thompson, 1956, p. 52).

Fossil leaves found in a sedimentary member of the Alta Formation in the Virginia City quadrangle suggest an Oligocene age (Axelrod, 1949, p. 1935).

KATE PEAK FORMATION

The Kate Peak Formation is named for Kate Peak in the Virginia City quadrangle (Gianella, 1936, p. 69). The variety of rocks increases as the formation is traced from the type locality at Kate Peak into larger regions of the Virginia City and Mount Rose quadrangles; the extension of the name to these areas is justified by the nearly continuous exposures and the abundance of the most distinctive rock types, which are (1) flows containing phenocrysts of plagioclase that are abundant (commonly 30 percent or more of the rock), large (3–5 mm), and nearly equidimensional; phenocrysts of pyroxene, with or without hornblende; and, occasionally, biotite; and (2) crudely stratified tuff-breccias containing heterogeneous blocks generally similar to the flow rocks (fig. 3). The larger blocks are subrounded "boulders" that are resistant to erosion and are generally strewn prominently on surfaces that are underlain by or are downslope from tuff-breccia. One of the less common varieties of the Kate Peak rocks, a vitrophyre, has been mapped separately as a member.

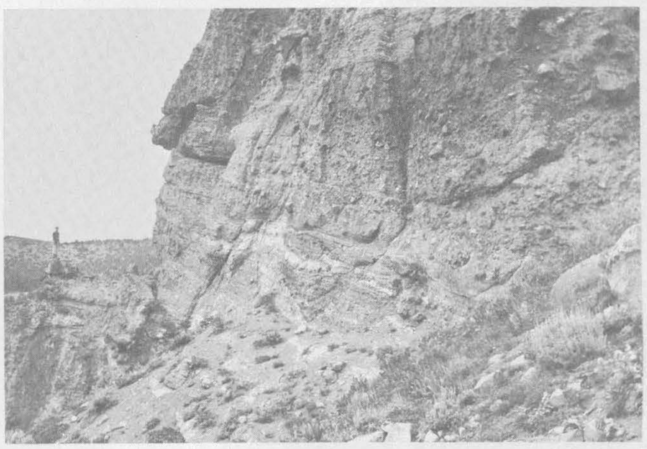

FIGURE 3.—Crudely stratified tuff-breccia of the Kate Peak Formation in the upper drainage basin of Thomas Creek.

Of the flows, flow breccias, intrusive bodies, and beds of tuff-breccia that constitute this great volcanic series, the tuff-breccias are especially abundant and characteristic; they are magnificently exposed in the northwestern part of the quadrangle in the Truckee River canyon. Flows, which generally form bold outcrops, are well exposed on Mount Rose, along the northeastern margin of the Carson Range in strike ridges near the contact with the Truckee Formation, and at many other places throughout the quadrangle (fig. 4). In some places flows grade into flow breccias which, in contrast to the tuff-breccias, are monolithologic.

The rocks of the Kate Peak Formation are diverse in appearance, mineralogy, and chemical composition. (See analyses 24 to 42, table 1.) The color ranges from light buff gray and pink to dark brown and red. Of

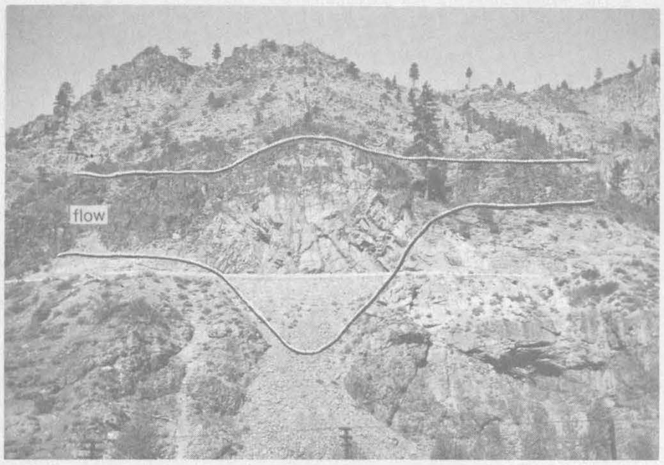

FIGURE 4.—Flow of andesite of the Kate Peak Formation filling an ancient valley cut in tuff-breccia of the Kate Peak. The flow is covered with later tuff-breccia of the Kate Peak. Flow structure is not consistently horizontal but instead conforms to the shape of the ancient valley. View is to the northwest in the Truckee Canyon at the mouth of Gray Creek, a mile west of the Mount Rose quadrangle; old Route 40 extends across center of photograph.

the mafic varieties, the most abundant is a pyroxene andesite that occurs in flows and pyroclastic beds in an area centered near the Mount Rose Relay Station. This rock has the chemical composition of an andesite with 54 percent silica (analysis 27, table 1). The phenocrysts are dark-green pyroxene and plagioclase which is zoned from labradorite to andesine. Although this rock resembles a common variety of andesite of the Alta Formation of the Virginia City quadrangle in texture and composition, it is interstratified with typical hornblende-bearing rocks of the Kate Peak Formation. North of Big Meadows, lavas bearing phenocrysts of pyroxene and a little hornblende contain 59 percent silica (analyses 28 and 29, table 1), and in Thomas Creek hornblende-pryoxene andesite contains 63 percent silica and has the composition of rhyodacite (analysis 24, table 1). The groundmass of all these rocks is aphanitic to glassy and is commonly microvesicular. Although the general range of mineralogic and chemical composition is represented by the rocks described, some of the flows, pyroclastic beds, and intrusions in the northern and western parts of the quadrangle contain biotite, and Lindgren (1897) mentions biotite andesites in the Truckee quadrangle. Biotite is not as common in the Kate Peak Formation of the Mount Rose quandrangle as in the Virginia City quadrangle.

VITROPHYRE MEMBER

A vitrophyre containing biotite as the only important phenocrysts covers a small area along Bronco Creek. It is light gray and porous, in places almost pumiceous. The silica content is 71 percent (analysis 41, table 1). The contacts are everywhere obscured, but from its general position the vitrophyre appears to be interstratified with normal flows and pyroclastic beds of the Kate Peak Formation. The vitrophyre is similar in composition to the vitrophyre member of the Kate Peak Formation in the Virginia City quadrangle (analysis 40, table 1) but is fresher and more vesicular.

The vitrophyre resembles the intrusive rock of the large body west of Hunter Creek in the northern part of the quadrangle (analysis 42, table 1). This intrusion contains a much more silicic rock than most of the intrusions of the Kate Peak Formation. Blocks of similar material are also abundant in some of the pyroclastic beds of the Kate Peak Formation.

STREAM GRAVELS

Lenses of water-worn gravel and sand are fairly common in the bedded tuff-breccias but are generally of very small extent. More extensive gravels are interbedded with the pyroclastics north of the word "Thomas" on Thomas Creek, where some beds are shown on the map to help define the structural deformation. The gravel, which ranges in size from small pebbles to boulders more than a foot in diameter, is composed almost entirely of Kate Peak volcanic rocks, both fresh and altered. A very few large boulders of granodiorite were found north of Thomas Creek.

The stream gravels in the Kate Peak Formation are exactly similar to the coarsest gravels in the Truckee Formation and in some places, as north of Thomas Creek, represent interfingering of the two formations.

INTRUSIONS

The intrusive rocks of the Kate Peak Formation are much like the lavas (fig. 5). They have a range of composition (analyses 36 to 39, table 1), texture, and appearance nearly identical to that of the lavas, and in many cases they probably represent the conduits from which flows and pyroclastics were erupted.

The large intrusions south of Fuller Lake in the northwestern part of the quadrangle are composed mainly of biotite-hornblende andesites much like those in the Virginia City quadrangle (Coats, 1936), where typical representatives (analyses 36 to 38, table 1) with silica contents of 60–64 percent have the composition of rhyodacite. These intrusions have disrupted and displaced large blocks of granodiorite. The largest block, south of Puny Dip Canyon, is nearly half a mile long and has been lifted about a thousand feet above the main body of granitic rock to the north. Along the southeast side of this block are excellent exposures of the chilled glassy border of the intrusion, with small inclusions of granodiorite. Small inclusions of andesite are also characteristic of the intrusions, as in the large intrusions in the Virginia City quadrangle.

721-882 O - 64 - 3

FIGURE 5.—Intrusive andesite of the Kate Peak Formation from quarry southeast of Virginia City, showing conspicuous phenocrysts of plagioclase and biotite. An inclusion of finer grained andesite is shown above the hammer head.

In a few places the intrusive rock grades into breccia similar to that in the extrusive tuff-breccia; this intrusive breccia, which is similar to that of intrusive dikes described by Durrell (1944), is interpreted as part of the eruptive feeder. Some of the intrusive breccia is monolithologic and thus resembles flow breccia, but some is composed of heterogeneous fragments of andesite, apparently derived partly from the walls of the conduit. The elongate intrusion a mile south of Fuller Lake, which forms a cone-shaped peak, is composed partly of breccia and partly of intrusive biotite-hornblende andesite containing inclusions of granodiorite. Hydrothermally altered tuff-breccia found at depth in GS–1 drill hole in the Low Terrace of Steamboat Springs is interpreted as intrusive (Sigvaldason and White, 1961) and was probably a feeder for extrusive tuff-breccias. Granodiorite overlies and also underlies this mass of tuff-breccia. Another example of what is probably intrusive tuff-breccia is shown in figure 6.

The large intrusion west of Hunter Creek and the very small one directly west of the large intrusion are composed of a biotite-bearing vitrophyre. Both intrusions are remarkable for enormous amounts of included granodiorite, and the small intrusion also contains blocks of metamorphic rocks. In the larger intrusion, the granitic rock, which in places makes up more than half the total volume, occurs both as large blocks several feet in diameter and as small masses down to the size of individual crystals of feldspar and quartz; the latter are abundant. The contacts of these intrusions are nowhere well exposed, but general relations indicate that the bodies are probably shallow intrusions that were connected with vents; near the

FIGURE 6.—Tuff-breccia of the Kate Peak Formation showing probable intrusive relation to granodiorite along abandoned railroad cut in canyon of Steamboat Creek below Little Washoe Lake. Boulder behind hammer head is granodiorite.

intrusions, some of the extrusive pyroclastic rocks of the Kate Peak Formation contain blocks of the vitrophyre and blocks of granodiorite. The composition of the vitrophyre is close to that of the vitrophyre member of the Kate Peak (compare analyses 40 to 42, table 1).

The effect of the large Kate Peak intrusions on structural deformation is discussed in the section on structure.

AGE AND REGIONAL RELATIONS OF THE KATE PEAK FORMATION

The upper part of the Kate Peak Formation intertongues with the Truckee Formation, which in the Virginia City quadrangle contains fossil leaves and diatoms indicating an age near the Miocene-Pliocene boundary (Calkins, 1944, p. 23), and which in the Reno quadrangle immediately north of the Mount Rose quadrangle contains fossil leaves and vertebrate fossils indicating a middle Pliocene Age (Axelrod, 1958). Intertonguing of the Kate Peak and Truckee Formations is well shown at the following localities: (1) The north end of Chalk Hills in the Virginia City quadrangle, where Truckee beds strike into and pass beneath Kate Peak flows, (2) along the east side of the Carson Range north of Thomas Creek, and (3) immediately north of the Mount Rose quadrangle in the bluffs along the south side of the Truckee River west of Hunter Creek, where thick pyroclastic beds of the Kate Peak Formation are interstratified with the normal Truckee Formation. As the fossils date only the upper part of the Kate Peak, the formation probably has an age span from Miocene to Pliocene.

The Kate Peak Formation probably interfingers with the underlying Alta Formation in the Virginia City quadrangle (see Thompson, 1956, p. 52, and the preceding discussion of Alta Formation), but the exact relations are obscure and may be understood only after reliable radioactive dating of the volcanics. Nowhere was clear evidence found for any long time break between deposition of the Alta and Kate Peak Formations. The Alta Formation, which clearly intertongues with the underlying Hartford Hill Rhyolite Tuff (Thompson, 1956, p. 50), contains leaves in the Sutro Member, which were judged by R. W. Brown to be middle or upper Miocene (Calkins, 1944, p. 15) and by Axelrod (1949, p. 1935), on the basis of larger collections, to be Oligocene. Unless the Oligocene age is in error, the Hartford Hill Rhyolite Tuff, the Alta Formation, and the Kate Peak Formation together were erupted during a very long period of sporadic volcanism.

Volcanic rocks similar to those of the Kate Peak Formation have been erupted from numerous vents over an enormous area in western Nevada and in the Sierra Nevada. The distribution of the Mehrten Formation, which is similar in age and composition to the Kate Peak Formation, is given by Curtis (1954, p. 455). The Mehrten is disconformably underlain by the rhyolitic Valley Springs Formation, which is of about the same age and belongs to the same cycle of volcanism as the Mehrten (Curtis, 1954, p. 454). These relations are similar enough to those of the Kate Peak Formation and Hartford Hill Rhyolite Tuff in the Virginia Range that a correlation of the Hartford Hill and the Valley Springs Formation might be suggested. But this correlation is unlikely if the fossil leaves in the Alta Formation, previously discussed, are Oligocene as interpreted by Axelrod. Independent means of determining the ages of the volcanic rocks are urgently needed.

Curtis (1954) explains voluminous tuff-breccias like those of the Kate Peak Formation by autobrecciation of rising viscous magma, aided by expansion of gas near the tops of conduits, and also by brecciation in viscous lava flows. This interpretation is supported by the large amount of breccia in intrusions and also by flow breccias in the area of the present study. Curtis follows earlier workers (Anderson, 1933) in regarding the crudely bedded tuff-breccias as a product of volcanic mudflows or lahars (a lahar is defined by Curtis on page 458 as "a volcanic breccia with a matrix of tuffaceous aspect which came to rest as a single unit and was originally mobilized by addition of water"), but the inference that mudflows are essential to the formation of all the tuff-breccias is probably incorrect. The definition would imply that temperatures were below 100°C to permit the presence of water rather than steam. There is little critical evidence to justify such an assumption. Hot and even incandescent flows of tuff-breccia containing steam are an alternative possibility. The numerous vents and the thick deposits near them show that transportation for long distances on low gradients is not necessarily required. Moreover, the potential and kinetic energy of breccia that is rapidly poured out on steep slopes is very large, and in such a mass semielastic collisions between the fragments may "liquify" or fluidize even dry rubble and permit it to continue moving on gentler slopes. Bragg (1948, p. 15–17) demonstrated that sand can be easily liquified by vibration. The larger a flow of tuff-breccia becomes under given conditions, the greater its content of energy relative to frictional losses of energy at its boundaries, for the volume increases more rapidly than the surface area. The special requirement that liquid water need always be abundantly available can thus be avoided, but the fact that water was involved in the deposition of part of the tuff-breccia is not in doubt.

The problems of distinguishing the very hot deposits of tuff-breccia from the moderately hot deposits and the tuff-breccias of true mudflow temperature (less than 100°) have not been solved. Paleomagnetism is a hopeful approach in distinguishing glowing avalanche deposits, for incandescence requires temperatures above the Curie point, and magnetic orientation should be consistent rather than random in the fragments of a single deposit; care must be taken to distinguish effects of hydrothermal alteration from initial orientation.

We are not aware of a reliable paleoindicator for the boiling point of water. Evidence for fumarolic activity from heat supplied by the deposit would indicate initial temperatures above 100°C (or a little less, depending on elevation), but this must be distinguished from the more commonly preserved evidence for hydrothermal activity superposed on and unrelated to the specific flow units. Charring of wood fragments is evidence for initial high temperatures, perhaps lower than the Curie point in some instances. Wood that is silicified was probably never exposed to temperatures of destructive distillation but is not a precise temperature indicator.

TRUCKEE FORMATION

King (1878, p. 412–423) was the first to use the name Truckee for the sedimentary rocks along the canyon of the Truckee River at the north end of the Carson Range. The name as now generally used refers to Miocene and Pliocene deposits of lakes and streams in northwestern Nevada. The deposits were never continuous over long distances but were laid down in numerous basins. Their composition and texture changes greatly even within one basin. These properties depended largely on the contemporary topography (structure) and, as the bulk of the clastic debris is of volcanic origin, on accidents of local volcanism. Axelrod (1958) used the name Coal Valley Formation for the same sedimentary rocks at the north end of the Carson Range and regards the Coal Valley as intertonguing with the Truckee Formation farther east. The distinction is based largely on an abundance of andesitic detritus in the Coal Valley Formation and basaltic debris in the (restricted) Truckee Formation of Axelrod, but as noted above these differences are not regular or consistent.

In the basin at the north end of the Carson Range between Reno and Verdi the following constituents of the Truckee Formation are each locally predominant: diatomite, debris from rhyolite, from granodiorite, from basalt, and from andesite of the Kate Peak Formation; but the andesitic detritus is by far the most abundant. The textural variation is from shale to conglomerate; tongues of tuff-breccia of the Kate Peak Formation are also included.

Near the northwest corner of the Mount Rose quadrangle and at several localities north of the quadrangle flows of basalt are intercalated with sandstone and shale and are mapped as a separate member; the flows have an irregular distribution, indicating that they probably flowed down valleys. One series of flows is splendidly exposed in cliffs along the Truckee River and Route 40, a mile east of Verdi. A flow three-fourths of a mile west of Hunter Creek and one-tenth of a mile north of the quadrangle is marked by charcoal and baked sandstone at its base and by sandy conglomerate with abundant basaltic pebbles in the overlying beds of the Truckee Formation. The basalt of the Truckee Formation is similar to the basaltic andesite of the Lousetown Formation but contains less silica (compare analyses 44 and 45 with analyses 48 to 59, table 1) and weathers to a more granular appearance, generally without the platy parting that is so commonly developed in the Lousetown flows. Moreover, the Lousetown Formation is much less deformed than the basalt of the Truckee Formation.

The lower contact of the Truckee Formation is well displayed on the southwest side of Alum Creek near the crossing of the Steamboat ditch and also 1½ miles west of Hunter Creek. At both localities the underlying rock is a flow of the Kate Peak Formation and the dips are roughly conformable. In Alum Creek the flow grades upward through coarse partly weathered rubble into conglomerate and sandstone of the Truckee. West of Hunter Creek the Truckee consists at its base of well rounded pebbles and cobbles of Kate Peak rocks. Near the northwestern corner of the quadrangle, the basalt member, underlain by a few feet of gravel containing pebbles of Truckee sedmimentary rocks, lies at the base of the formation.

The thickness of the Truckee Formation may amount to several thousand feet. Anderson (1909) made a rough measurement of 2,100 feet, but no accurate measurement is possible because of innumerable small faults and abrupt facies changes. Gravity measurements indicate that the thickness is much greater to the east of Hunter Creek than to the west and is probably about 1,800 feet (Thompson and Sandberg, 1958, p. 1276). An exploratory drill hole southwest of Reno is reported to have penetrated 1,890 feet of sedimentary rocks (Anderson, 1909, p. 487) and all or most of these are presumed to be Truckee.

The abundance of diatomite and of reedy plant remains throughout the formation indicate that the sediments were deposited in marshy flood plains and shallow lakes. The coarser sandstone and conglomerate are channeled and crossbedded, and undoubtedly they were deposited by streams between the swampy lakes and the hills or low mountains that were shedding the coarse sediments.

The volcanic eruptions that produced the Kate Peak Formation were partly contemporaneous with deposition of the Truckee Formation. As previously mentioned, thick pyroclastics are interbedded with Truckee shales and sandstones along the Truckee River west of Hunter Creek, and interfingering is also demonstrable in the Virginia City quadrangle. The exact degree to which the Truckee Formation is a sedimentary equivalent of the volcanic Kate Peak Formation is unknown, but partial contemporaneity is certain. Volcanic debris was washing into the basins between volcanic hills and low mountains while volcanism continued in the hills. Uplift of the hills and deformation along the margins of the basins were accompanied by continued deposition in the central parts of the basins. The highest altitudes at which Truckee Formation was found were 6,500 feet near Fuller Lake and 8,700 feet in Galena Creek. The latter occurrence is very small and probably represents a lens of fine-grained Truckee beds in a local basin that may have been formed by lava damming and not by structural deformation. It is unlikely that the Truckee Formation ever extended continuously over what are now mountain ranges, but it certainly once extended in many places far up the flanks of the present mountains and has since been removed by erosion.

The age of the Truckee Formation as indicated by leaves from a railroad cut a mile north of the quadrangle is Middle Pliocene (Axelrod, 1958).

TERTIARY AND QUATERNARY SYSTEMS

LOUSETOWN FORMATION

The Lousetown Formation, named by Thayer (1937, p. 1648) in the Virginia City quadrangle, is composed mainly of lava flows but includes also a few small intrusive plugs or necks and one cinder cone in the Mount Rose quadrangle. The flows, which thinly covered broad erosion surfaces and flowed around residual hills in the Carson Range, extend to the lower slopes at the north end of the range and also in eastern Steamboat Hills.

A few feet of stream gravel, composed mainly of pebbles of andesite and metamorphic rocks, underlie the lava in many places; the gravel is generally not a mappable unit in the scale of the quadrangle maps and is included in the Lousetown Formation; where it is most extensive, the gravel has been mapped separately as a member of the Lousetown Formation. In the detailed mapping of the Steamboat Springs thermal area of following chapters of this professional paper, the stream deposits are a separate mappable formation.

The fresh rock of the lava flows is a gray basalt or basaltic andesite; the weathered rock is darker. Sparse phenocrysts of plagioclase, pyroxene, and olivine rarely more than a millimeter in diameter are contained in an aphanitic groundmass. Olivine crystals, which are a common but not universal constituent, are rounded and altered, indicating partial resorbtion. A prominent platy parting, developed parallel to the flow banding and to the alined feldspar microlites, gives the outcrops of the Lousetown Formation a distinctive appearance in most places (fig. 7). However, the flows of Steamboat Hills lack the platy parting, contain prominent olivine and plagioclase phenocrysts to the exclusion of pyroxene and also contain a few partly resorbed quartz grains, probably as foreign inclusions.

In chemical composition, the Lousetown flows (analyses 48 to 59, table 1) are generally close to the boundary between basalt and andesite (analyses 48 and 55 are especially close to basalt). They contain a little quartz in the norm and the normative plagioclase ranges from labradorite to andesine. Some of the flows, including those in Steamboat Hills, are high in K_2O and TiO_2 (analysis 52, table 1) and show chemical affinities to the McClellan Peak olivine basalt of the Virginia City quadrangle (Thompson, 1956, p. 59–60; analyses 65 to 68, table 1); the only femic mineral in the phenocrysts is olivine.

FIGURE 7.—Platy parting in a flow of the Lousetown Formation west of Hunter Lake in the Carson Range. Parting is unusually well developed, probably because of active frost weathering.

The flows issued from several vents, of which the best marked are one at the high point of the flows in Steamboat Hills and one at the dissected cinder cone west of Davis Meadow. The high point 1½ miles south and a little west of Hunter Lake is marked by an unusual amount of scoria and may be the site of another vent; the same is true of an area about half a mile southwest of Big Meadows. The intrusion south and a little east of Incline Lake or the one at Rose Knob Peak west of the lake may have fed the flow capping a ridge south of the lake. The flows in the Virginia City quadrangle were also fed from many vents.

The Lousetown Formation is not easily distinguished from the basalt member of the Truckee Formation. In the Virginia City quadrangle remnants of basalt flows east and northeast of Chalk Hills were mapped as Lousetown but now seem more likely to be correlative with the basalt member of the Truckee Formation.

AGE

The Lousetown Formation lies on a post-mature erosion surface cut on the Truckee and pre-Truckee formations. The remnants of flows at lowest altitude in the northwestern part of the quadrangle rest on Truckee beds with an angular unconformity of about 35°. As the Truckee Formation is Pliocene, the Lousetown Formation must be Pliocene or younger. On the other hand it is older than late Pleistocene as evidenced by the degree of erosional dissection. West of Fuller Lake the Truckee River flows 1,000 feet below the lowest remnants of the Lousetown Formation, although the flows may once have extended to lower altitudes in the canyon. In the northwestern part of the quadrangle the low remnants of flows lie on the highest pediment cut on the Truckee, and this surface can be projected through several erosion remnants to the highest Truckee River gravels, which form a thin veneer on Chalk Bluff in the Reno quadrangle. These high river gravels contain a few boulders of basalt, which may be either from earlier Lousetown flows or from the basalt member of the Truckee Formation. The Lousetown flows are thus younger than the highest pediment and are either a little younger or a little older than the highest Truckee River gravels. The river gravels are nearly 400 feet above the present river and suggest that downcutting (or at least re-excavation of an aggraded valley) has amounted to to about 400 feet since the Lousetown flows spread onto the highest pediment. The canyon of Thomas Creek has been cut 200 feet below the lowest remnant of the Lousetown flow that came down that canyon. This erosion is greater than the dissection undergone by the oldest glacial deposits (Sherwin or older) that have been recognized in the area. The evidence suggests that the Lousetown Formation is early Pleistocene but might be partly or wholly Pliocene. If the interpretation of pediments and glaciofluvial river terraces, as discussed in the section on pediments, is correct, the Lousetown flows on the northern flank of the Carson Range are contemporaneous with or later than the McGee stage of Sierran glaciation (Blackwelder, 1931) but older than the Sherwin glaciation. The flows in Steamboat Hills are also pre-Sherwin but may be a little younger than Lousetown flows elsewhere, as noted below in the discussion of the gravel member of the formation.

New evidence of the age span of the Lousetown Formation comes from measurements of the direction of remanent magnetism in the flows. Preliminary determinations by Richard Doell of the U.S. Geological Survey indicate that the Lousetown Formation at the following localities is reversely magnetized (R1) and is therefore older than the present period of normal orientation of the earth's magnetic field (N1), which is thought to have begun in middle or early Pleistocene: (1) Steamboat Hills (same location as analysis 52, table 1), (2) Clark Mountain in the Virginia City quadrangle (analysis 51), (3) southeast of Mound House in the Dayton quadrangle (analysis 54), (4) southwest of Boca in the Truckee quadrangle (analysis 59). In contrast the flow remnant lying on the highest pediment in the northwestern part of the Mount Rose quadrangle (SE¼ sec. 21), which appears to be older because of its position, as described previously, is normally magnetized (N2?) and may therefore be early Pleistocene or late Pliocene. These inferences are supported by the fact that the Mustang Andesite and McClellan Peak Olivine Basalt, which directly overlie the Lousetown Formation at Clark Mountain in the Virginia City quadrangle, are both normally magnetized in accordance with the present magnetic field (N1).

INTRUSIONS

Shallow intrusions of basaltic andesite similar to that of the Lousetown flows occur in the southern part of the quadrangle: southwest of Rifle Peak; at Rose Knob Peak (analysis 57, table 1); south of Incline Lake; near Incline Creek; and in Little Valley (pl. 1). The rock, which is vesicular in places, generally displays a well-developed columnar jointing. A remarkable feature of the intrusions that cut granodiorite is local fusion of the granitic rock on the contacts or in inclusions to black glass resembling obsidian. All gradations from partial to complete fusion may be observed.

GRAVEL MEMBER

A few feet of gravel, generally a pediment gravel, underlies the lava flows of the Lousetown Formation at many places, but the gravel is separately mapped as a member only where its outcrop is widest. Well-rounded pebbles of fresh and altered andesite from the Kate Peak Formation are commonly the most abundant constituent and this makes the gravel difficult to distinguish from gravel lenses in the Kate Peak or Truckee Formations where exposures are poor. Southeast of Fuller Lake the gravels contain pebbles of metamorphic rock, for which no likely source is now exposed in the vicinity; presumably these pebbles came from the south or west and may have originated before the Truckee River cut through the Carson Range at its present location. In Steamboat Hills the gravels, which represent the capping of a well-marked buried pediment, contain granodiorite, andesite, metamorphic rock, chalcedonic hot-spring sinter, rhyolite pumice, and relatively abundant fragments of silicified wood.

In the northwestern part of the quadrangle, the gravel capping of the highest pediment cut on the Truckee Formation is correlated with the gravel member of the Lousetown Formation because remnants of the flows extend onto the pediment, as previously mentioned. The gravels of the lower younger pediments include fragments derived from the Lousetown. The highest pediment is tentatively correlated with the pediment in Steamboat Hills on which the Lousetown flows lie (see section on pediments), but a complex later history of erosion, deposition, and structural deformation make an exact correlation uncertain.

Other areas of gravel that may be correlative with the gravel member of the Lousetown Formation include the small patch lying on the Kate Peak-Truckee contact along Hunter Creek and perhaps the patch along a fault 4½ miles due east of Hunter Lake, although the latter patch is mapped as a younger gravel. Still other fairly extensive, though unmapped, areas of gravel are (1) in the Carson Range a mile southwest of Hunter Lake, (2) on the east side of the Lousetown flows 1½ miles northeast of Hunter Lake, and (3) on both sides of Thomas Creek near the lowest remnant of the Lousetown there.

Drill cores and cuttings from the thermal area at Steamboat Springs include relatively fine clastic sediments that White believes were deposited prior to the pediment gravels. These sediments occur along the eastern margin of Steamboat Hills at depths as much as 500 feet below the pediment surface upon which the Lousetown flows were erupted. Fragments of all local rocks older than the Steamboat flows have been recognized, but fragments having the characteristic texture of the Steamboat basaltic andesite are absent in these deep sediments in spite of proximity to present outcrops of the basaltic andesite. In summary, deep valleys were carved locally and were then filled as much as 500 feet deep before pediments were cut in bedrock on the flanks of the older valleys; the thickness of the deposits here included as a member of the Lousetown Formation was locally at least 500 feet.

STEAMBOAT HILLS RHYOLITE

The type locality of the Steamboat Hills Rhyolite is the dome of pumiceous rhyolite in the western part of Steamboat Hills. Although this dome represents the only outcrop of the rhyolite in the Mount Rose quadrangle, several domes were extruded at about the same time in the Virginia City quadrangle along the western and eastern margins of the Virginia Range (see analyses 60 to 63, table 1).

The rock is a light-gray pumiceous rhyolite (fig. 8), weathering to light tan, containing sparse phenocrysts of sanidine, quartz, sodic plagioclase, and biotite. Some of the rock is perlite and a very little is obsidian.

The dome in Steamboat Hills, which is half a mile in diameter and about 300 feet high, is capped by the eroded remnants of a crater, from which only a small amount of pyroclastic pumice was ejected. Explosive eruptions also preceded the extrusion of the dome and created a depression, now filled by the rhyolite but partly excavated by erosion on the south side. Ejected material includes pumice fragments and older rocks deposited on surrounding slopes as much as half a mile from the dome, particularly to the northeast, west, and south. The pyroclastic deposit is generally thin and has not been distinguished as a separate map unit on plate 1.

FIGURE 8.—Pumiceous Steamboat Hills Rhyolite, partly brecciated and welded, in a quarry near Sutro Springs on the east side of the Virginia Range.

Pumice fragments thought to be from the earliest eruptions are locally abundant in the gravels beneath the Lousetown flows in Steamboat Hills, and pumice lapilli from the last explosions are scattered very sparsely on the surface of the same flows. The andesitic and rhyolitic eruptions in Steamboat Hills were therefore contemporaneous or nearly so. As the andesitic eruptions in Steamboat Hills were probably among the latest of the Lousetown eruptions, the contemporaneous Steamboat Hills Rhyolite is certainly Pleistocene or late Pliocene. A Pleistocene age, probably pre-Sherwin, is supported by the following additional evidence: (1) Slight erosion, indicated by preservation of the summit crater, and (2) absence of pumice fragments from the surface of coarse pre-Lake Lahontan boulder gravels immediately west of the Steamboat Hills dome, contrasted with their presence on bedrock immediately east of the gravels; the rhyolite is therefore older than the gravels, which are believed to correlate with the Sherwin Glaciation in the Sierras, as discussed in a following section.

HOT-SPRING DEPOSITS

Hot-spring deposits at Steamboat Springs near the east end of Steamboat Hills consist mainly of siliceous sinter with a minor amount of travertine. The sinter terraces of active or recently active springs are composed of porous white opal near the surface. Dense chalcedonic sinter is abundant in the old terraces and beneath porous opaline sinter in some of the younger spring deposits. The mechanism of precipitation of opaline sinter and its reconstitution to chalcedonic sinter has been studied by White, Brannock, and Murata (1956) and will be considered more extensively in later reports.

The oldest spring deposits antedate the basaltic flows of the Lousetown Formation in Steamboat Hills, as demonstrated by pebbles of chalcedonic sinter in the gravel member at the base of the Lousetown. In addition, small remnants of once more extensive chalcedonic and opaline sinter occur on the flank of Steamboat Hills 1 mile southwest of the Steamboat post office; the deposits are at an altitude 400 feet above Steamboat Valley and lie on an extension of the pediment surface upon which the Steamboat flows of the Lousetown Formation were erupted. The exact age relations of the flows and sinter cannot be determined but are believed to be approximately the same. Sinters of several intermediate ages are known, and the youngest are still being deposited. White has mapped the complex relations of the long succession of spring deposits and has found that the springs have probably been active during much and perhaps even most of the Quaternary.

Small deposits of hot-spring travertine occur on the southeast flank of Steamboat Hills about 100 feet above the floor of Pleasant Valley, and Thompson briefly described local travertine deposits near the north end of the Virginia City quadrangle (Thompson, 1956, p. 60). The spring waters that deposited the travertines were probably influenced by contact with carbonate rocks and were probably only moderate in temperature (White, 1957, p. 1652–1653). The ages of the spring-travertine deposits are not known but local topographic relations are consistent with ages that are intermediate between the oldest and youngest sinter deposits.

GLACIAL DEPOSITS

All the main valleys that head above 9,000 feet are glaciated. Glacial debris covers large areas in the valleys of Galena Creek and Third Creek, which originate near Mount Rose and Slide Mountain in the highest part of the quadrangle. The drainage basins of Gray Creek and its west fork, Bronco Creek, and the upper parts of Whites Creek and Thomas Creek are also glaciated. As terminal moraines are inconspicuous or lacking, there is little to indicate the lower limits to which ice tongues protruded down the rocky canyons, but the glacier in Galena Creek certainly reached an altitude below 7,000 feet. All the glaciers have disappeared, but patches of snow occasionally linger over the summer in the highest cirque on Mount Rose.

Moraines of several ages are recognizable because of their various degrees of decay and dissection. In the drainage area of Galena Creek and Third Creek, four subdivisions have been separately mapped. Two of these represent the Tioga and Tahoe Tills (Blackwelder, 1931); one is older than these tills, and one is younger.

OLDEST GLACIAL DEPOSITS

The oldest glacial deposits are characterized by deep decay, complete destruction of the original topographic form of moraines, and high position along the sides of the valleys. Where exposed in roadcuts they are brown and iron-stained to a depth of more than 20 feet; granodiorite boulders are thoroughly softened by decay and the outer inch or so of andesite boulders is also decayed. Some residual boulders of granodiorite and many of andesite survive on the surface, however, where chemical decay is less rapid and where the fresh cores of the largest boulders are concentrated by erosion of finer debris. These features are well displayed on the Mount Rose road half a mile east of Grass Lake. The general features suggest correlation with the Sherwin Till of Blackwelder (1931). The soil is about as well developed as the pre-Lahontan soil recognized by Morrison (1961a, 1961b); however, at the depth at which the soils described by Morrison are enriched in carbonates, the soils on the pre-Tahoe tills are free of carbonate and contain instead veinlets of silica or clay minerals.

TAHOE TILL

The large lateral moraines of the Tahoe Glaciation are an impressive sight along Galena and Third Creeks, where their height above the present valley bottoms is as much as 400 feet. Although the lateral moraines are well preserved, end moraines are completely lacking. The mapped deposits extend down to altitudes of 6,250 feet on Galena Creek. The upper 3–5 feet of till may be stained tan by iron oxides; in the upper 3 feet the outer few inches of granodiorite boulders are softened, but andesite boulders have only begun to decay on their surfaces; at depths greater than 10 feet granodiorite boulders are perfectly fresh. The deposits are far less weathered than the older till but distinctly more so than the fresh light-gray Tioga deposits.

The terraces 60 and 75 feet above the Truckee River at Verdi, just north of the quadrangle, are covered with river gravel showing a comparable degree of weathering and soil formation; these terraces were correlated with the Tahoe stage by Blackwelder (1931, p. 893). All of the pre-Lake Lahontan gravels are more deeply and thoroughly weathered and are much older.

TIOGA TILL

Nested within the bulkier Tahoe moraines are much smaller moraines that are correlated with the Tioga glacial stage of the Sierras (Blackwelder, 1931). Breached remnants of several low end moraines cross the valley of Galena Creek at altitudes as low as 7,200 feet on Galena Creek. Weathering of the debris is less than that of Tahoe Till; faint iron staining and slight decay usually extend no deeper than 3 feet and granite boulders are generally almost fresh up to the surface; but weathering varies with location and the cover of vegetation. The Tahoe and Tioga Tills are difficult to distinguish on the basis of weathering alone except where good cross sections are available in roadcuts.

YOUNGEST GLACIAL DEPOSITS

Some of the high cirques contain small moraines, untouched by weathering and erosion, that can be no older than a few tens to a few hundreds of years. Two fine examples, both at an altitude of 9,200 feet, are a mile west and 2 miles northwest of the summit on the Mount Rose road (fig. 9). The cirques contain many young trees, only 10–20 feet tall, that have grown since the moraines were formed. In contrast, trees growing on the slopes outside of these cirques are mature and several times as large.

FIGURE 9.—Small moraine a mile west of the summit on the Mount Rose road, representing the youngest glacial deposits. Young trees in the hollow of the moraine contrast with larger trees up the slope. Mount Rose in the background.

These young moraines probably record the maximum glacial expansion that occurred about 1750 in many parts of the world, an event often called the "little ice age" (Matthes, 1942, p. 204–215; Lawrence, 1958, p. 90).

CORRELATION WITH LAKE LAHONTAN

The oldest glacial deposits recognized in the area are older than the deposits of the Tahoe stage of Blackwelder (1931) and are correlative with or include deposits of the next older stage, the Sherwin of normal Sierran terminology.

A very mature soil developed on Sherwin Till has been correlated by Morrison (1952, 1961b) with a soil in the Lahontan Basin that is developed on deposits older than Lake Lahontan. The soil and the Sherwin Till are therefore considered to be earlier than Lake Lahontan, and the till is middle Pleistocene in age. No attempt is made here to correlate glacial stages of the Sierras with the midcontinent; the problems have been reviewed recently by Richmond (1961).

Morrison (1961b) has correlated the Tahoe stage of glaciation of the Sierras with older upper Pleistocene deposits of Lake Lahontan and the Tioga stage with younger upper Pleistocene deposits of the lake; he believes that each advance of glaciers in the Sierras correlated closely with a rise in lake level. Morrison's evidence consists principally of soils of differing maturity; each soil is believed to have developed in a short period of time when climate was particularly favorable. Morrison's conclusions seem to be valid to the extent that they can be tested in the Mount Rose and Virginia City quadrangles.

LANDSLIDES

Large volumes of rock slid down the canyons in two kinds of situations, the first where erosion undermined tuff-breccia of the Kate Peak Formation or sedimentary rocks of the Truckee Formation, generally beneath sheets of basaltic lava, and the second where crushed and jointed granodiorite was tectonically oversteepened on the slopes of Slide Mountain.

LANDSLIDES RELATED TO KATE PEAK AND TRUCKEE FORMATIONS

The largest slides of the first kind are in the precipitous side canyons leading into the main canyon of the Truckee River. Parts of Bronco Creek and Puny Dip and Deep Canyons are completely choked with large and small blocks of basaltic andesite of the Lousetown Formation, lying near the angle of repose on talus-like slopes. Much of the material mapped as landslide can properly be described as talus, but it continues to move because the fine material beneath is being washed away; some is normal landslide and exhibits hummocky topography and small closed depressions. Fuller Lake and some of the ponds below it owe their origin to sliding in the Truckee Formation; the slide is comparatively small and has not been shown on the map.

LANDSLIDES FROM SLIDE MOUNTAIN

On the southeast side of Slide Mountain an enormous scar (fig. 10), which is a conspicuous landmark that gives the mountain its name, is the source of multiple slides extending 3 miles eastward into Washoe Valley. Fresh granitic rock exposed in the scar is dazzling white. The bulky debris tongue, which is composed wholly of granitic rubble, contains Upper and Lower Price Lakes at the base of the main scar and Rock Lake in a secondary landslide cirque at the north end of Little Valley; smaller ponds and depressions lie on the hummocky surface of the lower end of the debris tongue and are well shown west of Route 395 where the highway cuts through one lobe of the debris tongue.

721-882 O - 64 - 4

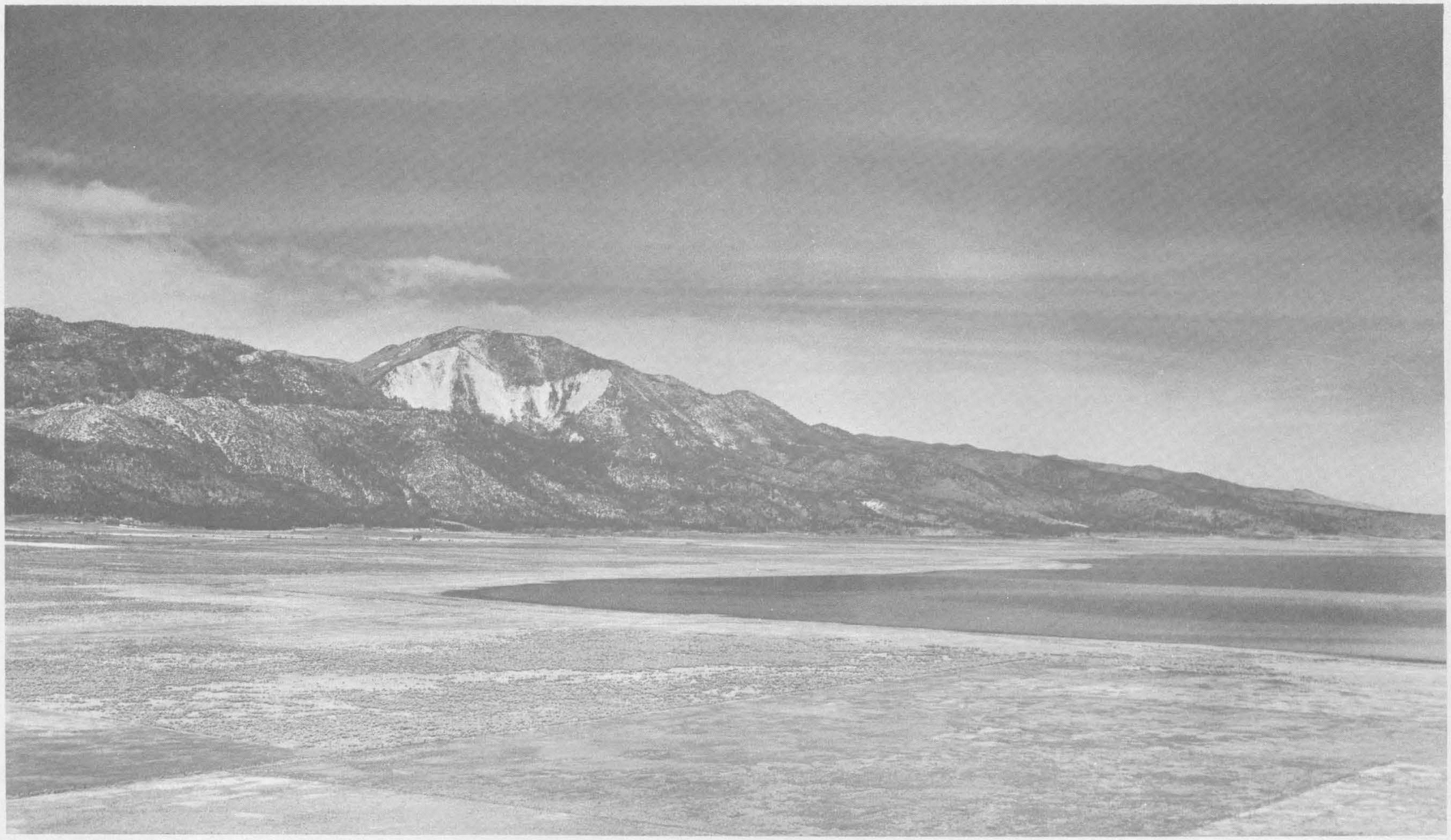

FIGURE 10.—Landslide scar on Slide Mountain; view northwest across Washoe Lake. The landslides, which blend with the granitic bedrock in the picture, moved downward and to the right, extending onto the floor of Washoe Valley. Little Valley, about halfway up the mountain, extends to the left from the scar.

The magnitude of the slides is best described quantitatively. The scar on Slide Mountain is sharply bounded, and by conservatively reconstructing the contour across the face of the scar, one can estimate that the volume of granodiorite that has slid away is 125 million cu yd (0.025 cubic miles). The debris tongues are exposed over an area of 1–2 sq mi but were once more extensive, for the lower ends are partly concealed beneath younger alluvium in Washoe Valley. If we assume that the area is 2 sq mi and that the material expanded 20 percent as it broke up, an average thickness of 75 feet of debris would account for the volume estimated at the source. That this thickness is not unreasonable is shown by the least eroded lobes, the fronts of which stand nearly 100 feet high. Conservative estimates of the volume of the scar and of the debris tongue are thus in agreement and indicate that this composite slide is one of the largest on record.

The face of the main scar descends from an altitude of 9,400 feet to 7,200 feet at an angle of 42°. The debris tongues slope down to 5,040 feet at an average angle of 8° or 700 feet per mile. The total difference in altitude is about 4,400 feet.

AGE AND ORIGIN OF SLIDES

Intermittent sliding over a long range of time is clearly demonstrated by differences in weathering and erosion of the debris and by alluvial burial of the lower parts of the oldest slides. Two main subdivisions of the debris tongue have been recognized.

The younger subdivision is composed of angular granitic rubble ranging in size from sand to abundant blocks 1–3 feet across and a few blocks more than 10 feet across. The texture is remarkably similar to that of tuff-breccia in the Kate Peak Formation. The upper part commonly has open spaces between the blocks, probably owing to washing away of the finer material; stream cuts show more fine material below the surface. The front of the younger slides stands as steep as 30°–35°, which is about the angle of repose, and rises abruptly to a height of nearly 100 feet. The slight decay and faint staining by iron oxide differ from place to place but in general range from that of the Tahoe Till to that of the Tioga Till. The younger subdivision is composed of more than one individual slide, as shown by interstratified stream gravels exposed in the inner trench of Ophir Creek. The present scar is covered with loose unstable material indicative of some recent sliding.

The older subdivision of the debris tongue is deeply weathered, dissected, and partly buried. In roadcuts granodiorite blocks are decayed and thoroughly softened to depths greater than 20 feet, but residual blocks are still plentiful on the dry upper surface and a few resistant blocks, such as those of aplite, survive at depth. The debris is brown and clayey, in contrast to the light-gray material of the younger slide material. In spite of the deep decay and considerable erosion, undrained depressions still persist on parts of the upper surface. The most easterly remnants, which are nearly engulfed by alluvium in Washoe Valley, display only the residual slightly weathered large blocks characteristic of the upper surface. These remnants thus appear less weathered than the debris exposed in roadcuts, but original flow fronts must have been at least 50 feet high and are now almost buried by alluviation; these relations indicate that the remnants are part of the older subdivision. The degree of decay characteristic of the older subdivision of the debris tongue is as great as, or greater than, that of the oldest Sherwin glacial deposits in the Carson Range and of pre-Lake Lahontan deposits of the Carson Desert (Morrison, 1961a, 1961b); on this evidence the age is pre-late Pleistocene, an interpretation that is supported also by the amount of dissection and alluviation.

Although the debris tongues have the superficial appearance of glacial moraines and have been identified as such, they clearly originated by landsliding. The altitude of the base of the scar (7,200 ft) is less than that of the lowest recognized glacial cirque (about 7,800 ft northeast of the crest of Slide Mountain) and the debris tongue also extends to a lower altitude than do any of the glacial moraines. More important, the only large gathering basin available for such a glacier would be in Tahoe Meadows at the headwaters of Ophir Creek, but the glacier originating in that basin flowed south toward Lake Tahoe. A complete lack of volcanic material in the Slide Mountain debris tongue proves conclusively that no glacier flowed down Ophir Creek from Tahoe Meadows, where volcanic rock is abundant. The present drainage of Tahoe Meadows into Ophir Creek is the result of recent capture, owing to the extremely steep gradient from Tahoe Meadows to Washoe Valley. Small glaciers occupied the high cirques on Slide Mountain but their deposits are very restricted.

The cause of the sliding can be found in a combination of factors, the most important of which are extensive tectonic crushing of the granodiorite, joints dipping southeast parallel to the slope (pl. 1), and over-steepened slopes ultimately related to tectonic movements. The major faults strike north-northeast in Little Valley and north on the face of Slide Mountain; the landslide scar is at the bend. Normal faults with downthrow on the western (mountain) side are an additional tectonic factor that probably resulted in unusual crushing. One large fault with this displacement extends from Galena Creek southward into the scar, converging toward the largest fault in Little Valley, which has downthrow in the opposite direction. The combination of multiple faults, large throw in both directions, a sharp change in strike, and very steep slopes seems adequate to explain unusual crushing and the formation of the voluminous slides. Any future movement on faults in the area could trigger new slides.

Once material was loosened and began to move on such steep slopes the available energy was so great that semielastic collisions between the blocks might make the slide fluid, like sand undergoing vibration (Bragg, 1948, pp. 15–17). That at least part of the debris was fluid enough to move rapidly seems to be demonstrated by the low gradient and backslopes around depressions at the lower end. Abundant water from rain and melting snow probably also helped to buoy and lubricate each slide.

PRE-LAKE LAHONTAN DEPOSITS

The older gravels, sands, and silts antedate the Tahoe and Tioga Tills and the upper Pleistocene deposits of Lake Lahontan. The largest expanses are stream gravels capping the extensive pediments in the northern part of the quadrangle and sands and gravels northwest of Steamboat Hills. In most of the latter area, very coarse boulder gravel covers the surface (fig. 11); the boulders range up to 12 feet or more in diameter and are commonly 3–5 feet. In a few roadcuts and gravel pits, the boulder-strewn surface is underlain by coarse bouldery deposits (fig. 12), but elsewhere the underlying deposits range up to cobbles in size, and large boulders are absent. Wells drilled on the boulder-strewn surface north of Steamboat Hills have demonstrated an absence of boulders at depth; relatively fine clastic sediments correlated with pre-Lake Lahontan deposits extend down at least 100 feet below the bouldery surface. Evidently the underlying material was laid down under very different conditions from the upper few feet of boulders.

FIGURE 11.—Boulder-strewn surface of pre-Lake Lahontan deposits immediately north of Steamboat Hills, looking west-northwest to Carson Range. Vegetation of foreground largely destroyed by recent brush fire.

FIGURE 12.—Pre-Lake Lahontan deposits and mature soil, roadcut half a mile south of mouth of Whites Canyon, Carson Range. Large granodiorite boulders, not evident in photograph, strewn on upper surface. Boulders of andesite are relatively fresh but those of granodiorite below original surface are decomposed; note pick imbedded in large boulder.

The deposits west of Washoe Lake and west of Little Washoe Lake are composed of sand and silt; they have been structurally uplifted and probably originated as deposits of a more extensive lake in Washoe Valley. A patch of gravel lying on granodiorite on the east side of Washoe Valley caps an erosion surface that slopes westward toward a gravel hill lying out in the valley. In the southwestern corner of the quadrangle, near the north shore of Lake Tahoe, are gravels containing boulders up to 6 feet in diameter; the degree of dissection and decay marks these also as pre-Lake Lahontan deposits.

The recognition and comparative dating of the pre-Lake Lahontan deposits are based largely on weathering and soils. Topographic position, structural deformation, amount of erosion, and stratigraphic relations are also important criteria. The soil is characteristically much deeper than on younger deposits in the same area (fig. 12); it is comparable to or even greater than that on the Sherwin Till. The horizon equivalent to the (B) horizon of Morrison (1961a, p. D111) is 2–4 feet thick where fully preserved and is reddish brown and clayey. Buried granodiorite boulders in this zone are thoroughly rotted, andesitic boulders less so, although boulders lying on the surface remain fresh and sound. The underlying light-colored horizon equivalent to Morrison's (Cca) horizon is from 4 to at least

10 feet thick; typically it is characterized by veinlets of clay or silica with little or no carbonate.

This relatively mature soil has the essential characteristics of the pre-Lake Lahontan and equivalent pre-Tahoe soils recognized by Morrison (1961a, 1961b) in the Carson Desert and Sierra Nevada and marks the deposits as pre-Tahoe in age. An upper limit to the age range is more problematical. In the northern part of the quadrangle, the pediment gravels contain a scattering of fragments derived from the Lousetown basaltic andesite, and in general the deposits mapped as pre-Lake Lahontan are post-Lousetown. But in many places no source of Lousetown is present in the drainage system and the criterion of relative age is lacking. Thus some gravels as old as or older than the gravel member of the Lousetown formation may be included, as in the Virginia City quadrangle (Thompson, 1956, p. 61).

Relatively fine-textured clastic sediments mapped as pre-Lake Lahontan deposits lie near the exit of Steamboat Creek from Washoe Valley (pl. 1). The abundant diatoms have been studied by K. E. Lohman of the U.S. Geological Survey (USGS diatom loc. 4188), who reported (written communication, 1958) that:

> Collection 4188 yielded an assemblage of 39 species and varieties of very well preserved freshwater diatoms * * * Most of the diatoms are normally littoral forms, although the dominant species present, *Stephanodiscus astraea*, is a pelagic one. This suggests somewhat deep * * * water * * * With the single exception of *Cymbella* sp. *A*, which is probably a new species, and as far as is now known may be extinct, all of the diatoms in this assemblage are still represented in living assemblages elsewhere. Many have long geologic ranges going back to the Miocene, and the absence of short ranging forms makes it difficult to arrive at a reasonable age assignment. Thirty-six percent of the diatoms * * * also occur in the Provo Formation of Late Pleistocene age in Utah and 39 percent occur in the Hagerman Formation of Early Pleistocene age in Idaho.

Sand and fine gravel also correlated with pre-Lake Lahontan deposits but stripped of the typical surficial very coarse boulder gravels were found to contain rather abundant diatoms at a depth of 14–16 feet in the South Steamboat well, immediately south of the limit of hot-spring sinter at Steamboat Springs (pl. 2). According to Lohman (written communication, 1958):

> Collection 4182 yielded a diatom assemblage of 27 species and varieties, of which all except two, *Surirella utahensis* and *Surirella* sp. *A* are living elsewhere today. *Surirella utahensis* has a known geologic range of upper Pliocene or lower Pleistocene (Tulare Formation of California) to Late Pleistocene (Provo Formation of Utah) but has not been reported living. *Surirella* sp. *A* is a new species and quite probably an extinct one. Eighteen percent of the diatoms in this assemblage also occur in the Provo Formation * * * of Late Pleistocene age and 37 percent * * * in the Hagerman Formation of Early Pleistocene age * * * This assemblage is composed dominantly of shallow, freshwater forms, with a number of species which live today in saline lakes, plus four species characteristic of thermal springs * * *

Evidence from diatoms in deposits considered to be parts of the pre-Lake Lahontan deposits is not diagnostic of specific age but is consistent with the Sherwin age assigned by correlation with glacial deposits of the Sierra Nevada and soil development. Further information on the age relations of the gravels is given in the discussion of pediments.

YOUNGER LAKE AND STREAM DEPOSITS

The younger deposits are contemporaneous with or later than the Tahoe and Tioga Tills and the sediments of Lake Lahontan. They are distinguished from older sediments by immature and shallow soil; by topographic position, and, in many places, continuing sedimentation; by a nearly complete lack of measurable structural deformation; and by their stratigraphic overlap on pre-Lake Lahontan deposits. The soil, where best developed, consists of a foot or so of leached light-brown to black material, which is poorer in clay than the older soils; an underlying zone of lime or silica enrichment is often indistinct or even absent and always much thinner than that of the older soils. Where cobbles or pebbles are present in the leached zone, those of granodiorite may be softened in their outer 1 or 2 inches, but andesite cobbles and pebbles have only begun to decay on their surfaces.

The lake deposits are mainly sand and silt. Those in the northeastern corner of the quadrangle were laid down in recent temporary lakes and in what was probably an extension of Lake Lahontan. A temporary lake forms and rises to an altitude of about 4,400 feet when the Truckee River is in flood (fig. 13). This ponding

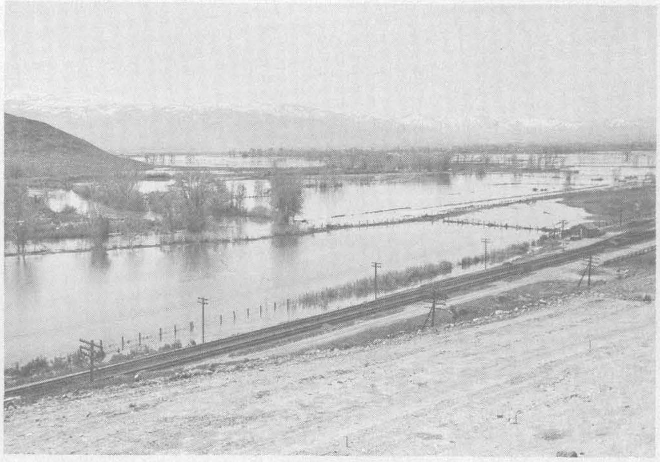

FIGURE 13.—View southwest of Truckee Meadows on May 11, 1952, flooded where the Truckee River enters the Virginia Range. River flows to the left in the foreground. Temporary lake extends south as far as Huffaker Hills.

results from the structurally maintained low gradient of the river where it enters the canyon through the Virginia Range.

During the high stage of Lake Lahontan a lake that is here called Truckee Meadows Lake was maintained at least for short periods of time in Truckee Meadows. A faint rocky beach and a thin deposit a fraction of an inch thick and considered to be tufa that was precipitated on the shore of a lake occur on the west side of a small hill of Kate Peak andesite that lies in the Vista quadrangle three-fourths of a mile north of the Virginia City quadrangle and 1¼ mile south of Route 40. The coating occurs up to an altitude of nearly 4,410 feet, which is about 20 feet above the surrounding floor of Truckee Meadows and 10 feet above the high level of driftwood found on this hill from the major flood of 1950.

According to Morrison (1961a), the highest altitude attained by Lake Lahontan in the Carson Desert area was 4,380 feet, which is a few feet above the bed of Truckee River immediately north of the tufa-bearing hill. Discharge through the Virginia Range must have been very sluggish at times of high stands of Lake Lahontan because of high discharge of the Truckee River and the low gradient. A water level as high as 4,410 feet and probably even higher during flood stages was necessary in Truckee Meadows Lake to provide the gradient necessary for Sierran runoff to discharge into the main body of Lake Lahontan. The existence of the tufalike deposit as high as 4,410 feet implies either enough evaporation in Truckee Meadows Lake for the pH to rise and calcium to be concentrated sufficiently for calcium carbonate to precipitate from waters that were initially relatively low in calcium, or else the deposit was formed by evaporation of wave splash immediately above lake level. In view of the large flow and low calcium content of the Truckee River and the small evaporation area of Truckee Meadows Lake (not more than 25 sq mi) formation of the tufalike deposit from splash appears more likely.

Lake sediments also surround and connect Washoe and Little Washoe Lakes, recording a formerly much larger lake in Washoe Valley. Beach sands on the east side of Washoe Lake have been blown by the wind into dunes, which are not separately mapped but are shown by the topographic contours. Dune sands derived in part from Washoe Valley also occur in pockets well up on the flank of the Virginia Range, and many rock outcrops have been abraded by blowing sand. The lake sediments near the southwestern corner of the quadrangle are beach sand and gravel of a higher stage of Lake Tahoe, which probably resulted from glacial damming of the outlet during the Tahoe or Tioga Glaciation. Other lake deposits represent the sediments of small glacial lakes and structural basins in the mountains.

The deposits from late Pleistocene and Recent streams range from boulder gravels deposited by mountain torrents to silt laid down by sluggish streams such as those in the lower part of Truckee Meadows. Boulders more than a few feet across are comparatively unusual in the younger stream deposits, a fact that may indicate some unusual combination of circumstances for the very coarse boulder gravels of the pre-Lake Lahontan deposits.

ALTERED ROCKS

The most drastic and spectacular alteration, shown on the maps of the Virginia City and Mount Rose quadrangles by a dotted overlay pattern, is a bleaching and iron staining that affects the more permeable rocks of the Alta and Kate Peak Formation in large areas and other rocks locally. Patches of metamorphic rocks and granodiorite east and southeast of Little Washoe Lake, for example, are bleached. The alteration is generally restricted to Kate Peak and older rocks, but near Steamboat Springs the younger rocks are also affected. In addition to the bleaching, alteration of a type generally known as propylitization is common in some areas; although not distinguished on the geologic map, it is described briefly and its origin discussed.

BLEACHING

Bleaching differs in intensity from a slight oxidation and partial breakdown of silicate minerals to clay, through complete conversion to clay and residual fine-grained silica, with or without iron oxide, to replacement of the original rock by porous or dense chalcedonic silica. The chemical changes similarly grade from slight leaching of magnesium and calcium with partial oxidation of iron, through nearly complete removal of everything but alumina, silica, and a little iron, to removal of all the main constituents but silica. Iron oxide and silica are especially subject to local redistribution, whereas magnesium, calcium, sodium, and potassium are progressively removed. Analyses 69, 70, and 71 (table 1) represent respectively a tabular upper zone, completely silicified, immediately below an old erosion surface; a middle zone in which silica and iron oxide have been enriched; and a lower zone in which titanium oxide has also been greatly enriched. The upper zone now consists almost entirely of chalcedony; material from the lower zone is soft and pulverulent and gives X-ray patterns of α-cristobalite, hematite, and anatase. These rocks were normal andesites of the Kate Peak Formation before they were altered. Analysis 72 (table 1) shows removal of some silica, along with magnesium, calcium, sodium, and some potassium, and residual enrichment of alumina as kaolinitic clay. Analysis 74 shows strong residual enrichment of silica as well as alumina. Pyrophyllite is, surprisingly, the dominant aluminum silicate; quartz is abundant, and kaolinite, anatase, and alunite are present. Analysis 73 represents an end product of silicification, with silica present as chalcedony.

The texture, density, and appearance of bleached rock are extremely variable, but most of the rock is light colored as implied by the name. Much is glaring white or light tan, streaked and stained by yellow, red, and brown iron oxide. The most thoroughly silicified parts are dull gray in places but elsewhere are rusty and varicolored. In the last stages of alteration all trace of the texture of the original rock is generally obliterated but may be well preserved in a few places where dense glassy opal or cristobalite has completely replaced the original minerals, as locally in the Steamboat Springs thermal area and southwest of Washington Hill in the Virginia City quadrangle.

Bleaching is a special type of weathering, resulting from attack by sulfuric acid formed where atmospheric oxygen reacts at and near the surface with disseminated pyrite or, in active thermal areas like Steamboat Springs, with rising hydrogen sulfide. Abundant pebbles of bleached andesite in the Truckee Formation show that the process began before the Truckee was deposited, and some of this bleaching may be the result of oxidation of H_2S that accompanied volcanic activity. Oxidation of pyrite disseminated in the rocks is still going on, as evidenced by the acidity of soils (pH as low as 3.5) and of runoff from bleached areas (pH as low as 2.4); present activity is best seen in young stream gullies, roadcuts or mine dumps.

The soil of the bleached areas favors the growth of yellow pines over nearly all other vegetation, and where the alteration is intense, yellow pines grow on bare yellow soil. Within the semiarid sagebrush and pinyon-juniper zones, local stands of tall yellow pines, surrounded by sagebrush, juniper, and pinyon on unaltered ground, are remarkable (fig. 14). The yellow pines in the Virginia Range grow nowhere except in isolated stands on bleached rock. In the sagebrush zone near Steamboat Springs yellow pines grow even in warm acid ground that is 99 percent hydrous and anhydrous silica. The distribution and causes of this botanical phenomenon were studied by Billings (1950), who concluded after considerable experimentation that the explanation lies in the inability of most plants except the yellow pine to grow in acid soils that are also very deficient in phosphorus, nitrogen, and exchangeable bases. The pH of the altered soil Billings used in his nutrition experiments was 4.32; the total replaceable bases were only 3.8 milliequivalents per 100 grams dry soil, in contrast to 30.4 milliequivalents in normal soil from unaltered andesite (Billings, 1950, p. 71–72).

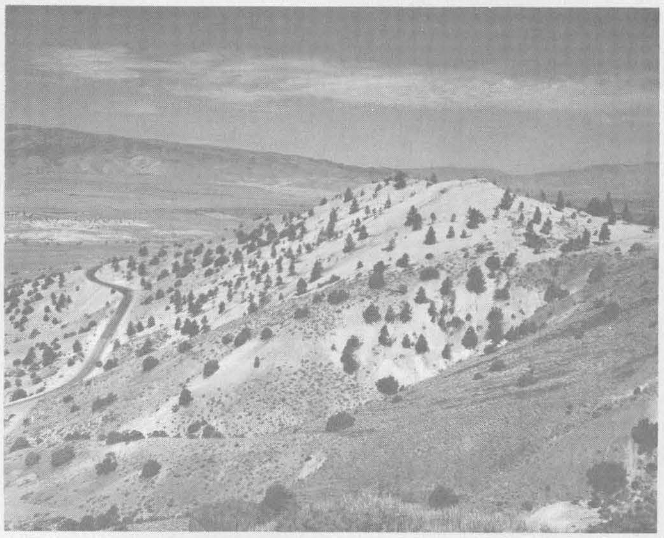

FIGURE 14.—Area along Geiger Grade in the Virginia City quadrangle, showing influence of bleached rock on vegetation. The most intensely bleached areas support yellow pine (pointed tops), less bleached and unbleached areas support pinyon pine (rounded tops) and sage. To the left in the distant valley, the altered area at Steamboat Springs is also dotted with yellow pine. View to the west.

Drill holes and mine workings in the Virginia City quadrangle show that bleached rock is generally confined to the zone of weathering and oxidation and grades downward into greenish-gray propylitized rock that is peppered with pyrite. Exceptions may occur where oxidizing groundwater descends to great depth in permeable zones or where oxidation took place before burial by younger rocks. Some specimens taken during the driving of the Sutro tunnel in the Virginia City quadrangle are bleached, but most of the rock is not, though pyrite is abundant. Pyritized rock on mine dumps crumbles and bleaches within a few tens of years. Gravity surveys also indicated that porous bleached rock is generally superficial (Thompson and Sandberg, 1958, p. 1277).

Introduction of sulfur to combine with part of the iron in ferromagnesian minerals is all that is required of the primary alteration that prepares a rock for bleaching. Even pyritic schist occasionally shows bleaching, as may be seen in upper Whites Creek. In the altered but unoxidized andesites, tiny pyrite crystals are in places arranged in the shape of original pyroxene or hornblende crystals; analyses suggest that iron generally has not been added, although it has certainly been redistributed locally and some may even have been removed. The bleached volcanic rocks commonly overlie propylitized rocks bearing pyrite or lie near propylitized rocks without pyrite.

PROPYLITIZATION

As stated by Calkins (1944, p. 20):

The term propylite was coined by von Richtofen for what he regarded as an independent species of volcanic rock, especially well exemplified in this district (the Comstock Lode). Becker (1882) showed that propylite was merely altered andesite, and the noun 'propylite' has consequently fallen into disuse, but the adjective derived from it is still applied to a certain sort of alteration that has never been very clearly defined. As the Comstock district is the type locality for propylitization, the term is probably applicable to changes that are well exemplified on the Ophir Grade and the adjacent flume (in Virginia City Quadrangle). The more evident effects of this alteration are the acquisition of greenish hues and a thorough cementation of porous rocks, such as tuffs and breccias, which makes them compact and hard. In detail the changes in the minerals are complex and irregular, but the commonest alteration products are chlorite, epidote, and calcite.

Coats (1940, p. 16) defined propylitization as a process of alteration of andesite that results in the formation of epidote comparable in size with feldspar phenocrysts, secondary albite replacing plagioclase, and chlorite, calcite, and epidote replacing mafic minerals. But Wilshire (1957, p. 244) preferred not to restrict the definition to alteration of andesite but to include alteration of any rock, unattended by strong deformation, which produces in andesites some combination of secondary epidote, albite, chlorite, and calcite. The latter definition is used here, because rhyolite and even granodiorite are included in altered areas in the Virginia City quadrangle.

Propylitic alteration persists up to the surface only where little or no pyrite was formed; where pyrite is present, the propylitized rock is always bleached at the surface. Propylitic alteration is not as abundant or generally as intense in the Mount Rose as in the Virginia City quadrangle, but good examples may be seen in and adjacent to the two large Kate Peak intrusions along Puny Dip Canyon. Patches of propylitized rock also occur near the bleached zones and elsewhere, with little evidence of a systematic pattern. In the Virginia Range propylitized rocks are more extensive than bleached rocks, but on a large scale the two types of alteration seem to be associated, as can be seen on the geologic map of the Virginia City quadrangle by comparing the distribution of the Alta Formation, much of which is propylitized at least weakly, with the distribution of bleached rock.

Details of the alteration are revealed in cores and cuttings from a drill hole 4 miles north of Virginia City (SW¼NE¼ sec. 6, T. 17 N., R. 21 E.) which have been studied by petrographic methods and to a preliminary extent by X-ray methods. The surface rocks, as mentioned previously, are bleached andesite composed mainly of pyrophyllite and quartz, with a little kaolinite, anatase, and alunite, and with 72 percent silica (analysis 74, table 1). Below the bleached zone, which was not sampled but which apparently extends to a depth of 50 or 75 feet, the altered rocks extending down to a depth of at least 385 feet (analyses 75 and 76, table 1) contain epistilbite, stilbite, chlorite, montmorillonite, illite, potassium-feldspar, secondary quartz, pyrite, and locally carbonate, but no epidote.

At a depth of 540 feet or somewhat less, epidote appears, along with chlorite, sodic plagioclase, carbonate, some stilbite and epistilbite, and pyrite. At 586 feet and 622 feet, pyrite and clay minerals are lacking (analyses 77 and 78, table 1), and abundant epidote occurs with chlorite, carbonate, actinolite, and minor stilbite and epistilbite. At 656 feet slightly altered andesite contains a little epidote, actinolite, and chlorite but no pyrite or zeolite (analysis 79, table 1).

These data show that propylitic alteration that is characterized by epidote overlaps a zeolite-clay mineral alteration; presence or absence of pyrite is incidental. Below the weathered zone no large changes in chemical composition generally accompany alteration, except for addition of water, sulfur, and CO_2 and some mobilization of magnesium, calcium, sodium, and potassium; several analyses also suggest movement of some silicon and aluminum (analysis 76, also 80 and 81 from Sutro tunnel, table 1). The alteration in the upper part is principally chloritic, argillic, and zeolitic; that at greater depth is characterized by abundant epidote and replacement of part of the original plagioclase by a more sodic feldspar. The latter type is evidently a higher grade alteration, which generally takes place at greater depth and at greater temperature and pressure than the zeolite-clay mineral alteration.

Regionally, propylitic alteration is most intense around certain centers, the most important of which is in the Comstock Lode district; an example of a smaller center is at the two large Kate Peak intrusions on the western border of the Mount Rose quadrangle. In the Virginia Range altered rocks as distant from the Comstock Lode as those near Washington Hill (Thompson, 1956, p. 63) generally have less epidote and are characterized by argillic and pyritic alteration and an abundance of opal or cristobalite instead of quartz or chalcedony. According to White, Brannock, and Murata (1956, p. 54), opal is generally restricted to lower-temperature environments (below about 140°C) and is converted to chalcedony with sufficient time and at higher temperatures.

Wilshire's study of gradations in intensity of propylitization in the Ebbetts Pass region of California indicates that chlorite and calcite appear without epidote in mildly altered zones (Wilshire, 1957, p. 250–251); his description of the minerals also indicates epidote is not always present (p. 246–247). Zeolites and quartz occur in veins in propylitized rocks but opal predominates as a vein mineral outside altered zones (1957, p. 250). Albitization of plagioclase is rare or absent (1957, p. 253). Evidently the Ebbetts Pass rocks exhibit similar gradations in rank or intensity of alteration, and possibly depth zoning not yet clearly recognized.

ORIGIN AND STRUCTURAL CONTROL

Most workers have concluded that propylitization is a hydrothermal alteration and have sought the source of the water in an underlying magma. Recent evidence indicates that hydrothermal solutions, as represented by hot-spring waters in areas of recent volcanism such as Yellowstone Park and Mount Lassen, contain no more than about 5 percent magmatic water (Craig and others, 1956, p. 35–36; White, 1957). The source of the heat must certainly be magmatic and the sulfur and CO_2 are probably magmatic.

A synthetic alteration of diabase, with the formation of epidote, albite, calcite, quartz, sericite, and clay, has been carried out in water containing carbon dioxide by Pellizzer (1957). The range of conditions in the experiments was as follows: temperature 375°–540°C, pressure 225–350 kg per cm^2, pH 3.8–9.

The close association and intergradation of strongly to weakly propylitized, zeolitized, and pyritized rocks strongly suggest that all these varieties may be formed contemporaneously as part of the same general alteration process. Differences are attributed to variations in temperature and pressure (depth in part), duration of the processes, and variations in the solutions such as concentration of hydrogen sulfide. In addition, differences in alteration due to composition of the original rocks were demonstrated by Wilshire (1957).

Epidote is the last mineral of the propylitic association to be found in altered rocks of hot-spring systems. The first published observation of epidote in hot-spring systems has been described by Naboko and Piip (1961, p. 110) from the Pauzhetskii Hot Springs of Kamchatka. Dacite and andesite tuffs and breccias show hydrothermal addition of potassium-feldspar and zeolites to a depth of 240 m (800 ft) and are propylitically altered at greater depths. Epidote occurs near and below 750 m (2,400 ft), where the temperature is a little below 180°C. At Wairakei, New Zealand, A. Steiner (oral communication, 1961) has found epidote at depths below 1,800 feet, where temperatures are about 250°C; at Reykjavik, Iceland, G. Sigvaldason (written communication, 1961) has found epidote at depths below 4,000 feet, where temperatures exceed 125°C. Epidote appears to form only at temperatures and pressures that are higher than those found in the near-surface parts of hot-spring systems.

The available evidence suggests that epidote does not ordinarily form at depths of less than 1,500 feet; the fact that it occurs in the altered rocks of the Virginia Range drill hole at depths of more than 500 feet suggests that at least 1,000 feet of volcanic rocks have been stripped from above the present ground level after the alteration occurred.

The generally more intense alteration of the lower part of the section in any given locality supports the dependence of the observed variations on temperature, pressure, and duration; thus in the Comstock Lode district, where the volcanic section is extraordinarily thick and complex, the younger, shallower units are the least altered. The presence in the same area of a plutonic rock of Tertiary age, the Davidson Granodiorite, suggests a more intense heat source there than elsewhere. Moreover, the temperature gradient in the mines is about three times normal and decreases away from the lode in the Sutro tunnel (Becker, 1882, p. 245, 265). Wilshire (1957, p. 258–259), supplementing earlier unpublished work of Curtis in the Ebbets Pass region of California, also describes alteration in early volcanic breccias fading out upward in younger flows. Finally, the more widespread and intense alteration of older rocks is surely due to greater opportunity for alteration by continuous or sporadic activity over a long span of time. Hot water containing CO_2 and H_2S was found in the Comstock mines, and such waters are probably still actively altering the rocks.

The distribution of alteration on a large scale is not markedly controlled by structure, as can be seen by the distribution of bleached rock shown on the geologic quadrangle maps. Instead, the alteration is diffuse because of the rather high original porosity and permeability of these rocks. In local detail, however, altered zones in some places follow faults, and spread out in permeable tuff breccias. Coats (1940, p. 13) also noted propylitization increasing in intensity near systems of steeply dipping minor fractures. The bleached area in the northern part of the Carson Range is more elongated than most, and it lies parallel to the general eastward strike of the volcanic rocks there.

RELATION TO ORE DEPOSITS

The principal Tertiary ore deposits (all in the Virginia City quadrangle) are within altered areas, and the bleached areas have been intensively prospected everywhere, although generally without great reward. The Comstock Lode is a center of intense propylitization and pyritization, and other associations of propylitization with major ore deposits around the world are far too numerous to mention; the great silver deposits of Pachuca-Real del Monte in Mexico are one example.

A possibility that must be considered is that strong and extensive alteration is essential to the formation of certain kinds of ore deposits. More specifically, alteration of big volumes of rock might release and mobilize minor elements that may or may not then become concentrated into ore deposits. Propylitization destroys all glass in the groundmass of the andesites, attacks all ferromagnesian minerals (which contain most of the minor metals of igneous rocks), and, where intense, destroys nearly every original mineral. A concentration ratio of only about 100 to 1 for the gold and silver in ordinary rocks is all that is necessary to make ore of minable grade.

The possibility of deriving the gold and silver of the Comstock Lode from the altered rocks was long ago proposed by Becker (1882, p. 221–225), who, on the basis of assays, considered the hypothesis proved. The idea has not been widely accepted since Becker's time, and his data show such scatter as to be virtually worthless as evidence. The fundamental problem of what happens to the minor elements and how they are redistributed during alteration should be restudied in the search for a basic understanding of ore deposits.

SUMMARY OF VOLCANISM

VOLUME

The volume of Cenozoic volcanic rocks in the two quadrangles is large, as shown in table 3.

TABLE 3.—*Volume of Cenozoic volcanic rocks*

Rocks	Estimated		
	Area (square miles)	Average thickness (feet)	Volume (cubic miles)
Basaltic rocks (Lousetown Formation and McClellan Peak Olivine Basalt)	30	100	1
Andesitic rocks (Kate Peak and Alta Formations)	280	1, 500	84
Rhyolitic rocks (Hartford Hill Rhyolite tuff, Steamboat Hills Rhyolite, and Washington Hill Rhyolite)	8	1, 000	2
Sedimentary deposits (largely derived from Kate Peak and Alta Formations)	125	1, 000	25

The andesitic rocks (Kate Peak and Alta Formations) are by far the most abundant. If the sediments derived from andesites are added, the total of this volcanic material, which is silicic to intermediate composition, amounts to more than 100 cubic miles. This volume may be compared to the volume of a mountain range 10 miles wide and half a mile high extending 17 miles from the northern to the southern boundary of the quadrangle with a volume of 85 cubic miles. The fact that the volume of erupted volcanic rocks is so large clearly indicates a probable genetic relation between volcanism and structural deformation.

LOCATION OF VENTS

Numerous vents for the basaltic rocks (including basalts and basaltic andesites) are easily recognized but those for the Alta and Kate Peak andesites are concealed by later erosion, deformation, and deposition, or are inherently more difficult to recognize. The basaltic vents are generally high in the ranges and are still surrounded by extensive remnants of flows, as can be seen near the dissected cinder cones in the western part of the Mount Rose quadrangle, in Steamboat Hills, and also in the Virginia Range. A single probable vent of Lousetown basaltic andesite lies at the eastern margin of basin sediments 1 mile north of the foot of Geiger Grade (analysis 53, table 1). Elsewhere, if basaltic vents were present in what are now basins, the flows have been largely eroded or buried, a circumstance that is unlikely in the case of the youngest basalts. In addition, basaltic flows have not been recognized in the hundreds of water wells drilled in Truckee Meadows and Washoe Valley. The basaltic vents clearly seem to have been located preferentially in the ranges.

Although the evidence is less direct, the andesite vents seem also to have been mostly in what are now mountains. Many intrusions, some of which surely represent the roots of vents, are evident in the ranges. But are intrusions merely concealed in basin areas? The volcanic complexes seem to thin toward the basins, for pre-Tertiary rocks are commonly exposed along the margins of the ranges (both sides of Virginia Range, Steamboat Hills, and north end of Carson Range). But there are exceptions; the region around Slide Mountain, in the highest part of the Carson Range, is free of volcanic cover, and the lack of Cenozoic intrusions suggests that there were no vents in that region. In places the complexity of the volcanic pile increases toward the higher part of the ranges, as best shown in the Comstock Lode district. Intertonguing of volcanic rocks of the ranges with contemporaneous sedimentary rocks of the Truckee Formation in the basins constitutes further evidence that vents were mainly in the ranges. A genetic connection between volcanism and structural uplift has been suggested (Thompson, 1952).

SEQUENCE OF VOLCANIC ROCKS

The sequence of volcanic rocks is not rigidly related to their composition, which ranges widely. Rhyolite and basalt, for example, are among the most recent volcanic products. On a broader scale, however, the Hartford Hill Rhyolite Tuff is followed by the andesites and these in turn by widespread basaltic rocks. The basalts are relatively small in volume but the younger rhyolitic rocks are much less abundant, and are far smaller in volume than the early Hartford Hill. These relations are necessarily based on observations over a larger region than the two quadrangles. There is thus a general evolutionary trend from rhyolitic through andesitic to basaltic rocks, but there is also much deviation from this general trend.

STRUCTURE

The structural history is divided into two parts, a Mesozoic orogeny and the Miocene to Recent deformation that shaped the present basins and ranges. Virtually nothing is known of the transition between these two periods of deformation because rocks of early Cenozoic age are lacking.

A brief summary of the pre-Tertiary structure is given in the following paragraphs. The Cenozoic structure is then described for each of the ranges and basins. Finally, the problems of the Cenozoic deformation and its origin are discussed.

PRE-CENOZOIC DEFORMATION

The pre-Tertiary sedimentary and volcanic rocks were intensely folded, regionally metamorphosed, and intruded by granitic rocks. Only a few details have been added to this general history, for pre-Tertiary rocks are concealed over most of the area and of the exposed part most is granitic rock.

The strike of the metamorphic rocks ranges from north-northwest to northeast and has a common trend to the north-northeast, and the steep dips generally range from 45° east to 45° west. Subtracting the effects of Cenozoic deformation does not change this range appreciably. In Steamboat Hills and in the hills lying to the southeast, the average strike is a little east of north. In the Virginia City quadrangle, the sedimentary rocks near Castle Peak and those south of Virginia City strike north-northeast. Metamorphic rocks in the northwestern part of the Mount Rose quadrangle are locally distorted near Cenozoic faults.

The attitude of the metamorphic rocks is of interest because it is a measure of the physical anisotropism that must have influenced all later structures. It seems especially significant that the Cenozoic normal faults have an average trend of north-northeast, parallel to the average trend of the dominant Mesozoic structure.

The granitic rocks, where they exhibit any well-marked internal structure, have a planar structure apparently developed in the last stages of intrusion. This foliation is well developed only in the southwestern part of the Virginia City quadrangle. A weaker foliation is plotted in a few places where it is fairly prominent, as along the north side of Ophir Creek east of Slide Mountain. The average attitude of the foliation is nearly north and steep, like the bedding of the metamorphic rocks. The anisotropism of the granitic rocks may also have influenced the Cenozoic structures.

The best-exposed intrusive contact of the granitic rocks is southeast of Little Washoe Lake, where the contact is exposed for a length of more than 2 miles. The trend is west-northwest and the dip is steep. The granitic rock here cuts directly across the strike of the country rocks but sends swarms of aplite and pegmatite dikes and quartz veins along the bedding. Except for the concordance of dikes and small apophyses, discordance between intrusive contacts and stratification of the country rocks is the general rule both in the quadrangle and in the nearby areas, but the granitic rock near the southwestern corner of the Virginia City quadrangle is exceptional and forms a concordant intrusive contact with the metamorphic rocks.

CENOZOIC STRUCTURE

PREVIOUS WORK

Previous workers have had considerable divergence of opinion on the Cenozoic deformation in this general region. One extreme is represented by Lindgren (1897, p. 6) who found (with one exception) "no conclusive evidence in favor of post-andesitic faulting on either side of the Carson Range" in the Truckee quadrangle. He thought that the Tahoe-Truckee depression was outlined by pre-andesite faulting On the other hand, he recognized recurrent fault movements during the period of volcanism "along the old break at the eastern base of Carson Range," but he regarded the movement along the mountain front north of Mount Rose as pre-andesite (Lindgren, 1911, p. 194). At the other extreme, Louderback (1907, p. 667) listed a post-andesite fault with vertical throw of 5,500 feet at the same locality. Reid (1905 and 1911) regarded the topography as a direct expression of late faulting very little modified by erosion. Louderback (1903, 1907, 1923, 1924) established that extensive normal faulting took place after the eruption of the andesite and the deposition of the Truckee sediments. The general sequence according to Louderback (1924, p. 4 and 5) is: (1) Tertiary volcanic rocks deposited on a surface of pre-Tertiary rocks; (2) continental sediments deposited on the volcanics, generally with only slight angular discordance; (3) deformation of the sediments into gentle folds; (4) erosion surface of very gentle relief cut across the sediments and volcanics; (5) basic lavas poured out on this erosion surface;

(6) deformation of the erosion surface and its partial cover of basic lavas, mainly by block faulting; and (7) dissection.

The evidence from the present geologic work in the Mount Rose quadrangle necessitates fundamental revisions of these earlier viewpoints. The following points are worth emphasizing: (1) The ranges, though much lower and smaller in extent, were already roughly outlined in early or middle Pliocene time when the Truckee sediments were being deposited. This is indicated in the Truckee Formation by the abundant coarse clastic material that could have been derived only from highlands adjacent to the basins of deposition. It is indicated also by the thickness of the Truckee Formation of 2,000 feet or more in the basins. (2) Uplift of the Carson Range in the northern half of the quadrangle, where the stratigraphic evidence is conclusive, was by flexure. The faults alone, many of which are antithetic, would make the area a structural and topographic depression (pl. 1, section *C–C′*). This uplift must have been partly coincident with Truckee sedimentation, to account for the thickness of the sedimentary rocks. (3) The high pre-Lousetown erosion surface was dissected by canyons at least 1,000 feet deep before eruption of the basic lavas; the steep mountainous topography preceded, rather than followed, the lava flows. (4) There is no great post-high erosion surface and postbasalt uplift along normal faults in the northern half of the quadrangle. Rather the present topographic relief was produced by continued or renewed flexing and by great erosional excavation of Truckee sediments from the margins of the basins.

The relations outlined are strictly applicable only in the northern half of the quadrangle. Farther south, it is clear that normal faults are responsible for more of the structural relief, although warping is also important. As stated long ago in a discussion of basin-range structure, true anticlines are very rare "but many ranges are built of faulted and dislocated rock masses with imperfect anticlinal arrangement" (Powell, 1876, p. 24, quoting G. K. Gilbert).

CARSON RANGE

The Carson Range splits from the main Sierran block 25 miles south of the Mount Rose quadrangle. It extends northward and diverges from the Sierran block, which trends north-northwest; the intervening structural depression is the Tahoe-Truckee trough. The range ends in the valley of the Truckee River along the northern border of the Mount Rose quadrangle.

The warped and faulted block which forms the Carson Range has a structural relief with respect to bedrock of the basins on its east side of more than 8,000 feet. Considering the great depth of Lake Tahoe (1,600 ft) lying to the west, the structural relief on that side must be about as great. On the north and northeast sides the range forms a partial dome, which is broken by many normal faults, most of which are antithetic; that is, the displacement is down on the mountain side of the faults. Farther south, along the east side, the domical boundary is replaced by great normal faults along which the range is uplifted with respect to the basins. In the interior of the range, normal faults are abundant and the strata are complexly warped and tilted, but the overall structure is that of a faulted anticlinal uplift.

NORTHERN PART OF RANGE

It will be convenient to discuss first the structure of the Carson Range in the northern half of the quadrangle, north of Whites Canyon. This part is bounded by a domical fold, or plunging anticlinal structure, which is well marked on the geologic map by the lower contact of the Truckee Formation where the beds dip away from the dome at an average angle of 35°. Along its northeastern part, near Evans Creek, the dome is complexly faulted and bulged eastward toward Huffaker Hills; the Truckee Formation, if present at all in this part of Truckee Meadows, is concealed by alluvium. A remarkable set of concentric and radial faults also borders the dome. The concentric faults with few exceptions have their downthrown sides toward the interior of the dome. The structural relief produced by faulting thus subtracts from that due to upwarping. The total effect is like that above salt domes, as if the strata of the dome were stretched over the rising interior, causing the central part to fail along normal faults and form grabens.

RELATIVE DEFORMATION OF TRUCKEE AND KATE PEAK FORMATIONS

The first point to be noted is the very small structural discordance of the Truckee and Kate Peak Formations. In most places along the boundary the measured dips in the two formations agree within 5° or 10°, and this fact is remarkable in view of the high initial dips that are possible in the flows and breccias of the Kate Peak Formation. Further evidence for approximate structural concordance can be found in the intertonguing of the two formations. This relation is illustrated by the tongue of Truckee Formation in the Kate Peak half a mile north of Fuller Lake and by tongues of stratified breccia of the Kate Peak Formation in the Truckee along the south side of the Truckee River half a mile west of Hunter Creek and a quarter of a mile north of the quadrangle. Intertonguing on a small scale can be seen at many other places.

In contrast to the evidence of general structural conformity is much evidence of local unconformity because of erosion and nondeposition. The Truckee was laid down directly on granitic rocks near Lawton in the basin north of the quadrangle, and some beds of sandstone are composed mainly of sand derived from the granite. Near Fleish, in the northwestern corner of the quadrangle, the basalt member of the Truckee, with a few feet of gravel at its base, was laid down on metamorphic rock. Elsewhere, the Kate Peak Formation directly beneath the Truckee Formation may consist of either flows or breccia, weathered or unweathered. Similar lensoid depositional features occur within the Truckee Formation as well as at its base and are to be expected in any series of stream, lake, and volcanic deposits. Everywhere that the two formations lie in depositional contact, however, they are deformed to roughly the same degree.

FOLDS

In addition to the main domical structure of the northern Carson Range, a few small folds occur. They are best shown in the Truckee Formation, where they are generally small, diverse in trend, and commonly plunge steeply. The small syncline along the northern border of the quadrangle 2 miles west of Hunter Creek can be traced for more than a mile in an easterly direction. It probably marks the axis of the structural basin between the Carson Range and Peavine Peak to the north (pl. 1, section *A–A'*). It plunges eastward and disappears against a cross fault. A mile to the east another small syncline appears; it trends northeastward for a short distance and then disappears east of Hunter Creek in a zone of eastward dips where the basin becomes much deeper. Other small folds plunge steeply to the north or northeast off the main domical uplift. One of these folds lies half a mile east of Hunter Creek and another 3½ miles west of it.

Small folds are much more difficult to recognize in the heterogeneous Kate Peak Formation. Two are mapped in the northern part of the range; one is 2 miles south and east of Fuller Lake and the other is in the headwaters of Thomas Creek. Similar folds lie farther south. All of these folds trend a little west of north.

The folds are everywhere broken by numerous small faults which appear to have developed contemporaneously with the folding.

FAULTS

The faults, so far as is known, are all normal faults, based on the evidence of a few of the large faults and of many small ones exposed in irrigation canals and in roadcuts (see Anderson, 1909, p. 479). The faults in the interior of the range trend mainly west of north, parallel to the folds. Along the north and northeast margins of the range, however, the faults are roughly concentric and radial to the edge of the dome. The two concentric faults that bring granodiorite to the surface in the northwest part of the quadrangle (one at Fleish and the other 2 miles to the east) are of special significance and are discussed separately in the next section.

The radial faults and those within the range are downthrown either to the east or west without apparent pattern. The concentric faults, with a few exceptions, are antithetic and are downthrown to the south or west toward the interior of the dome. Along one fault, between Evans and Thomas Creeks, sandstones and shales of the Truckee Formation are downthrown against the Kate Peak rocks (pl. 1, section *C–C'*). This fault dips 40° westward and the stratification dips 35° eastward. The abnormally low dip for a normal fault can reasonably be explained if the faulting began at nearly the same time as the doming and if the fault initially dipped 75°. As doming and faulting proceeded together, the fault plane was gradually rotated to a dip of only 40°. Evidence for the displacement on some of the other antithetic faults on the east side includes longitudinal topographic troughs along the mountain fronts, some containing closed depressions with small lakes. Gravel beds mapped locally as key beds near the top of the Kate Peak Formation also provide stratigraphic evidence of the displacement (pl. 1, section *G–G'*).

TRAPDOOR FAULT BLOCKS AND DEFORMATION BY ANDESITE INTRUSIONS

Two blocks of granodiorite and metamorphic rock, one at Fleish and the other 2 miles to the east, protrude from volcanic rocks of the Kate Peak Formation. They are bounded on their south sides by short faults with large displacement (pl. 1, section *A–A'*). These structures can be visualized as thick trapdoors, opened along faults on their south sides and principally hinged or bent on the other sides. Oversteepening of beds by the uplift is well shown north of the eastern block.

The large intrusions of Kate Peak andesite in the northwestern part of the quadrangle suggest an explanation for the trapdoor faults. The intrusion 3 miles south of Fleish contains enormous blocks of granitic rock, the largest of which is nearly half a mile across and lies more than 1,000 feet above its normal position in the canyon of the Truckee River (pl. 1, section *A–A'*). Chilled contacts of the andesite against the granitic rock are well exposed. Several smaller inclusions of granitic rock are shown on the geologic map, and there are also many blocks too small to show. Granitic inclusions also appear near the southeast end of the elongate intrusion a mile south of Fuller Lake. Granitic and also metamorphic inclusions occur in the vitrophyric intrusion a mile east of Fuller Lake.

Finally, the vitrophyric intrusion half a mile southeast of the eastern trapdoor block is locally crowded with innumerable inclusions of granodiorite.

These relations suggest that the trapdoor blocks were probably lifted by Kate Peak intrusions, as indicated on plate 1, section *A–A′*. The size of the intrusions that are exposed and their demonstrable ability to lift large granite masses more than 1,000 feet prove the adequacy of this explanation. The time of faulting is indicated by the very thick section of Kate Peak breccias that strike and dip into the western trapdoor block on its south side. On the north side these rocks are very thin or missing below the Truckee Formation. These facts show that faulting accompanied or closely followed deposition of the Kate Peak extrusive rocks and that the thin or missing section on the north side must be explained either by nondeposition or by erosion prior to deposition of the Truckee Formation. Fault movements certainly ceased before the lavas of the Lousetown Formation flowed over the eastern fault. Thus the time relations are entirely consistent with a theory of uplift of the blocks by Kate Peak intrusions contemporaneous with or slightly later than the local extrusions.

Another instance of probable deformation by an intrusion was indicated in the anticline lying west of Virginia City in the adjacent quadrangle (Thompson, 1956, p. 65). Early movements on the Comstock fault, which lies east of the anticline, are also inferred to be connected with the intrusion of the Davidson Granodiorite.

At this point it is tempting to explore the possibility that concealed intrusions are responsible for some of the larger structures, perhaps even the domical form of the northern Carson Range. The large bleached area extending for 6 miles across the northern end of the Carson Range may indicate voluminous intrusions at depth, but this is far from certain. The abundance of volcanic vents and intrusions in the ranges as compared to the basins has already been mentioned and further suggests a direct connection of volcanism with structural deformation.

LATEST DEFORMATION

Although structural movements continue to the present, as indicated by occasional small earthquakes in the area, the youngest deposits that clearly record structural deformation are the lavas of the Lousetown Formation and the deposits of Quaternary alluvium. When the Lousetown volcanoes were erupting, the main features of the present topography and structure were already established. The Truckee River north of the quadrangle has cut down no more than 400 feet since the eruptions. Canyons like that of Thomas Creek were already formed and have since been deepened about 400 feet. Faulting of the flows and of alluvium occurred in some areas but not in others.

Lava flowed down an erosion surface on the north slopes of the Carson Range and is still preserved in small remnants. In this area there is no certain evidence that the flows have been faulted, although warping, which would be difficult to detect, possibly occurred. Northeast of Fleish, patches of lava extend northward onto remnants of the highest pediment surface cut on the Truckee Formation. This pediment projects toward the highest river terrace, which lies about 400 feet above the Truckee River on Chalk Bluff, north of the quadrangle. Fault displacements of the highest pediment or of younger pediments on the north slope are generally lacking, but a few faults exposed in excavations north of the quadrangle cut terrace gravels. One of the most prominent of these is directly north of and in line with the hot spring at Lawton.

On the east slope of the range south and southeast of Hunter Lake, as in Steamboat Hills and the Virginia Range, the Lousetown flows are cut by faults and were apparently tilted some at the time of faulting. Farther east the pre-Lake Lahontan alluvium is also cut by swarms of small faults; this general area has obviously been more active during the Pleistocene than the area north of the range.

MOUNT ROSE-SLIDE MOUNTAIN AREA

To the south, faulting plays a more prominent role in forming the boundaries of the Carson Range. The range is still a broad upwarp, which is capped at Mount Rose by a horst (pl. 1, section *D–D′*) and farther southwest by a graben (pl. 1, section *F–F′*).

East of Mount Rose, on the slope of the range, well-stratified tuff-breccias of the Kate Peak Formation are downthrown to the east along a fault of large displacement (pl. 1, section *D–D′*; fig. 15), which can be traced for only a mile. It is interesting to note that this fault, which was described by Louderback (1903, p. 344), has little direct topographic expression. Farther down on the east slope the eastward dip of the volcanics accounts for part of the structural relief.

A large fault cuts the granodiorite on the east flank of Slide Mountain, but along this fault downthrow is to the west. An excavation along this antithetic fault near Grass Lake exposed a wide crushed zone in granodiorite dipping westward 70°. The longitudinal valley formed along the fault contains Hidden Lake and Grass Lake, which are of glacial origin. Parallel and farther east is another large fault which bounds the high part of the range and has normal downthrow to the east.

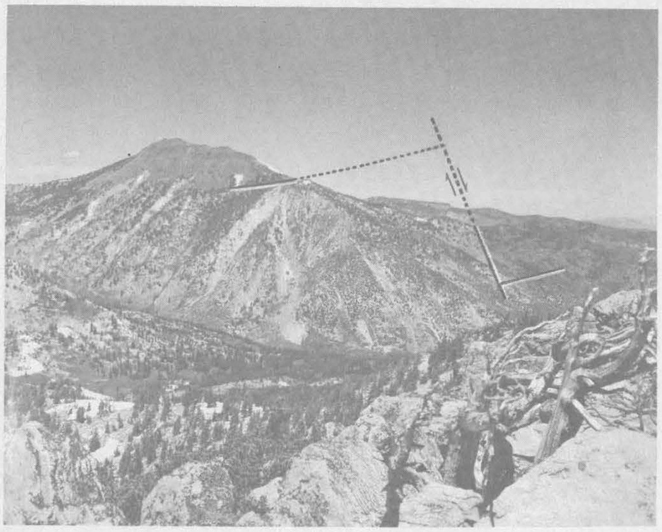

FIGURE 15.—Andesite of Kate Peak Formation, overlying granodiorite, caps Mount Rose and is dropped down along a normal fault. (See also pl. 1, section *D–D'*). View from Slide Mountain across the glaciated valley of Galena Creek.

West of Mount Rose and Slide Mountain the normal faults are mostly antithetic; they displace the interior of the range downward with respect to the flanks. In the same area, stratification dips east and west away from the interior of the range. The anticlines shown on the geologic map, like the faults, generally extend for only a mile or two.

LITTLE VALLEY

Southeast of Slide Mountain the longitudinal trough of Little Valley interrupts the precipitous escarpment of the eastern front of the Carson Range (fig. 10). The valley is half a mile in width and its floor is alluviated. Alternative structural interpretations of the origin of the valley are possible; the interpretation must be based largely on topographic evidence for the rock is granitic except for a small intrusion of basalt. On the west side of the valley an escarpment rises 2,000 feet to the summit of the range. East of the low hills forming the east side of the valley another escarpment drops 1,600 feet to the foot of the range in Washoe Valley. From their magnitude and form both of these escarpments must represent large faults. A map published by Reid (1911) shows rhyolitic rocks lying south of the quadrangle displaced where the two faults would project. Little Valley may thus be likened to the tread of a step and the two faults to the risers above and below the step.

This simple picture does not account for the hills forming the east side of Little Valley. One possible interpretation is simply that erosion along the western fault zone has lowered the back of the step with respect to its outer edge, but this hypothesis does not satisfactorily account for the alluviation of the valley floor. A second possibility is that the step was tilted westward. And a third possibility is that a subsidiary antithetic fault on the east side of Little Valley makes the valley a graben block (pl. 1, section *H–H'*). The last interpretation is favored because other antithetic faults are present in the area; in particular the fault cutting the east flank of Slide Mountain and passing through Grass and Hidden Lakes projects southward in line with Little Valley.

VIRGINIA RANGE

Although only a small part of the Virginia Range lies in the Mount Rose quadrangle, an understanding of the structure is essential in relation to the Carson Range and the intermontane basins.

On the east side of Washoe Valley the structure of the Virginia Range resembles a mirror image of the structure of the Carson Range. East and west of Washoe Lake the ranges are bounded by frontal faults along which the basin block is depressed. On the slopes of the ranges are antithetic faults with longitudinal valleys formed along them (pl. 2). Farther north along Washoe Valley the rocks in both ranges are tilted downward toward the basin. Faulting is less important than warping in outlining the ranges, although faults are numerous and complex in pattern. The dip of the rocks in both ranges toward the axis of the basins continues, with many complications in details, to the northern limit of the Mount Rose and Virginia City quadrangles.

Broadly the Virginia Range, in contrast with the domical or plunging anticlinal structure of the northern Carson Range, consists of west-tilted blocks bounded by normal faults. Open folds can be discerned in a few places and one larger anticline occurs near the mass of Davidson Granodiorite west of Virginia City. The irregular boundaries of both the Carson and Virginia Ranges are determined by folds and faults of the same systems that comprise the internal structures of the ranges.

STEAMBOAT HILLS

Steamboat Hills is in many ways a miniature replica of the major ranges except that its long axis trends northeast, transverse to the trends of the ranges. The hills have an anticlinal form produced by warping and by tilting of fault blocks (pl. 1, section *E–E'*). On the southeast flank of Steamboat Hills, north of Pleasant Valley, basal tuff-breccias of the Kate Peak Formation dip 45° to the southeast under the valley and reappear again on the southeast side of the synclinal structure in Pleasant Valley. Pre-Tertiary rocks crop out in many places near the crestline of the hills and in general decrease in altitude to the northwest and are concealed by younger volcanic rocks. The structural relief, as in the Carson Range, has been produced by a combination of normal faults dipping away from the hills and tilting that more than compensates for antithetic faults dipping toward the hills.

At least three systems of normal faults have been recognized in the hills. One major set strikes northeast, parallel to the axis of the hills; many of these are antithetic. A second set of faults strikes northwest, nearly at right angles to the first. The third set strikes nearly north and is prominent near Steamboat Springs; many of this set are antithetic in dipping toward the structural crest of Steamboat Hills but their strikes are not parallel to the axis of the hills. Although faults of the three systems are not of distinctly separate ages, those of the north-striking system show evidence of being most active recently, for many of them cut the pediment surface developed on pre-Lake Lahontan deposits. On the other hand, few of those of the northwest and northeast systems cut this surface.

The western bounding fault of Steamboat Hills has been one of the most active in late geologic time. It cuts the alluvium along part of its trace, and in its strike and sense of displacement it obviously belongs to the same extensive system of faults that cuts the pre-Lake Lahontan alluvium west and north of the hills (fig. 16 and pl. 1). A fault bounding part of the southeast side of Steamboat Hills at Pleasant Valley must also have moved fairly recently to account for the lack of larger fans where Galena and Steamboat Creeks debouch on the floor of Pleasant Valley.

Several andesitic intrusions of the Kate Peak Formation, a dome of pumiceous rhyolite, and a vent for basaltic andesite, indicate that volcanism was active in the hills over a long span of time.

Many thermal springs issue from fissures and isolated vents over an area of about a square mile near the northeastern end of the hills near the western edge of the Virginia City quadrangle. The hills lie in the midst of an area of faults of Quaternary age which are well shown in the alluvium west and north of the hills (fig. 16). Continuing thermal activity is undoubtedly connected with late structural movements and volcanism.

INTERMONTANE BASINS

The trough between the Carson and Virginia Ranges is divided into a chain of basins by transverse hills extending out from the ranges. The principal individual basins are Truckee Meadows, which is constricted and divided into two parts at the Huffaker Hills, and Washoe Valley. Between these main basins lie the smaller Pleasant and Steamboat Valleys, the latter entirely in the Virginia City quadrangle. West of Reno the basin of Truckee Meadows protrudes westward to form the Verdi basin, which bounds the Carson Range to the north. The pattern of irregular chains of basins continues north of Truckee Meadows and south of Washoe Valley. Branching chains of oval basins scattered among ranges that G. K. Gilbert compared to an "army of caterpillars marching toward Mexico" are in fact characteristic of the Basin and Range province.

The basins, which are generally roughly oval in plan, are not simple grabens but are formed by faulting, tilting, and warping. Moreover, instead of being built of simple blocks with distorted margins, the basins, like the ranges, are distorted throughout, and the structural level of their bottoms results from the summation of many smaller structures, some of which are additive and some subtractive. Finally, erosion and deposition played a major role in shaping the present topographic basins.

Because of their oval shape and their size, it is tempting, in view of the abundant volcanic manifestations, to compare the basins with calderas or volcano-tectonic depressions, as Pakiser, Press, and Kane (1960) and previously Williams (1941) have concluded for the Mono Basin. But the comparison suffers in the details, for the volcanic rocks seem to be thickest in the ranges, and pre-Tertiary rocks are exposed at many places on the margins or within the basins. Probably a less direct connection exists between volcanism and basin-range deformation. Some sort of probable, although quantitatively poorly understood, relation has long been recognized by many geologists at many places. The association of Permian igneous activity and faulting in the Oslo and Rhine grabens of Europe, the Triassic igneous activity and block faulting in Eastern United States, the Cenozoic faulting and volcanism of the African rift valleys, and the widespread volcanism of the Basin and Range province are familiar examples. On the other hand, block faulting is by no means always accompanied by volcanism nor vice versa.

TRUCKEE MEADOWS

The part of Truckee Meadows lying northwest of the Huffaker Hills will be discussed first. The western side of this basin is formed by the dome of the Carson Range. Numerous normal faults along that side are antithetic; they step the strata downward to the west, in opposition to the eastward dip of the beds. The deepest part of the basin, containing roughly 2,800 feet of sediments as indicated by gravity measurements, lies to the north of Huffaker Hills (Thompson and Sandberg, 1958, p. 1275). The southeastern side of the basin is probably bounded by a system of faults concealed by young alluvium along the Huffaker Hills. Faults of this system are clearly shown in the pre-Lake Lahontan alluvium southwest of the hills. To the northeast, where the Huffaker Hills join the Virginia Range, faulting is re-

FIGURE 16.—Vertical aerial photograph showing swarms of en echelon fault scarps cutting alluvium west and north of Steamboat Hills. Width of photograph 4 miles; north is at the top. The two conspicuous creeks are Thomas and Whites; Steamboat Hills in southeastern part of photograph.

inforced by westward dips in the Virginia Range to form the eastern margin of the basin. The major structural break is apparently not at the base of the range but about a mile to the west, concealed under a thin cover of alluvium (Thompson and Sandberg, 1958, p. 1275–1276).

The basin underlying the part of Truckee Meadows south of Huffaker Hills is deepest along Steamboat Creek a little to the east of the Mount Rose quadrangle and immediately north of Steamboat Springs (Thompson and Sandberg, 1958, p. 1274); the gravity evidence indicates a depth of fill in the order of 1,000 feet. The western side of the basin is formed by the east-dipping flank of the Carson Range. The normal faults in pre-Lake Lahontan alluvium north and northwest of Steamboat Hills face mostly westward and thus oppose the dip, but a few face eastward (fig. 16 and pl. 1). The eastern side of the basin, along the Virginia Range, is formed mainly by the westward-dipping strata in that part of the Virginia Range and is thus symmetrical with the western side, but faulting complicates the structure.

FAULTS SOUTHWEST OF TRUCKEE MEADOWS

A system of fault scarps with topographic expression and displacement of an erosion surface is strikingly displayed on the alluviated slopes southwest of Truckee Meadows (fig. 16). The displaced surface is a gravel-covered pediment, cut partly on Kate Peak Formation but mainly on pre-Lake Lahontan stream deposits, probably of Sherwin age. These faults exhibit characteristics common to all the normal faults of the region: (1) their traces are individually short and curved or angular, with a change in strike of as much as 60°, (2) they form an overlapping or intersecting pattern, and (3) the vertical displacement along many of them is opposed to the structural relief of the ranges or basins within which they occur. These characteristics explain why the major structures such as mountains and basins can be understood only in terms of the aggregate structural relief of warping, tilting, and of faulting along swarms of individual breaks.

To the northeast in Truckee Meadows similar faults are presumably concealed beneath younger alluvium, which is still being actively deposited. Sporadic fault movements over a long part of the Quaternary Period have cut up the older alluvium but in the short span of existence of the younger alluvium have not yet deformed it appreciably. A fault immediately north of Steamboat Hills (sec. 30) exhibits repeated movement especially well where it is crossed by Whites Creek. This fault and a second fault about a mile to the west (sec. 25) bound a horst, across which Whites Creek has cut a narrow arroyo. The scarp of the eastern fault has a maximum height of about 35 feet; former stream channels in the horst block are not truncated at the fault, and a cut terrace adjacent to Whites Creek demonstrates that faulting occurred at least twice and was separated by a period of erosion.

VERDI BASIN

The Verdi Basin, which contains a large volume of sedimentary rocks of the Truckee Formation, projects westward from the northern part of Truckee Meadows at Reno. Only its southern part, in which the beds dip northward off the dome of the Carson Range, lies in the Mount Rose quadrangle. The deepest part of the basin is west-southwest of Reno, near where Alum Creek leaves the quadrangle. Gravity measurements indicate at least 1,800 feet of sedimentary rocks in that area, thinning westward to no more than a few hundred feet west of Hunter Creek (Thompson and Sandberg, 1958).

STEAMBOAT AND PLEASANT VALLEYS

The relatively small basins of Pleasant and Steamboat Valleys are similar in that both have fairly young concealed faults along their northwestern sides.

Detailed gravity measurements suggest that the bedrock floor of Steamboat Valley was tilted westward against faults and was buried by sediments to a maximum depth on the order of 500 feet (Thompson and Sandberg, 1958, p. 1273–1274). To the north and south a gentle rise of the floor of the valley probably forms the structural boundary at both ends. The Steamboat fault projects southwestward from Steamboat Springs and may split, with a western branch close to the west margin of the valley and an eastern branch projecting south-southwest to the large fault a mile east of Pleasant Valley. This large fault has been inactive south of Steamboat Valley for a long time, however, for remnants of the pediment surface upon which the Lousetown flows of Steamboat Hills were erupted projects to the south and east across the fault, with little or no displacement.

Near the north end of Steamboat Valley and farther to the north the fault system provides the structural control for Steamboat Springs. Drill hole and gravity data demonstrate an absence of appreciable displacement of the granitic bedrock contact on the prominent system of fissures that control thermal activity of the principal spring terraces. Any major movements that may have occurred on the fissures were earlier than stream deposits here included with the basal pediment gravels of the Lousetown Formation. The largest fault of the system appears to be at the east base of the Main Terrace. A well drilled in 1959 by Nevada Thermal Power Co. penetrated volcanic rock (probably soda trachyte of the Alta Formation) at a depth of 350 feet, and granodiorite at 1,050 feet. In a drill hole only 750 feet to the west and 90 feet higher in altitude, andesite of the Kate Peak Formation was cut at a depth of 135 feet, and granodiorite at only 154 feet. A total vertical displacement of at least 1,000 feet is indicated for the large fault; two or more periods of movement are indicated, with major displacement older than the Kate Peak Formation, smaller but significant movement younger than the Lousetown Formation, but probably none that is younger than the pre-Lake Lahontan deposits.

Along the west side of Pleasant Valley the young gorge cut by Galena Creek through the Steamboat Hills and the small fan where the Creek debouches into the valley give evidence of recent faulting. Indications that the valley has also sagged to form a syncline are found in the valleyward dip of the Kate Peak Formation on both sides. Like Steamboat Valley, Pleasant Valley is probably limited at the ends by gentle inward plunge of the bedrock floor.

WASHOE VALLEY

Washoe Valley drains toward the interior during the dry season, but when precipitation is plentiful Washoe Lake overflows through Little Washoe Lake into Pleasant Valley. Faults partly bound the basin on both sides. Those on the west side, at the foot of the Carson Range, have been active late enough to displace the pre-late Pleistocene landslides east of Slide Mountain. Gravity measurements also indicate that the greatest thickness of sediments—at least 1,800 feet—underlies the western part of the basin (Thompson and Sandberg, 1958, p. 1272–1273).

The bounding faults die out northward, and there the base of the Kate Peak Formation on both sides of the basin dips gently basinward. At the north end, several small faults project southward into the basin, where they are concealed by the young alluvium.

At the outlet of the basin, Steamboat Creek occupies a peculiar position. It is entrenched in a narrow gorge 250 feet deep, which is half a mile west of the low pass where Route 395 leaves Washoe Valley. These relations are interpreted as indicating eastward tilting at the north end of the Valley or an eastward shift of the axis of this part of the structural syncline in late Quaternary time after entrenchment of the outlet had started. The faulted and uplifted lake beds half a mile west of Little Washoe Lake are also consistent with this interpretation.

The geologic history of Washoe Valley has certainly been complex, although most of the record is concealed beneath the youngest sedimentary deposits. In addition to deformation along the margins of the basin near the north end, which has already been mentioned, old lake beds a mile west of Washoe Lake at the south edge of the quadrangle are upfaulted and are now being eroded. Old gravel caps a hill of granodiorite in the Virginia Range near the eastern edge of the Mount Rose quadrangle, and the surface of the gravel deposit projects toward a gravel-covered hill a mile to the west in Washoe Valley. Erosion followed by later sedimentation has destroyed the connection between these isolated patches of gravel. Washoe Valley has probably contained larger lakes repeatedly while Quaternary structural deformation competed with erosion and deposition.

AGE OF THE BASINS

The question of when the basins came into existence is largely a problem of the location of basins of deposition of the Truckee Formation, which comprises the oldest sedimentary deposits preserved in any of the present basins. These deposits are extensively exposed in the Verdi Basin and at the foot of the Carson Range west of Truckee Meadows but are concealed if present in the basins farther south. Remnants also remain in the ranges, at altitudes of as much as 6,500 feet near Fuller Lake, 8,700 feet in Galena Creek, and 7,000 feet at Chalk Hills in the Virginia Range. Do these high remnants mean that the sedimentary deposits once extended continuously across the present ranges and have since been eroded?

The evidence indicates instead that the general position of highlands and lowlands during deposition of the Truckee Formation was roughly similar to the present distribution, although boundaries of the basins certainly shifted and large changes occurred in the magnitude and location of topographic features. This conclusion is based partly on the local source of much of the sedimentary debris, most of which is derived from the Kate Peak Formation. Near its base the Truckee Formation contains cobbles and boulders that could have been transported only a few miles; this indicates highlands in close proximity to the depositional basins. Partial intertonguing of the Truckee Formation in the basins with the Kate Peak Formation in the ranges supports the same conclusion. The fact that structurally active basins, which are undergoing sedimentation at present, contain very thick deposits from the past also argues for the persistence of basins in approximately their present locations. That the margins of the basins shifted and that some early basins have become parts of the ranges, however, is shown by the remnants of Truckee Formation in the ranges.

In summary, basins were formed during the eruption of the Kate Peak volcanic rocks (at least during the later stages) and as sediments accumulated during and after the principal volcanism, structural deformation continued, perhaps sporadically; basins were deepened, a few new ones were formed, and perhaps some old ones were destroyed by erosion; but the general locus of uplands and lowlands may have persisted from early Pliocene to the present.

REGIONAL STRAIN

It has been pointed out (Thompson, 1956, p. 64) that in this region the horizontal components of strain due to normal faulting and folding (or warping) are opposite in sense, and since the two kinds of deformation were contemporaneous, the net strain should be less than appears from normal faulting alone. Furthermore, the simplest explanation for the Cenozoic deformation is differential vertical uplift (or depression), which results in both folding and normal faulting; either one alone would have to be accompanied by large horizontal movements unless the faults are vertical, which they seem not to be.

This explanation works particularly well for such small structural units as the trapdoor fault blocks in the northwestern part of the Mount Rose quadrangle. It also has considerable appeal as an explanation for some large structural units such as the partial dome and associated faults forming the northern part of the Carson Range.

In some areas, however, normal faults are certainly too numerous for the horizontal component of dip slip to be balanced by folding and tilting. The swarm of late Pleistocene faults cutting old alluvium southwest of Truckee Meadows, for instance (fig. 16 and pl. 1), must represent a considerably greater extension than can be balanced by tilting and folding in the same area. Similarly the horizontal extension on normal faults measured along section *F–F′* (pl. 1) exceeds the contraction on folds by a factor greater than two, although along section *D–D′* the extension and contraction are roughly equal. In general, many faults too small to map appear in good exposures and must be cumulatively important, but the same is not true of folds. Still further evidence of an excess of extension over contraction during basin-range deformation is supplied by geodetic measurements of deformation that accompanied the Dixie Valley-Fairview Peak earthquakes in 1954 (Whitten, 1957). These measurements show that although the valley block was tilted at the time of faulting, the net extension is equal to that expected on the faults alone (Thompson, 1959). One must conclude that the basin and range region has been undergoing net lateral expansion even though part of the extension caused by normal faulting is offset by tilting and folding.

The direction of maximum principal strain (extension) could be estimated if the normal faults were purely dip slip. As this condition is unlikely and as it is generally impossible to measure a component of strike slip on the faults, we may resort to considering only the change of area. The dip-slip component of movement on a normal fault causes an increase in area, whereas a strike-slip component has no effect on area. One may say that the horizontal projection of the strain ellipse is increased in area by elastic strain preceding normal faulting but is merely changed in shape by that preceding strike-slip faulting.

A rough estimate of 5 percent increase in area has been made for the basin-range region (Thompson, 1959). Study of the maps and sections of the Mount Rose and Virginia City quadrangles indicates that a 5 percent expansion is a reasonable order of magnitude in that area also.

PROBLEMS OF ORIGIN

Most of the Basin and Range province, like the Plateau province to the east, stands at a high altitude, but unlike the Plateau province, it is broken up by block faults. The reasons for this difference are not clear, but as Nolan (1943, p. 185) pointed out, the block faults are restricted to the region of earlier folding, a fact that indicates a deep-seated origin and a long duration for the difference.

The origin of basin and range structures and of the broad highland on which they are superimposed is probably tied up with volcanism (as discussed previously in the section on intermontane basins) and with changes in the crust-mantle system of the earth. The inferred expansion of area which was discussed above seems best understood if the upper part of the crust was carried along on an expanding substratum.

Recent evidence of crustal structure in the Basin and Range province (Press, 1958; Berg and others, 1959; Thompson and Talwani, 1959) is consistent with this conclusion. Here, the crust and anomalous upper mantle (defined as material of lower velocity than normal mantle) together are thick, and the lower part of this system is transitional in physical properties between normal crust and normal mantle. These conditions must represent in part the additions of low density material necessary for isostatic compensation of the highlands. We may reasonably assume that the broad region rose in response to such additions. The vertical rise and lateral expansion of the area seem to require vertical and lateral expansion of material at depth. What mechanisms could explain this kind of structural evolution? Additions to the crust or anomalous upper mantle by phase changes in the mantle (Robertson and others, 1957; Lovering, 1958; Ringwood, 1958) and distention of the crust by igneous intrusions are perhaps the most attractive hypotheses.

A more complete discussion can be found in Thompson (1960).

The tilting of very narrow blocks, the uplifting of small domes, and the depression of small closed basins are examples of deformation that are easiest to understand if connected with igneous activity at relatively shallow depths. Possibly some structures of the dimensions of mountain ranges may be underlain by large intrusions (Thompson, 1952) and the shape of some of the basins suggests a kinship with large volcanic depressions or calderas.

GEOMORPHOLOGY

HIGH EROSION SURFACE

The most prominent erosion surface of relatively low relief in the Carson Range lies now at altitudes of 8,000–9,000 feet and gives the range a deceptively level or rounded appearance when it is viewed from a distance. This surface has been considerably modified since it was first formed and may not represent the landscape at a particular short period of time. Moreover, parts of the surface, for example at and near Big Meadows, Hunter Lake, and Tahoe Meadows, have been more recently modified by damming by flows of the Lousetown Formation and by deposition connected with glaciation. The surface was never a peneplain, for many high ridges and isolated hills or mountains projected above it.

Canyons were cut into the surface along main drainage lines before the Lousetown flows were poured out on it, as is well shown by the remnant of a flow in the canyon of Thomas Creek, 1,000 feet below the general level of the Lousetown flows. The flows are underlain by a thin sheet of rounded gravel in many places. In general the high erosion surface was presumably a pediment or pediments but was earlier and distinct from the lower pediments described in the following section.

The Truckee sediments in the northern half of the quadrangle project upward over the flanks of the Carson Range and small remnants are preserved on the flanks. But these same sediments show abundant evidence in their gravel and sand beds, which are composed of andesitic debris, of derivation from a nearby source, presumably partly in the Carson Range. While the area that is now the higher part of the Carson Range was being eroded, the sediments were evidently being deposited on what are now the flanks of the range and the basins. As the structural relief increased by relative rise of the range and (or) depression of the basins, the topographic relief remained relatively low because of erosion and deposition. After most of the present structural relief had accumulated, the high-level erosion surface reached its optimum development—its cuts across all pre-existing structure and is much less deformed than the Kate Peak and Truckee Formations. The optimum development was probably attained in late Pliocene or early Pleistocene time. The stable base level required for this optimum development may indicate a time of relative structural quiescence, or merely a dynamic equilibrium between erosion and sedimentation.

Structural movements of the same kind, continuing in Pleistocene time, resulted in warping and faulting of the high surface (with its partial cover of Lousetown flows) around the flanks of the Carson Range. The magnitude of the post-Lousetown deformation is relatively small—probably not more than one-tenth of the earlier deformation.

It must be emphasized that the high surface cannot be a correlative of any general eroson surface in the basins, as has been supposed by Louderback (1907, 1923), Reid (1911), and others. Reid (1911, p. 107) regarded the surface as "undoubtedly a continuous peneplain before the uplift of the Sierra." There is no evidence that this surface was ever reduced to so low a relief that it might be regarded as a peneplain, and far from being a continous erosion surface of regional extent, it must have been relatively local, for the basins were in many places being filled with sediments before, after, and presumably during the time that the surface was cut.

RELATION TO HIGH SURFACE IN VIRGINIA RANGE AND STEAMBOAT HILLS

The relations outlined for the northern part of the Carson Range are somewhat similar to those in the Virginia Range, where a prominent high-level surface of low relief is found as low as 5,600 feet in the western prong of the Virginia Range, rising to 7,500 feet near the crest. An erosion surface close to 5,800 feet near the crest of Steamboat Hills may correlate with the high surface of the ranges. It is evident that the intervening basins, containing Pliocene sediments, already existed at the time this surface was cut. The lower altitude of the surface in the Steamboat Hills and the Virginia Range is probably to be explained partly by original difference in erosion of terrains with different local base levels but mostly by difference in structural deformation since the formation of the surface.

In the foregoing discussion we have implied that the late Pliocene or early Pleistocene topography was much more extensively dominated by gentle slopes than it is now and that remnants of such a topographic surface can be recognized and correlated. However, we must admit that the evidence is tenuous and that if the topography of the past was anything like that of the present it is hazardous to correlate separate topographic remnants. Complete destruction of the constructional volcanic topography of the Kate Peak Formation and even exposure of the pre-Kate Peak surface show that erosion and deformation have gone on concomitantly ever since the Kate Peak volcanism, and possibly before. The discussion of younger erosion surfaces that follow in the next section emphasizes the difficulties of correlation of even the young pediment remnants because of original slope, deformation, and erosion.

PEDIMENTS

Vigorous erosion, beginning before the extrusion of at least the later Lousetown flows, cut broad pediments on the upturned Truckee beds around the margins of the Carson Range (fig. 17) and excavated deep canyons in the higher part of the range. The pediments discussed here are later and at lower altitudes than the high erosion surface considered in the previous section. Whatever established the base levels toward which the successive pediments were cut, local deformation along the northern margin of the range was slight, for even the oldest pediment surfaces there are not generally warped or faulted. West and southwest of Truckee Meadows, on the other hand, even younger pediment surfaces are offset by many small faults. Deformation in the central parts of the basins, or where the Truckee River enters the Virginia Range, are possible causes for the changing base level; other influences are the changes in climate and drainage connected with successive glaciations.

The erosional remnants of two high pediments below and distinct from the high erosion surface of the previous section are well displayed around the northern and northeastern margin of the Carson Range. The high pediment is older than the Lousetown flows in the same area and has an altitude range of about 5,000–5,800 feet. The summit of Chalk Bluff, which is on the north side of the Truckee River a mile north of Hunter Creek Reservoir, is capped with Truckee River gravels at an altitude of almost 5,000 feet (400 ft above the river); the higher pediment appears to project toward this remnant of a high river terrace. The lower pediment is roughly 250 feet below the upper and is younger than local flows of the Lousetown Formation, for pebbles from the Lousetown are included in the gravel capping. The lower pediment projects toward remnants of Truckee River terraces that lie about 160 feet above the river.

The pediment surfaces, which slope about 300 feet per mile, generally have a gravel capping 5–10 feet in thickness (fig. 18), and gravel washed down from the capping conceals older rocks in most of the lower slopes around the remnants of pediments (fig. 17). The gravel is composed of angular to rounded boulders and cobbles in a matrix of pebbles and sand. Rocks of the Kate Peak Formation are the most abundant constituents, but all the rocks upslope are represented, except that gravel derived from the Lousetown Formation is absent on the higher pediment surface, indicating the younger age of the local rocks of the Lousetown Formation. The gravel of the river terrace on Chalk Bluff is distinguished by abundant granitic and metamorphic rocks and by boulders as much as 6 feet in diameter.

Several lower erosion surfaces and subdivisions of those just described are evident in the field, but none seem to have had such widespread development as the two highest ones. On the other hand, some remnants, such as that lying immediately west of Hunter Creek, lie between the two discussed here and are graded to a

FIGURE 17.—View northwest from Hunter Lake road, showing broad, partly dissected pediment cut on Truckee Formation, which is mostly concealed by gravel washed down from the capping. Truckee beds dip about 30° to the right (northeast).

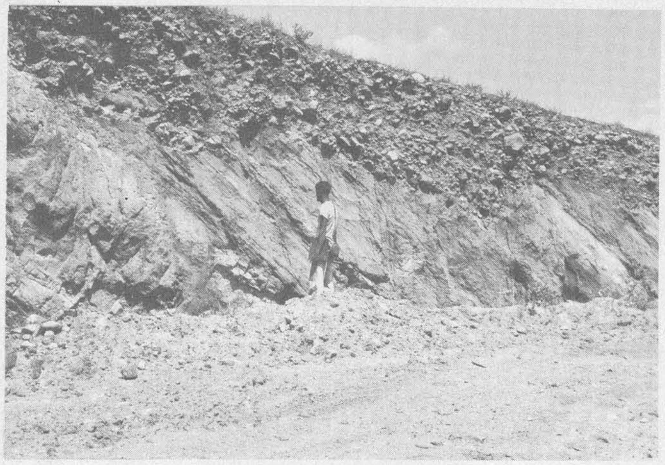

FIGURE 18.—Pediment gravels lying on the Truckee Formation with an angular unconformity of 35°, along the Steamboat Ditch near Hunter Creek.

terrace of the Truckee River between the 400- and 160-foot terraces.

The broad sloping surface southwest of Truckee Meadows is also a pediment or modified pediment cut largely on pre-Lake Lahontan deposits. The well-developed pre-Lake Lahontan soil recognized by Morrison (1961a, 1961b) formed on the cut surface. In Steamboat Hills and in the hills to the southeast are other important remnants of pediments. On the west side of the Virginia Range, in Huffaker Hills, and in Washoe Valley are still other remnants of what may have been pediments. These are described and an attempt is made to correlate them in the next section.

AGE AND CORRELATION

The Truckee River terrace toward which the lower of the two high pediments projects stands about 160 feet above the river and was identified by Blackwelder (1931, pp. 889–900) as a possible glaciofluvial terrace of Sherwin age. Results of later study by Birkeland[1] are consistent with the above and indicate that the Sherwin terraces stand 100–250 feet above the Truckee River between Lake Tahoe and Verdi. If this correlation is tentatively accepted, the lower pediment is of Sherwin age and the higher, which is correlated with the 400-foot river terrace, is possibly of McGee age in Blackwelder's terminology (1931). The following observations indicate that the 400-foot river terrace might be of glaciofluvial origin: (1) The coarseness of the gravel, which includes boulders as large as 6 feet in diameter, and (2) boulders of granitic and metamorphic rocks, derived from sources far upstream. Granitic boulders are more characteristic of glacial periods, when they are produced by frost-shattering and glacial quarrying, whereas granitic sand is generally formed under weathering conditions of nonglacial times like the present. The several erosion surfaces that are lower and less extensive than the Sherwin terrace, and which are not discussed in detail here, would then correspond with the glaciofluvial terraces of Tahoe age, which stand 60–75 feet above the present Truckee River (Blackwelder, 1931, p. 893), and those of Tioga age, which stand a maximum of 40 feet above the river (Birkeland[1]).

The two prominent high pediments at the northern end of the Carson Range can be correlated only roughly with erosional surfaces farther south. The correlation is not simple because of the rather steep slope of the surfaces, because of erosional modification in differing degree since they were first formed, and because of structural deformation since the surfaces were formed.

The broad surface lying southwest of Truckee Meadows and heading at Jones and Galena Creeks is principally a pediment. The surface slopes 300–400 feet per mile; in its upper part it is cut partly on the Kate Peak Formation, as demonstrated by outcrops between Thomas and Dry Creeks, and partly on the Truckee Formation, as shown along Dry Creek, but on the lower slopes the surface is cut on older alluvial sand and gravel of the pre-Lake Lahontan deposits. This alluvium, which is seldom coarser than pebble or cobble in roadcuts and wells, is overlain by a few feet of gravel with boulders up to 12 feet in diameter and commonly 2–4 feet. The contrast between the bouldery surface layer and the underlying sandy alluvium is strikingly shown in numerous excavations.

The altitude range of this surface, 4,600–6,000 feet, and the presence of Lousetown basaltic andesite fragments in the alluvium on which it is cut indicate that the surface correlates with the lower of the two high pediments described on the northern margin of the Carson Range and is probably of Sherwin age. Locally near Steamboat Springs, pre-Lake Lahontan deposits and individual boulders correlated with these deposits are found as much as 60 or 80 feet above the nearby pediment surface. Although local structural deformation may account for a part of this difference in altitude, cutting of the lower surface by pedimentation has been the major factor; this also accounts for the large boulders that are a residual concentration on the pediment surface.

Subsequent to the cutting of the lower pediment, erosion partly dissected the surface, and structural movements disrupted its continuity. Rare destructive cloudbursts that move 4-foot boulders down Galena Creek do occur (fig. 19), but most of the boulder gravel covering the surface accumulated as a residual concentration while the pediment surface was being cut.

[1]Birkeland, P.W., 1961, Pleistocene history of the Truckee River area north of Lake Tahoe, California: Stanford University, PhD. thesis.

FIGURE 19. House destroyed by a flash flood on Galena Creek, in the summer of 1952. Four-foot boulders covered highway.

A prominent older pediment in the eastern part of Steamboat Hills stands roughly 250 feet higher than the surface just described. The pediment in Steamboat Hills, which has an altitude range of 4,800–5,200 feet, is cut mainly on granitic rocks and is covered by basaltic andesite flows of the Lousetown Formation. Beneath the flows a few feet of pediment gravel are preserved in most places; the gravel is composed of pebbles and cobbles derived locally from Steamboat Hills. The pediment, older than the basaltic andesite flows that lie upon it, is thus similar to the higher of the two prominent pediments along the northern margin of the Carson Range. This fact does not prove that the surfaces are exactly correlative, however, for the flows in each area came from local sources and may differ in age. The altitude difference of about 250 feet between the Steamboat Hills pediment and the broad pediment southwest of Truckee Meadows strengthens the suggested correlation, however. Tentatively, then, the Steamboat Hills pediment like the higher pediment on the north flank of the Carson Range was possibly cut during the McGee Glaciation.

Another remnant of the Steamboat Hills pediment is cut on metamorphic rocks east of Pleasant Valley at an altitude of 5,200 feet and on Kate Peak Formation south of Steamboat Valley in the Virginia City quandrangle at about the same altitude. No remnants of gravels have been discovered on these surfaces, where a protecting cover of lava is lacking.

Farther north along the west side of the Virginia Range as far as Huffaker Hills there are no well-preserved remnants of a comparable surface. The summits of Huffaker Hills that have altitudes of 4,800–5,000 feet may once have been part of the surface. On the west slope of the Virginia Range adjacent to Huffaker Hills, a bench at an altitude of about 4,900 feet was interpreted as a questionable fault block (Thompson, 1956) but may instead be a remnant of the erosion surface. A higher bench in the same area lies at an altitude of 5,300 feet; it might also be erosional and its presence illustrates the difficulty and uncertainty of detailed correlation of surfaces.

To the south of Steamboat Hills, the three valleys, Washoe, Pleasant, and Steamboat, formed a single broader valley during the earlier period of extensive pedimentation. Inspection of the 5,200- and 5,400-foot contours around Pleasant and Washoe Valleys gives a clue to the former topography. A flat-topped hill northwest of Little Washoe Lake, on the divide between Pleasant and Washoe Valleys, has an altitude of 5,240 feet and is nearly continuous with the broad erosion surface extending westward. Another remnant lies on the hill east of Little Washoe Lake at an altitude of 5,240 feet, and farther east are alluvial gravels that, although partly reworked and largely covered by windblown sand, may be as old as the higher pediment surface. Still other gravel deposits including a remnant on the west side of Pleasant Valley at the main adit of the Union lead mine and remnants between Washoe and Pleasant Valleys lie a little lower at about 5,000 feet and may be correlative with the lower pediment. Structural movements later than the erosion surface have certainly complicated the relations in this area, for deformation is required to maintain the lakes in Washoe Valley in spite of erosion and deposition. The entrenched outlet of Washoe Valley is perhaps a result of both superposition from an old alluviated surface and of eastward tilting of the outlet area.

On the east side of Washoe Valley two areas of older alluvium cap hills, one at an altitude of 5,200 feet and the other at 5,480 feet. The alluvium is obviously a product of an earlier cycle of erosion and may also be correlative with one of the broad pediments.

Detailed relations revealed from drill holes at Steamboat Springs demonstrate complexities in the geomorphic history of the region, some of which have been mentioned previously. At least near Steamboat Hills the cutting of the high erosion surface near the crest of the hills (close to the 5,800-ft contour) was followed by very deep erosion to a base level presently at an altitude at least as low as 4,340 feet, which is nearly 270 feet below present local base level and 30 feet below the Truckee River at its entrance into the Virginia Range. Valleys formed at that time were then alluviated to depths of at least 450 feet (200 ft above present local base level) prior to the cutting of the higher pediment surface upon which basaltic andesite flows were erupted.

After the basaltic andesite was erupted, erosion proceeded again, this time down to a base level presently close to 4,500 feet in altitude, which is about 100 feet below present base level. This period of erosion was followed by deposition of the pre-Lake Lahontan deposits of probable Sherwin age; at least locally the upper 60–80 feet of these deposits were removed by pedimentation that preceded development of the pre-Lake Lahontan soil.

The evidence for the postulated early deep erosion and a later less deep erosion is provided by drill cores and cuttings from drill holes in the thermal area. Alluvium (in many places altered or cemented by hydrothermal action) was found as deep as 4,340 feet in altitude or 270 feet below present local base level. A careful investigation of hundreds of pebbles and dozens of thin sections has failed to find any fragments in the deep alluvium that are identifiable as the Steamboat basaltic andesite phase of the Lousetown Formation; this is in spite of the fact that such lava is now directly upslope from all of these drill holes, and its texture is so characteristic that it is distinguishable from all the other volcanic rocks of the region, generally even where completely replaced by hydrothermal minerals. The deep alluvium without basaltic andesite fragments is assigned to the alluvium of Pliocene or Pleistocene age. In contrast, the shallower alluvium that does contain basaltic andesite pebbles is assigned to the pre-Lake Lahontan deposits of probable Sherwin age.

The absence of fragments of Steamboat basaltic andesite in the deep alluvium can be explained reasonably only if the alluvium is older than the flows. In the sequence of events, deep erosion of the highest surface was followed by extensive alluviation. A stillstand or steady state was then attained when pedimentation cut back into bedrock walls of the old alluvium-filled valleys. The first recognizable hot-spring activity then commenced, and this was followed by eruption of basaltic andesite flows on the pediment surface and perhaps on parts of the deep alluvial fill. The next period of erosion then removed alluvium with relative ease, but bedrock margins of the older valleys were more resistant.

The preceding explanation accounts for the fact that the high pediment surface occurs as a narrow fringe around the present Steamboat Valley. The lower valley margins are largely the exhumed margins of the old deep valley, and the narrow fringing pediment is thereby closely related to the form of the present valley.

SUMMARY

Former landscapes very different from the present have evolved by continuous modification at variable rates. Relics of certain stages in the evolution of the landscapes are recognizable and partly decipherable, and three of these, the low-relief erosion surface high in the ranges and two broad pediment surfaces around the margins of the ranges, are discussed in some detail. The high surface is no older than Pliocene and probably reached its optimum development in late Pliocene or early Pleistocene time, but it may not represent the landscape at a particular short period of time. The two best developed pediments were possibly cut during the McGee and Sherwin Glaciations.

The high surface is well developed in various parts of the Carson Range at altitudes of 8,000–9,000 feet, in the Virginia Range at altitudes of about 5,600–7,500 feet, and near the crest of Steamboat Hills at about 5,800 feet. Although some geologists have assumed that the surface is older than the basins and ranges, evidence found in the present study indicates instead that basins and low ancestral ranges were already formed when the surface was being cut. Differences in altitude of the high erosion surface in the two ranges are then attributable to original differences in what may have been a compound pediment or rolling upland and to later structural deformation. Because no completely reliable criteria exist for correlation of the separate topographic remnants, the erosional history may be considerably more complicated than we imply. A still higher and older erosion surface may be represented on the summits above 9,000 feet, but no attempt is made here to reconstruct it.

Below the prominent high surface are numerous remnants of erosion surfaces that defy regional correlation, and below these are two main systems of broad pediments. Several lower and narrower pediments are so obviously related to terraces along present valleys that they need not be discussed. The higher of the two high pediments on the northern margin of the Carson Range was cut about the same time as the prominent lava-capped pediment in Steamboat Hills. Other remnants to the south and along the Virginia Range show that the higher pediment was widely developed and during its development the chain of valleys from Washoe Valley to the Truckee River was broader than at present, was filled with alluvium to heights of 200 feet above the present level of Truckee Meadows, and probably had an axial gradient of less than 50 feet per mile, compared to a slope of about 300 feet per mile of the pediment toward the valleys. Local evidence at Steamboat Springs indicates that the cutting of this high pediment was both preceded and followed by cycles of deep erosion, alluviation, and pedimentation by streams. The first cycle was earlier than basaltic andesite flows of the Lousetown Formation and the second cycle was later and probably Sherwin in age. The lower of the two high pediments in the northern part of the Carson Range was then cut to the same general base level as the broad boulder-strewn surface southwest of Truckee Meadows. A widely developed pre-Lahontan soil then formed on the pediment surface, which was then displaced by faults with as much as 75 feet of vertical movement. Events contemporaneous with and later than Tahoe and Tiogan Glaciations have had relatively minor geomorphic effects.

MINERAL DEPOSITS

METALLIC DEPOSITS

Because of the proximity of the Comstock Lode district with its great bonanzas of silver and gold ore, prospecting in the Mount Rose quadrangle was carried on vigorously in the latter part of the 19th century. Hardly an intrusive contact of the granitic rocks or a patch of altered rock has escaped the prospector's pick, but it is generally difficult now to determine why pits were dug where they are. The most persistent efforts at mining and prospecting have been at the Union lead mine (near the mouth of Galena Creek), at the gold prospects on the east side of Pleasant Valley, at the lead-silver prospects in Steamboat Hills, at the gold and tungsten prospects 2 miles southeast of Little Washoe Lake, at mercury and gold prospects around Steamboat Springs, and at the Wheeler Ranch mercury prospect on Evans Creek near the northeast corner of sec. 4, T. 18 N., R. 19 E.

PRE-TERTIARY DEPOSITS

The mineralization, with the exception of that around Steamboat Springs and the Wheeler Ranch mercury prospect, is pre-Tertiary in age. The commonest type of mineral occurrence is in quartz veins along faults in either granitic or metamorphic rocks near the intrusive contacts, but some deposits, like that of the Union lead mine, are along faults where no intrusive rocks are exposed. Brown and yellow iron oxide and a green stain of copper minerals generally mark the outcrops. Below the oxidized outcrops, pyrite and, less commonly, arsenopyrite are the two most abundant sulfides, but galena, chalcopyrite, sphalerite, and other sulfides are occasionally seen. Gold and silver associated with the more abundant sulfides stimulate prospecting. In the prospects on the east side of Pleasant Valley one of the main veins, which produced some gold from arsenopyrite ore in 1907 (Overton, 1947, p. 66), strikes northeast and dips 55° west (pl. 1). In the prospects southeast of Little Washoe Lake a principal vein strikes eastward and dips 55° south (NW. cor. sec. 29 and NE. cor. sec. 30, T. 17 N., R. 20 E.). This vein is explored in the workings of the Denver mine, in which the intrusive contact and numerous granitic dikes near the contact can also be seen.

Small quantities of the tungsten mineral scheelite are widely disseminated in contact deposits where granitic rocks have intruded and metamorphosed limestone. The largest zone of this kind is southeast of Little Washoe Lake near the eastern edge of the quadrangle. Contact deposits of magnetite, although developed on the east side of the Virginia Range, are unknown in the Mount Rose quadrangle.

UNION LEAD MINE

The Union lead, or Commonwealth, mine has produced zinc-lead-silver ore intermittently since 1860. During World War II, when ore was shipped to smelters near Salt Lake City, the production was 14,624 tons with a gross value of about $232,000 (Overton, 1947, p. 65; Geehan, 1950).

A mineralized fault zone about 30 feet wide strikes N. 60° E. and dips 55° S. The main fault, which lies along the footwall of the mineralized zone, is generally marked by a black clayey gouge about 6 inches thick. In the hanging wall above the main fault are many subparallel faults, which are marked by silicified breccia zones instead of gouge and cross-fractures, along which the better ore is concentrated.

The rock cut by these faults is chiefly a massive hornfels of pre-Tertiary age, which in some places shows a clastic texture. A metaconglomerate exposed at the surface is composed of well-rounded cobbles of basalt or andesite as large as 6 inches in diameter. The bulk of the hornfels, however, was probably tuffaceous sandstone and shale before it was metamorphosed. An irregular dike of hornblende andesite of the Kate Peak Formation was met in the underground workings and a large dike which is probably connected with that underground is exposed on the surface. The dike rock is distinctive in appearance; needlelike hornblende crystals about 3 mm long and equidimensional plagioclase crystals about 2 mm in diameter are embedded in a light-gray groundmass. The dike is younger than the ore but is locally bleached by later alteration possibly related to oxidation of sulfides. The youngest rock in the mine is near the main portal, where the adit is cut through a boulder gravel resembling the modern gravel in Galena Creek but lying far above the present creek. These old gravels are probably roughly correlative with the gravels of the broad pediments. Interbedded with the boulder gravel about 100 feet in from the portal is a crudely bedded taluslike rubble made up entirely of the hornblende-andesite dike rock. A young fault about 250 feet in from the main portal separates the gravel and talus from metamorphic rocks.

The chief ore minerals below the zone of oxidation, which in some places extends downward to the No. 1 level, 175 feet above the main portal, are galena and sphalerite, the latter including both ordinary sphalerite and the dark iron-rich variety. The other sulfides are chalcopyrite, pyrite, and arsenopyrite. The gangue minerals are quartz and a small amount of calcite. Economically important amounts of silver are associated with the galena. The higher grade ore occurs in irregular shoots controlled by small faults and fractures in the hanging wall of the main fault. A bend of the main fault convex to the north may have localized the ore shoot explored by the main winze.

The U.S. Bureau of Mines conducted a drilling project at the mine in 1947 to test for horizontal and vertical extensions of the ore body and to prospect for parallel ore bodies. Extensions of the ore zone were established, but no parallel ore-bearing structures were found (Geehan, 1950). Future prospecting might well include a search for intersections of the mineral zone with limestone lenses, granitic intrusive rock, or other concealed rocks of properties contrasting with those of the hornfels.

TERTIARY DEPOSITS

The large areas of bleached Tertiary rocks in the western and especially in the northern part of the quadrangle have naturally invited the attention of prospectors, but in spite of a considerable expenditure of effort no substantial production of metals has resulted. The numerous shallow prospect pits were presumably dug in the search for gold and mercury, but ore minerals were observed only at the Wheeeler Ranch mercury prospect on Evans Creek. The only common sulfide in the bleached areas is pyrite; the abundant clay and silica result from acid weathering of the pyritic rock.

The Wheeler Ranch mercury prospect, which is on Evans Creek near the northeast corner of sec. 4, T. 18 N., R. 19 E., is described by Bailey and Pheonix (1944, p. 193), from whom the following information is taken. The deposit was discovered in 1875 and has been worked several times since then, most recently in 1940. In the workings, which have a length of about 260 feet, cinnabar occurs as veinlets in northward-trending faults and as disseminations in altered wall rocks. Pyrite, jarosite, gypsum, chalcedony, and reportedly native mercury are associated with the cinnabar.

Little Valley, on the east flank of the Carson Range, has produced gold from placers beneath the Tertiary volcanic rocks, but the productive ground all lies a little to the south of the Mount Rose quadrangle. The placers are described by Reid (1911) and production figures are given by Overton (1947).

QUATERNARY DEPOSITS

In most of the world's mining districts, deposition of ore minerals ceased long ago, but around Steamboat Springs the ore-forming processes have continued to the present time. Indeed the episodes of hydrothermal activity in a larger area are not finished and dead, if we may judge from the hot water and hydrogen sulfide in the Comstock mines, the numerous thermal springs in the general region, and the even more numerous occurrences of warm water in wells.

All of the metallic deposits known to be as young as Quaternary, however, are in the Steamboat Springs district and consist mainly of several occurrences of cinnabar which have yielded a small amount of mercury. The locality is world famous among geologists because mercury, antimony, and iron sulfides, gold and silver in unknown forms, and other metalliferous minerals have been deposited from the hot springs and have been studied to learn how ore deposits are formed. The results of these studies and the history of mineral production are summarized by White (1955, 1957) and Bailey and Phoenix (1944). References to earlier work are given in White's papers.

A mercury and sulfur deposit was discovered in 1875 at about the location of the present "silica pit" which is nearly a mile west of the active hot springs, in the SW¼NE¼ sec. 32, T. 18 N., R. 20 E. Both mercury and sulfur were produced in the early years, and several attempts at mining were made since then. In 1930 the ground was worked for silica, clay, and mercury in a large opencut (see section on nonmetallic deposits). Steam and hydrogen sulfide are rising through fractured granodiorite and basaltic andesite in the bottom of the opencut, and these rocks, along with a layer of pediment gravel beneath the lava are bleached white by acid resulting from the atmospheric oxidation of the hydrogen sulfide. The cinnabar occurs near the surface as thin films in the fractured bleached rock. Both the cinnabar and sulfur were presumably deposited from rising vapors. Farther west and north in section 32, altered ground has been explored by many pits, trenches, and short adits.

In the old (chalcedonic) sinter in the northeastern part of section 32 and in the adjacent section 29 to the north, cinnabar, stibnite, and gold and silver in small amounts have been found at several places. The younger (opaline) sinter near the highway (in the Virginia City quadrangle) generally contains no visible cinnabar at the surface, but a short adit driven below the crest of the sinter mass disclosed small amounts of cinnabar and stibnite in chalcedonic sinter. Hot water and sulfurous gases hamper deep prospecting in the Steamboat Springs district.

NONMETALLIC DEPOSITS

A variety of nonmetallic materials useful in construction are readily available in the Mount Rose quadrangle and some have been exploited for local needs. Alluvial sand and gravel are abundant; the pre-Lake Lahontan gravels, since they are deeply weathered, should be avoided where aggregate of highest quality is desired.

Coarse sand at the foot of steep slopes on crushed or disintegrated granitic rock is commonly used for road material and is available in large quantities at many places. Windblown sand, uniformly well sized, could be obtained in large quantities from dunes on the eastern side of Washoe Lake and from pockets on the slopes east of Washoe Valley.

Pumiceous rhyolite in a dome in Steamboat Hills is mostly of rather high density, and moreover similar material is more readily available northeast of Steamboat Hills in the Virginia City quadrangle. Another material commonly used in lightweight aggregate, basaltic cinder, occurs in the cone near Bronco Creek in the western part of the quadrangle, but this area is steep and inaccessible compared with other cinder cones in the region near the Mount Rose quadrangle.

Silica and clay, generally intermixed, are available in large quantities in the areas of bleached rock. Clay from a bleached area in the northern part of Steamboat Hills (NW¼ sec. 32, T. 18 N., R. 20 E.) has been mined by the Reno Press Brick Co. and used in their products. Half a mile to the southeast at the "silica pit" an attempt was made about 1930 to obtain pure silica for the glass industry from bleached granodiorite, basaltic andesite, and alluvium. This snow-white altered rock is nearly pure silica, with a trace of cinnabar and native sulfur, and grades into kaolinite clay and unaltered rock. The silica is a mixture of porous opal and residual quartz from the granodiorite. The ground in the bottom of the pit is warm, moist, and acid; alteration by rising H_2S gas that combines with atmospheric oxygen is still in progress.

THERMAL WATERS

Thermal waters underlying a large area in the northeastern part of Steamboat Hills and extending eastward to the discharge area at Steamboat Springs have been used locally for mineral baths, a swimming pool, and local heating of buildings. A very large supply of thermal energy is continually dissipated to the streams and the atmosphere. The heat flow is estimated as 7×10^6 cal per sec over an area of 1.9 square miles (White, 1957); this is equivalent to the heat from about 100 tons of coal or 500 barrels of oil per day. If the difficulties of corrosion of pipes and clogging by mineral precipitation from the highly mineralized waters can be overcome economically, some of this thermal energy will be available for heating purposes or geothermal power. Exploratory drilling by the U.S. Geological Survey in 1950 and by private interests in 1959–60 found temperatures as high as 186°C in hot water that yields saturated steam upon release of pressure (White, 1961). The outlook is not favorable for supersaturated steam.

Thermal waters are also available and are utilized locally at Bowers Mansion on the west side of Washoe Valley and at Moana Hot Springs on Moana Lane, south of Reno. Hot water is reported at 1,200 feet in a deep well southwest of Reno in SE¼ sec. 21, T. 19 N., R. 19 E. (Anderson, 1909, p. 487). This locality is 3 miles east and a little south of a hot spring at Lawton in the Reno quadrangle. Other shallow water wells at various places in the quadrangle contain warm water.

REFERENCES

Anderson. C. A., 1933, Tuscan formation of northern California, with a discussion concerning the origin of volcanic breccias: California Univ., Dept. Geol. Sci. Bull., v. 23, no. 7, p. 215–276.

Anderson, Robert, 1909, Geology and oil prospects of the Reno region, Nevada: U.S. Geol. Survey Bull. 381, p. 475–493.

Axelrod, D. I., 1949, Eocene and Oligocene formations in the western Great Basin [abs.]: Geol. Soc. America Bull., v. 60, no. 12, pt. 2, p. 1935–1936.

———1958, The Pliocene Verdi flora of western Nevada: California Univ., Dept. Geol. Sci. Bull., v. 34, no. 2, p. 91–159.

Bailey, E. H., and Phoenix, D. A., 1944, Quicksilver deposits in Nevada: Nevada Univ. Bull., v. 38, no. 5, Geol. and Mining ser. no. 41, 206 p.

Becker, G. F., 1882, Geology of the Comstock Lode and the Washoe district: U.S. Geol. Survey Mon. 3, 422 p.

Berg, J. W., Jr., Cook, K. L., and Narans, H. D., Jr., 1959, Seismic studies of crustal structure in the eastern Basin and Range province [abs.]: Geol. Soc. America Bull., v. 70, no. 12, pt. 2, p. 1709.

Billings, W. D., 1950, Vegetation and plant growth as affected by chemically altered rocks in the western Great Basin: Ecology, v. 31, no. 1, p. 62–74.

Blackwelder, Eliot, 1931, Pleistocene glaciation of the Sierra Nevada and Basin Ranges: Geol. Soc. America Bull., v. 42, no. 4, p. 865–922.

Bragg, William, 1948, Concerning the nature of things: New York, Dover Publications, 231 p.

Brannock, W. W., Fix, P. F., Gianella, V. P.. and White, D. E., 1948, Preliminary geochemical results at Steamboat Springs, Nevada: Am. Geophys. Union Trans., v. 29, no. 2, p. 211–226.

Calkins, F. C., 1944, Outline of the geology of the Comstock Lode district, Nevada: U.S. Geol. Survey, 35 p.

Calkins, F. C., and Thayer, T. P., 1945, Preliminary geologic map of the Comstock Lode district, Nevada; U.S. Geol. Survey, scale 1: 24,000.

Coats, R. R., 1936, Intrusive domes of the Washoe district, Nevada: California Univ., Dept. Geol. Sci. Bull., v. 24, no. 4, p. 71–84.

———1940. Propylitization and related types of alteration on the Comstock Lode: Econ. Geology, v. 35, no. 1, p. 1–16.

Craig, Harmon, Boato, G., and White, D. E., 1956, The isotopic geochemistry of thermal waters, (Chap.) 5 *of* Nuclear processes in geologic settings: Natl. Research Council, Comm. Nuclear Sci., Nuclear Sci. Ser. Rept. 19, p. 29–44.

Curtis, G. H., 1954, Mode of origin of pyroclastic debris in the Mehrten formation of the Sierra Nevada: California Univ., Dept. Geol. Sci. Bull., v. 29, no. 9, p. 453–502.

Curtis, G. H., Evernden, J. F., and Lipson, J., 1958, Age determination of some granitic rocks in California by the potassium-argon method: California Div. Mines Spec. Rept. 54, 16 p.

Durrell, Cordell, 1944, Andesite breccia dikes near Blairsden, California: Geol. Soc. America Bull., v. 55, no. 3, p. 255–272.

Geehan, R. W., 1950, Investigation of Union zinc-lead mine, Washoe County, Nevada: U.S. Bur. Mines Rept. Inv. 4623, 10 p.

Gianella, V. P., 1936, Geology of the Silver City district and the southern portion of the Comstock Lode, Nevada: Nevada Univ. Bull., v. 30, no. 9, Geol. and Mining Ser. no. 29, 105 p.

King, Clarence, 1870, The Comstock Lode: U.S. Geol. Explor. 40th Parallel (King) v. 3, p. 11–96, with atlas.

——— 1878, Systematic geology: U.S. Geol. Explor. 40th Parallel (King), v. 1, 803 p.

Larsen, E. S., Jr., Gottfried, David, Jaffee, Howard, and Waring, C. L., 1954, Age of the Southern California, Sierra Nevada, and Idaho batholiths [abs.]: Geol. Soc. America Bull., v. 65, no. 12, pt. 2, p. 1277.

Lawrence, D. B., 1958, Glaciers and vegetation in southeastern Alaska: Am. Scientist, v. 46, no. 2, p. 89–122.

Lindgren, Waldemar, 1897, Description of the Truckee quadrangle, California: U.S. Geol. Survey Geol. Atlas, Folio 39.

——— 1911, The Tertiary gravels of the Sierra Nevada of California: U.S. Geol. Survey Prof. Paper 73, 226 p.

Louderback, G. D., 1903, Basin Range structure in the Humboldt region: Geol. Soc. America Bull., v. 15, p. 289–346.

——— 1907, General geologic feature of the Truckee region east of the Sierra Nevada: Geol. Soc. America Bull., v. 18, p. 662–669.

——— 1923, Basin Range structure in the Great Basin: California Univ., Dept. Geol. Sci. Bull. v. 14, no. 10, p. 329–376.

——— 1924, Period of scarp production in the Great Basin: California Univ., Dept. Geol. Sci. Bull. v. 15, no. 1, p. 1–44.

Lovering, J. F., 1958, The nature of the Mohorovicic discontinuity: Am. Geophys. Union Trans., v. 39, no. 5, p. 947–955.

Matthes, F. E., 1942, Glaciers, *in* Meinzer, O. E., ed., Hydrology, v. 9 *of* Physics of the earth: New York, McGraw-Hill Book Co., p. 149–219.

Morrison, R. B., 1952, Stratigraphy of Lake Lahontan and associated Quaternary deposits in the Carson Desert area, near Fallon, Nevada [abs.]: Geol. Soc. America Bull., v. 63, no. 12, pt. 2, p. 1367.

——— 1961a, Lake Lahontan stratigraphy and history in the Carson Desert (Fallon) area, Nevada: Art. 329 *in* U.S. Geol. Survey Prof. Paper 424–D, p. D111–D114.

——— 1961b, Correlation of the deposits of Lakes Lahontan and Bonneville and the glacial sequences of the Sierra Nevada and Wasatch Mountains, California, Nevada, and Utah: Art. 332 *in* U.S. Geol. Survey Prof. Paper 424–D, p. D122–D124.

Naboko, S. I., and Piip, B. I., 1961, Contemporary metamorphism of volcanic rocks in the region of the Pauzhetsk hot springs [Kamchatka]: Akad. Nauk SSSR Lab. Vulkanologii Trudy, vyp. 19, p. 99–114.

Nockolds, S. R., 1954, Average chemical compositions of some igneous rocks: Geol. Soc. America Bull., v. 65, no. 10, p. 1007–1032.

Nolan, T. B., 1943, The Basin and Range province in Utah, Nevada, and California: U.S. Geol. Survey Prof. Paper 197–D, p. 141–196.

Overton, T. D., 1947, Mineral resources of Douglas, Ormsby, and Washoe Counties [Nevada]: Nevada Univ. Bull., v. 41, no. 9, Geol. and Mining Ser. No. 46, 91 p.

Pakiser, L. C., Press, Frank, and Kane, M. F., 1960, Geophysical investigation of Mono Basin, California: Geol. Soc. America Bull., v. 71, no. 4, p. 415–447.

Pellizzer, Renato, 1957, Trasformazioni sperimentali alle condizioni pneumatolitiche e idrotermali di rocce ofiolitiche Appenniniche (Experimental changes of ophiolitic rocks from the Apennines by pneumatolytic and hydrothermal conditions): Rendiconti Soc. Mineral. Ital., v. 13, p. 295–324.

Powell, J. W., 1876, Report on the geology of the eastern portion of the Uinta Mountains and a region of country adjacent thereto: U.S. Geol. and Geog. Survey Terr. (Powell), 2d div., 218 p.

Press, Frank, 1958, Velocity distribution in the crust [abs.]: Am. Geophys. Union Trans., v. 39, no. 3, p. 528.

Reid, J. A., 1905, The structure and genesis of the Comstock Lode: California Univ., Dept. Geology Bull., v. 4, p. 177–199.

———, 1911, The geomorphogeny of the Sierra Nevada northeast of Lake Tahoe [Nevada]: California Univ., Dept. Geology Bull., v. 6, no. 5, p. 89–161.

Richmond, G. M., 1961, New evidence of the age of Lake Bonneville from the moraines in Little Cottonwood Canyon, Utah in Art. 334 *in* U.S. Geol. Survey Prof. Paper 424–D, p. D127–D128.

Ringwood, A. E., 1958, Consequences of the olivine-spinel transition, [pt.] 3 *of* The constitution of the mantle: Geochim. et Cosmochim. Acta, v. 15, no. 3, p. 195–212.

Robertson, E. C., Birch, A. F., and MacDonald, G. J. F., 1957, Experimental determination of jadeite stability relations to 25,000 bars: Am. Jour. Sci., v. 255, no. 2, p. 115–137.

Russell, I. C., 1885, Geological history of Lake Lahontan, a quaternary lake of northwestern Nevada: U.S. Geol. Survey Mon. 11, 288 p.

Sigvaldason, G. E., and White, D. E., 1961, Hydrothermal alteration of rocks in two drill holes at Steamboat Springs, Washoe County, Nevada: Art. 331 *in* U.S. Geol. Survey Prof. Paper 424–D, p. D116–D122.

Smith, D. T., 1912, Vein systems of the Comstock: Eng. Mining Jour., v. 94, p. 895–896.

Thayer, T. P., 1937, Petrology of later Tertiary and Quaternary rocks of the north-central Cascade Mountains in Oregon, with notes on similar rocks in western Nevada: Geol. Soc. America Bull., v. 48, no. 11, p. 1611–1651.

Thompson, G. A., 1952, Basin and Range structure south of Reno, Nevada [abs.]: Geol. Soc. America Bull., v. 63, no. 12, pt. 2, p. 1303–1304.

——— 1956, Geology of the Virginia City quadrangle, Nevada: U.S. Geol. Survey Bull. 1042–C, p. 45–77.

——— 1959, Gravity measurements between Hazen and Austin, Nevada—a study of Basin-Range structure: Jour. Geophys. Research, v. 64, no. 2, p. 217–229.

——— 1960, Problem of late Cenozoic structure of the Basin Ranges: Internat. Geol. Cong., 21st, Copenhagen 1960, pt. 18, p. 62–68.

Thompson, G. A., and Sandberg, C. H., 1958, Structural significance of gravity surveys in the Virginia City-Mount Rose area, Nevada and California: Geol. Soc. America Bull., v. 69, no. 10, p. 1269–1282.

Thompson, G. A., and Talwani, Manik, 1959, Crustal section across California and Sierra Nevada [abs.]: Geol. Soc. America Bull., v. 70, no. 12, pt. 2, p. 1688.

White, D. E., 1955, Thermal springs and epithermal ore deposits: Econ. Geology, 50th Anniversary Volume, 1905–1955, p. 99–154.

——— 1957, Thermal waters of volcanic origin: Geol. Soc. America Bull., v. 68, no. 12, pt. 1, p. 1637–1658.

——— 1961, Preliminary evaluation of geothermal areas by geochemistry, geology, and shallow drilling: United Nations Conf. New Sources of Energy, Rome, Italy, 1961 (preprint), 12 p.

White, D. E., Brannock, W. W., and Murata, K. J., 1956, Silica in hot-spring waters: Geochim. et Cosmochim. Acta, v. 10, nos. 1–2, p. 27–59.

Whitten, C. A., 1957, Geodetic measurements in the Dixie Valley area: Seismol. Soc. America Bull., v. 47, no. 4, p. 321–325.

Williams, Howel, 1941, Calderas and their origin: California Univ., Dept. Geol. Sci. Bull., v. 25, no. 6, p. 239–346.

Wilshire, H. G., 1957, Propylitization of Tertiary volcanic rocks near Ebbetts Pass, Alpine County, California: California Univ., Dept. Geol. Sci. Bull., v. 32, no. 4, p. 243–272.

INDEX

[Major references are in italic]

U.S. GOVERNMENT PRINTING OFFICE: 1964 O—721-882