Duration of Hydrothermal Activity at Steamboat Springs, Nevada, From Ages of Spatially Associated Volcanic Rocks

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GEOLOGY AND GEOCHEMISTRY OF THE STEAMBOAT SPRINGS AREA, NEVADA

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CONTENTS

ILLUSTRATIONS

TABLES

DURATION OF HYDROTHERMAL ACTIVITY AT STEAMBOAT SPRINGS, NEVADA, FROM AGES OF SPATIALLY ASSOCIATED VOLCANIC ROCKS

By M. L. Silberman, D. E. White, T. E. C. Keith, and R. D. Dockter

ABSTRACT

Steamboat Springs is a presently active equivalent of epithermal gold-silver ore-forming systems. Hot-spring sinter deposits contain small amounts of gold, silver, mercury, antimony, and arsenic. Hot-spring activity probably started before extrusion of the basaltic andesite of Steamboat Springs. Old sinter from the Steamboat Springs system occurs in gravels above and below the basaltic andesite. Intense hydrothermal alteration, including almost complete replacement by hydrothermal potassium-feldspar, has affected the basaltic andesite. Three plagioclase separates of differing potassium content from fresh basaltic andesite yielded potassium-argon ages of 2.52–2.55 m.y. Basaltic andesite almost completely replaced by potassium-feldspar yielded an age of 1.1 m.y.

The thermal area lies approximately on a line connecting four rhyolite domes, the largest of which is 3 km southwest of the springs and the others are to the northeast. Several domes occupy vents from which tephra of pre- and post-basaltic andesite age was erupted, as indicated by pumice blocks up to 5 cm diameter in gravel underlying the andesite, and isolated pumice fragments lying on the andesite. The source of energy for the thermal convection system is probably a large rhyolitic magma chamber that supplied the pumice and from which the rhyolite domes were emplaced. Sanidine and obsidian from four of the rhyolite domes yielded potassium-argon ages of 1.15–1.52 m.y. and obsidian from one of the northeastern domes yielded apparent ages of 2.97 and 3.03 m.y.

The data indicate that hydrothermal activity has occurred at Steamboat Springs, possibly intermittently, for more than 2½ m.y. These data agree with other radiogenic age studies indicating 1- and 2-m.y. lifetimes for the hydrothermal systems that generate epithermal gold-silver deposits.

INTRODUCTION

The timespan and continuity of hydrothermal systems are important to establish for two reasons. First, the geothermal energy that may be recoverable from active systems has been related to the age of hot-spring activity and its volcanic heat source, through Smith and Shaw's (1975) magmatic model. Second, a major group of active hydrothermal systems, closely associated with large rhyolitic systems, is similar in water composition, metal content, subsurface temperature, and hydrothermal alteration to epithermal gold-silver-depositing systems (White, 1955, 1974). Because some active hydrothermal systems, including Steamboat Springs, seem to be the present-day equivalents of the ore-forming systems, they help to enhance our understanding of ore-forming processes. Few direct data on the limits of age, duration, and continuity of activity are yet available either for active hot-spring systems or for epithermal ore-forming systems of the past.

Our study provides much age data for a notable active hot-spring system. Steamboat Springs and the surrounding region have been thoroughly studied by geologic, geochemical, and hydrologic methods (Thompson and White, 1964; White and others, 1964; White, 1968). The hot-spring activity has occurred over a relatively long period of time. Uplift and erosion of the thermal area alternated with burial by volcanic rocks, alluvium, and hot-spring deposits, thus preserving much geologic evidence that permits reconstruction of its history. Potassium-argon ages, now available for many of the fresh and altered young volcanic rocks in and near the thermal area, permit precise estimates to be made for the total time span of hydrothermal activity. However, we do not yet know whether the activity has been continuous or intermittent through this timespan.

REGIONAL GEOLOGIC RELATIONS

The Steamboat Springs area is located near the northeast end of Steamboat Hills, a small northeast-trending range that lies between the north-trending Carson Range on the west and the Virginia Range on the east (figs. 1 and 2). The springs are noted for transport and deposition of mercury, antimony, gold, silver, and other metals and gangue minerals (White and others, 1964).

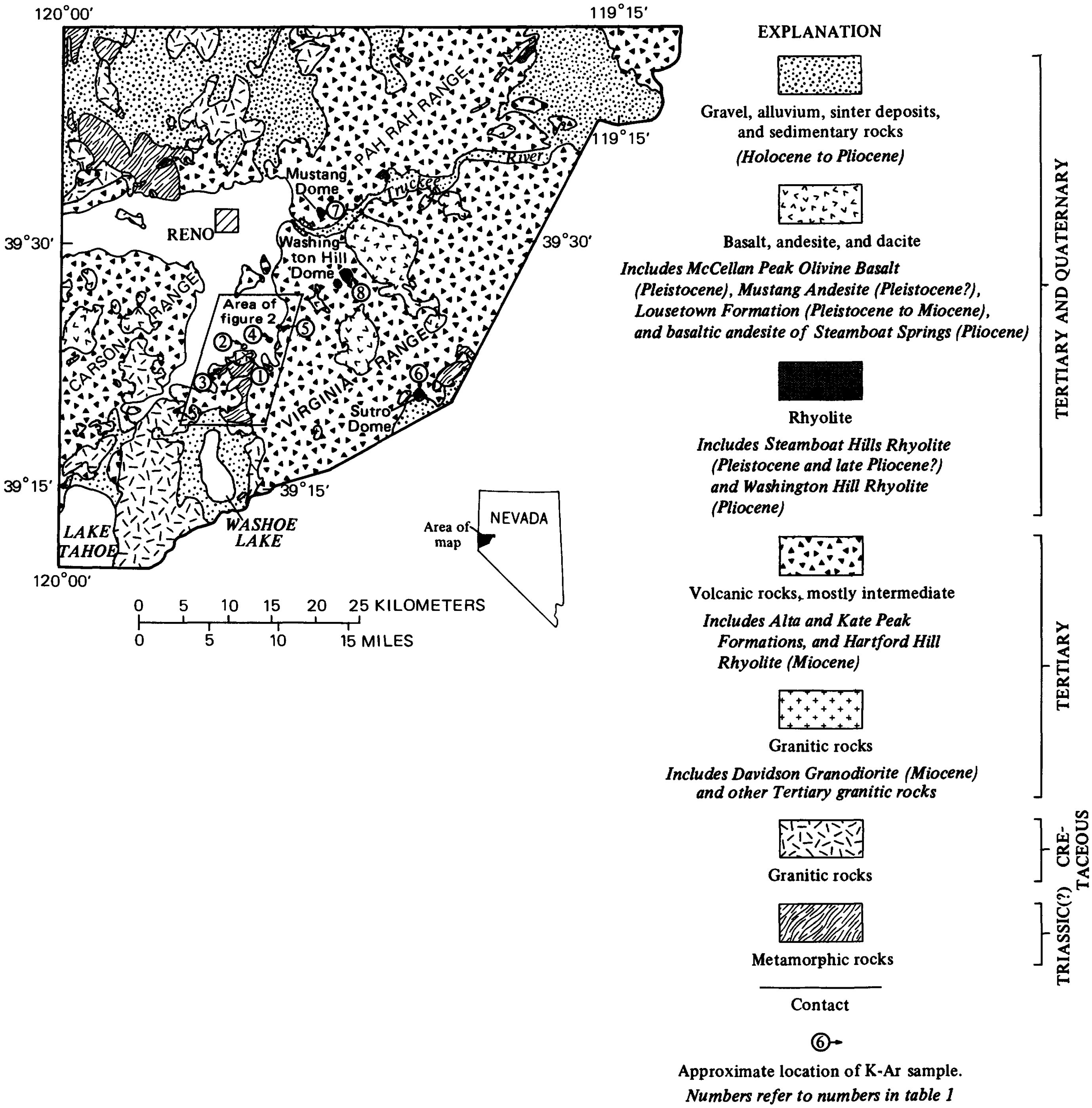

FIGURE 1.—Generalized geologic map of Storey and southern Washoe Counties, Nevada. Geology modified from Bonham (1969).

In the region surrounding Steamboat Springs (fig. 2), pre-Tertiary Triassic (?) metamorphic rocks and Cretaceous granitic rocks that form the basement are overlain by a sequence of middle Tertiary to Pleistocene volcanic rocks that has been discussed in detail by Thompson and White (1964) and Bonham (1969). Rock units considered in this report and generalized on figure 2 include the Alta and the Kate Peak Formations of Miocene age, the Lousetown formation, including local basaltic andesite of Steamboat Springs of Pliocene age, the Steamboat Hills Rhyolite of late Pliocene (?) and Pleistocene age, and alluvium of late Pliocene to Holocene age. The geologic relations in the vicinity of Steamboat Springs are summarized here from earlier papers in this series (Thompson and White, 1964; White and others, 1964). Ages of the volcanic rocks in the region are based on potassium-argon dates, largely summarized by Silberman and McKee (1972). Two rhyolite domes northeast of the area of figure 2 are Miocene in age, and one rhyolitic dome east of the area of figure 2 is correlated with the Steamboat Hills Rhyolite and is Pleistocene in age.

GEOLOGY OF THE AREA NEAR STEAMBOAT SPRINGS

Pre-Tertiary metamorphic and granitic rocks are exposed in and near the thermal area (fig. 2). Soda trachyte flows of the Alta Formation and andesite flows of the Kate Peak formation are the most widespread volcanic rocks near the thermal area. Dikes and extrusive tuff-breccia of the Kate Peak Formation occur at depth (as shown in drill core) beneath younger alluvium and sinter. A basaltic andesite flow of the Lousetown Formation up to 24 m thick, informally designated the Steamboat basaltic andesite by White, Thompson, and Sandberg (1964), and herein called basaltic andesite, caps much of the area to the west, southwest, and south of the hot spring sinter deposits (fig. 2). Four small pumiceous rhyolite domes lie on a line that extends from about 3 km southwest to 5 km northeast of the hot springs (Thompson and White, 1964, pls. 1, 2). A related rhyolite, perhaps a shallow intrusion, is suspected from structural evidence to underlie sinter deposits of Sinter Hill along the same linear trend (White and others, 1964, pl. 2, p. B51).

Rhyolitic pyroclastic material was ejected from several of the vents now occupied by domes of Steamboat Hills Rhyolite, shown on figure 2. Other rhyolitic domes occur in the region and are shown on fig. 1 and discussed by Thompson and White (1964). These include the Mustang dome and other smaller intrusions in the Pah-Rah Range north of the Truckee River canyon (loc. 7, fig. 1) and the Sutro and Washington Hill domes in the Virginia Range (locs. 6 and 8, fig. 1). These domes range in age from about 1.2 to 10.9 m.y., according to the radiometric data, indicating a long history of rhyolite magma generation in the region.

Alluvial deposits of pre-Lousetown and post-Lousetown age (pre-Lake Lahontan) are exposed near the springs. The pre-Lousetown alluvium lies on a pediment surface cut before eruption of the basaltic andesite, and is included with this andesite in figure 2. Chalcedonic and opaline hot-spring sinter deposits occur over large parts of the thermal area, particularly in its northern and eastern parts. At the borders of the sinter terraces, sinter is interbedded with alluvial deposits, and some sinter is still forming from the active springs.

Alteration has affected nearly all of the rock units exposed in the thermal area. Veins and veinlets of opal, chalcedonic quartz, quartz, and calcite cut the granitic and volcanic rocks and sediments, as shown in drill core. The opal and chalcedony veinlets are largely restricted to depths of less than 25 m. Veins up to 2½ m thick of interlayered quartz, chalcedony, and calcite are relatively common at greater depths below the hot spring terraces, with calcite being especially abundant at depths greater than 50 m.

Present spring discharge is localized along the Low and Main Terraces at altitudes below 1,530 m (4,670 ft) in the eastern part of the thermal area (fig. 2). The top of the convecting thermal water under the western part of the thermal area at altitudes ranging from 1,540 to 1,660 m (4,700 to 5,050 ft) is now from 15 to more than 35 m below the surface. Discharge at the surface in these western areas is now restricted to warm and hot vapors, but these western areas were formerly major loci of hot-water discharge and sinter deposition, related to higher ground-water levels in the past.

STRUCTURAL CONTROL OF HOT-SPRING ACTIVITY

Extensive normal faulting has occurred in and near the Steamboat Springs area (generalized in fig. 2). Most major fault movement occurred after Kate Peak volcanism and before cutting of the pre-Lousetown pediment (White and others, 1964). No displacement is demonstrably younger than the Lake Lahontan and Holocene sinter deposits and sediments (White and others, 1964), although some younger faulting seems likely because local earthquakes are frequent in the area (White and others, 1964, p. B53–B54). Fissures controlling the thermal activity may be maintained in part as open channels by periodic slight movement. However, no vertical displacement is evident on fissures in near-surface sinter. The fracture systems of the sinter terraces are clearly tectonic in origin, although chemical attack by sulfuric acid and disintegration adjacent to these fissures maintain them as open channels of varying width near the surface. The Steamboat Springs fault system, which bounds the eastern margin of Steamboat Hills and underlies the Low and Main Terraces (shown as faults, fig. 2) is a major fault system with displacements of at least 300 m, mostly older than the pre-Lousetown pediment gravels (White and others, 1964).

The thermal area lies along a line of Quaternary pumiceous rhyolite domes (fig. 2) that extend 3 km southwest of Steamboat Springs and 5 km to the northeast. Structural tilting away from the crest of Sinter Hill (loc. 2, fig. 2) suggests the presence of another rhyolite dome or shallow intrusion beneath

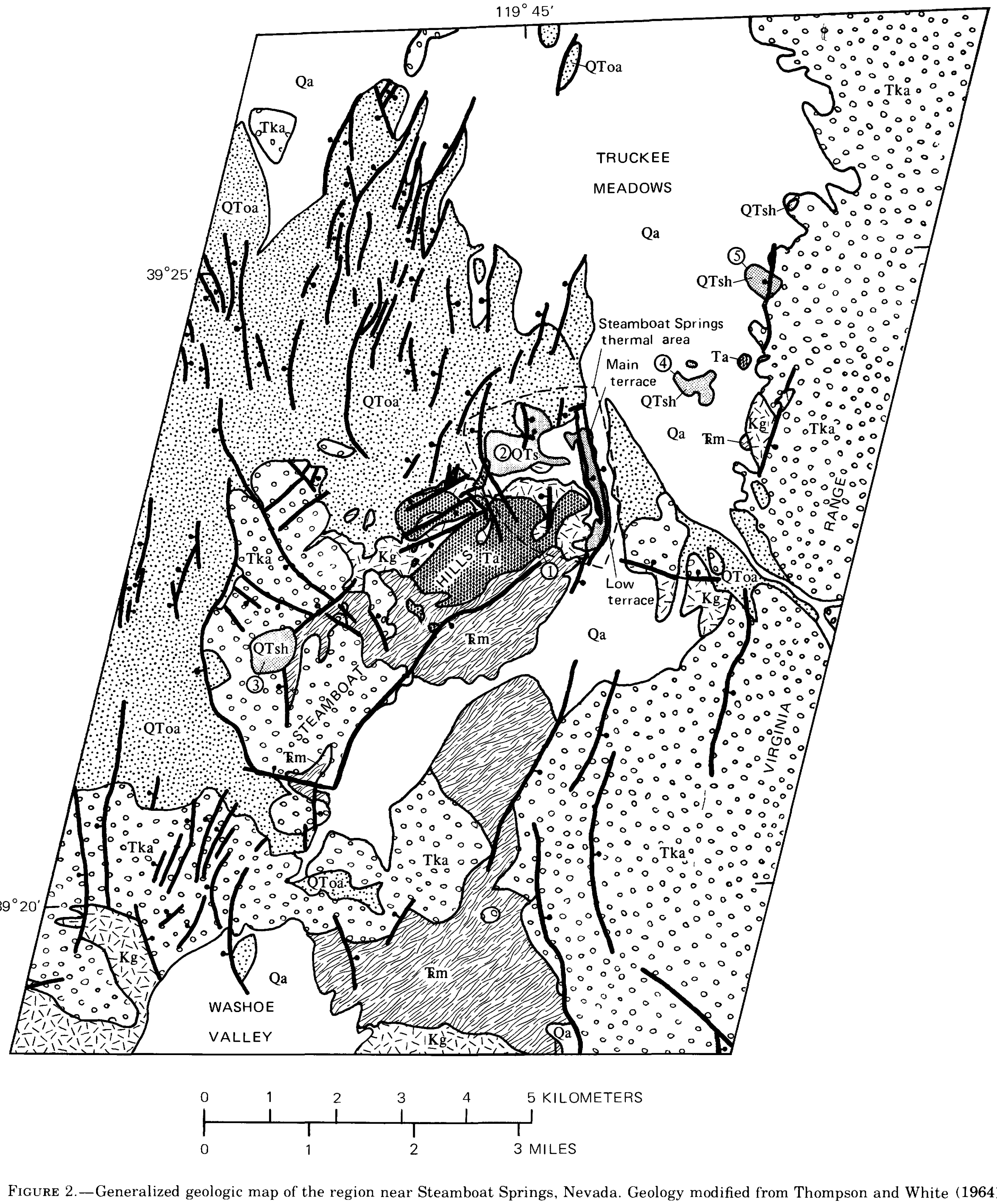

FIGURE 2.—Generalized geologic map of the region near Steamboat Springs, Nevada. Geology modified from Thompson and White (1964).

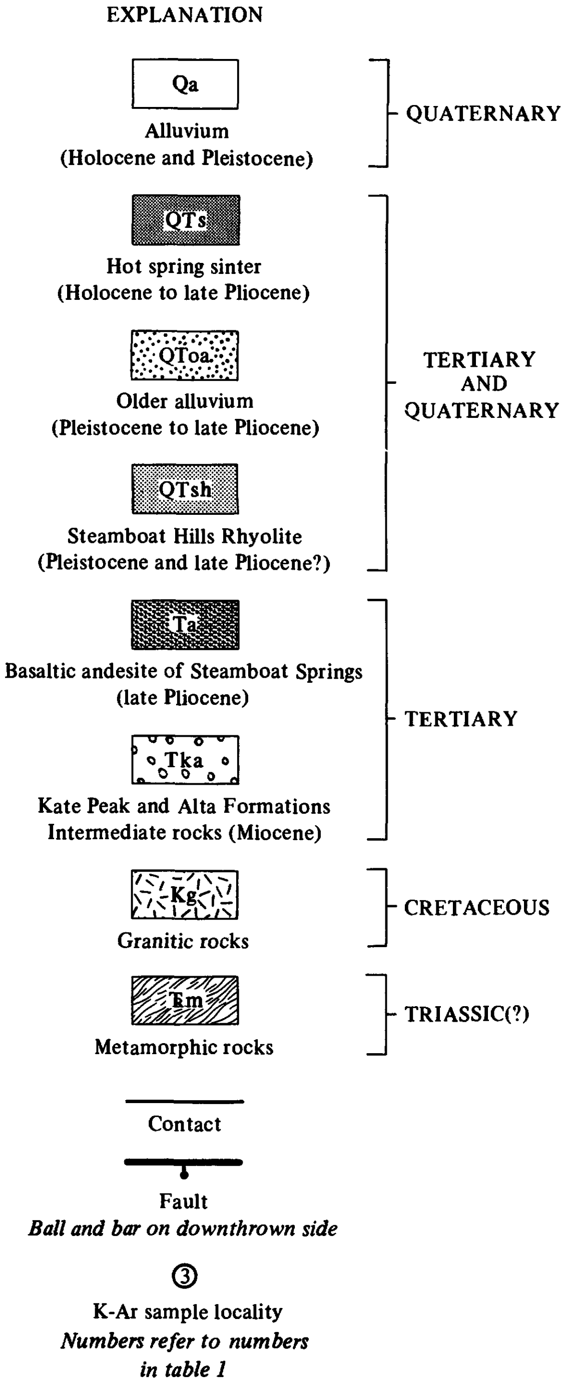

FIGURE 2.—Continued.

this hill (White and others, 1964, pl. 2). The line of rhyolite domes also parallels a major set of faults in the Steamboat Hills (not shown on fig. 2; see White and others, 1964, pl. 1). The hills have a northeast-trending anticlinal form produced by a combination of warping and tilting of fault blocks (Thompson and White, 1964). The alinement of domes and the trend of the hills probably have a common origin as parts of a major structural zone of weakness.

ALTERATION AND ORE MINERALS

Hydrothermal alteration has affected most of the igneous rocks and sediments in the thermal area. The characteristics of hydrothermal alteration of the Steamboat Springs system and the extent of near-surface leaching are summarized by White, Thompson, and Sandberg (1964) and Schoen, White, and Hemley (1974). Acid sulfate alteration is generally dominant near the surface and above the water table where hydrogen sulfide rises and oxidizes to sulfuric acid, chemically attacking the rocks (Schoen and others, 1974). Where generation of acid has been abundant, the end result is replacement of the silicate minerals by opal or cristobalite, except for relict original quartz. Below the water table of an acid-leached area, the silicate minerals are generally replaced by alunite and kaolinite (Sigvaldason and White, 1962). Drill cores show that this advanced argillic alteration dies out with depth. In one drill hole in the Thermal area, kaolinite extends to -83 m. At greater depths in this hole and in other drill holes at shallower depths, albite and the clay minerals montmorillonite, illite, and chlorite dominate most alteration assemblages. Pyrite is usually disseminated in the altered rocks. Calcite and rarely manganiferous calcite occur as veins. Hydrothermal monoclinic potassium feldspar (adularia) is an abundant replacement of plagioclase in several drill holes but does not occur in veins. Adularized rocks change at greater depth to a more complex assemblages of illite, other clay minerals, adularia, calcite, chlorite, and pyrite. A sample of basaltic andesite that was almost completely replaced by adularia, quartz, and minor celadonite was dated in this study.

ORE MINERALS, TRACE-METALS, AND ISOTOPIC COMPOSITIONS

Sinter at Steamboat Springs generally contains detectable quantities of gold and silver, and as much as 10 ppm of gold and nearly 40 ppm of silver have been reported in siliceous muds deposited from the springs (Brannock and others, 1948). These muds also contain up to 0.02 percent mercury and 3.9 percent antimony. Fine-grained cinnabar is disseminated in old chalcedonic sinter of Sinter Hill, and elemental mercury has been identified in vapors from several drill holes and hot-spring vents. Stibnite has been deposited as needlelike crystals in hot-spring pools, and in veinlets and cavities to depths as great as 45 m. Veins of chalcedony, quartz, and calcite up to 2½ m thick contain a few rich stringers of pyrargyrite (Ag_3SbS_3).

White (1955, 1974) has emphasized the similarity in trace-metal association and alteration mineralogy between hot-spring systems such as Steamboat Springs and epithermal gold-silver deposits. The veins and alteration assemblages at Steamboat Springs are indistinguishable from those of many vein gold-silver deposits, but as yet no large mass of ore metals of economic grade has been found by deep drilling at Steamboat Springs or in any other active hot-spring system.

Stable isotopes of spring waters indicate an almost exclusively meteoric source for the thermal waters, but the data are such that as much as about 10 percent of magmatic water would not be detectable (White, 1974). The isotope relations show an oxygen shift, or increase in $\delta^{18}O$, of 2 to 3 per mil, which is common in high-temperature geothermal waters. The shift results largely from interaction of meteoric water and rock silicates during alteration and, commonly in small part, from boiling (Truesdell and others, 1977). The system is probably recharged from both the Virginia and Carson Ranges, with only minor near-surface dilution from local precipitation. Isotopic studies of the waters and rocks from active hot-spring systems and epithermal ore-depositing systems (Taylor, 1974; O'Neil and Silberman, 1974; and White, 1974) demonstrate that many of these systems have been strongly influenced by water of meteoric origin. Steamboat Springs has isotopic similarities to many high-temperature hydrothermal systems, both active and fossil, elsewhere in the Great Basin and the world. Hence, information on the duration of hydrothermal activity in this presently active system yields valuable limits on the time spans of fossil ore-depositing systems.

GEOLOGIC LIMITS ON THE TIME SPAN OF HYDROTHERMAL ACTIVITY

The limits on the time span of activity in the Steamboat Springs systems come largely from the mapped geologic relations of the volcanic and sedimentary rock units (fig. 3) and from radiogenic ages of fresh and altered rocks in and near the thermal area. The youngest sinter deposits are porous opaline silica that is amorphous by X-ray. With increasing depth of burial, consequent higher-than-surface temperatures, and time, opaline sinter is first converted to β-cristobalite, then frequently to α-cristobalite, and finally to chalcedonic sinter (White and others, 1964). The interrelations of the sinters, the alluvial

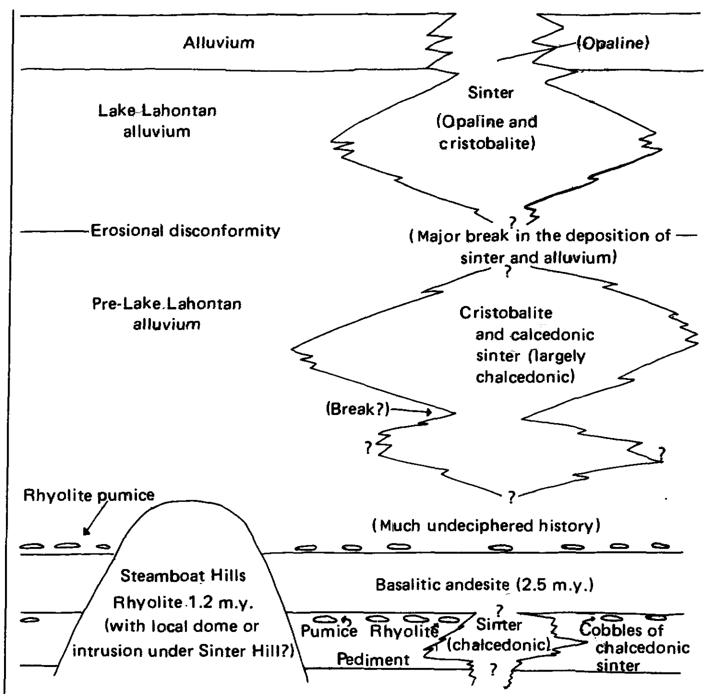

FIGURE 3.—Composite stratigraphic relations at Steamboat Springs, Nevada. Constructed from geologic data of White, Thompson, and Sandberg (1964) and Thompson and White (1964).

deposits, and the dated volcanic rocks provide geologic limits on the inferred time span of hydrothermal activity. However, specific ages are not nearly abundant enough to determine whether or not activity was continuous in the area, or if the recognized intervals of nondeposition of sinter at specific sites represent short intervals or major parts of the total time span. Our present interpretations are represented by figure 3, but times of sinter deposition may be greatly exaggerated.

A complex history of erosion alternating with deposition of alluvium and hot-spring sinter throughout the late Tertiary and Quaternary is revealed from surface mapping and subsurface drill-hole data (White and others, 1964). In general, the hot-spring deposits are interpreted as local facies of the sedimentary formations during periods of alluviation; interfingering of alluvium and sinter is largely inferred rather than observed because the few drill holes were sited near the centers of hot-spring activity and sinter deposition, which are usually positive landforms rising above surrounding areas of alluviation. Interpreted relations are best illustrated in sections A–A′ and B–B′ of plate 2 of White, Thompson, and Sandberg (1964).

Indirect evidence indicates that the oldest hot-spring deposits antedate the local basaltic andesite flows of the Lousetown Formation (reviewed by White and others, 1964, p. B34). Chalcedonized cinnabar-bearing sinter in the western part of the thermal area was incorporated in pediment gravels that underlie the basaltic andesite. Thus, basaltic andesite is the most important single stratigraphic unit in establishing the total time span of hydrothermal activity.

Post-Lousetown sinter deposits are thick and extensive and have been mapped by White, Thompson, and Sandberg (1964) as two major units; the older is largely chalcedonic and cristobalitic, and the younger consists of largely of amorphous opal. Each major unit probably consists of an older and a younger part that can be distinguished locally, especially in drill holes, but not throughout the area. The minimum age of the older part of the chalcedonic sinter is established by the radiometric age of hydrothermal potassium feldspar that replaced the underlying basaltic andesite, as discussed in a later section.

The drill-hole data were first interpreted (White and others, 1964) as indicating continuous or nearly continuous activity during much of the Quaternary, especially including the last three glacial periods when alluviation and water tables were at high levels. During periods of deep erosion and low water table, according to this explanation, springs did not always emerge at the surface, but thermal waters continued to circulate below the surface. A second alternative assumes that surface and near-surface activity was not continuous. Perhaps for relatively long periods of time, no water was discharged from the system, even below the surface into surrounding ground. The first alternative was preferred by White, Thompson, and Sandberg (1964), but the span of time indicated by present data is so long and the required quantities of total heat and rare soluble constituents are so large that the hypothesis of continuous activity must be reevaluated.

The rhyolite domes nearby (fig. 2) are next in importance to the basaltic andesite in establishing the time span of activity in the thermal system. A pumiceous rhyolite dome was extruded about 5 km southwest of the thermal area. Three pumiceous domes are from 2 to 5 km to the northeast (locs. 3, 4, and 5), fig. 2, and another pumiceous dome is 20 km to the east-southeast (loc. 6, fig. 1). All of these domes are similar in mineralogy, bulk chemistry, abundance of phenocrysts, and degree of erosion, and all were included by Thompson and White (1964) in their Steamboat Hills Rhyolite.

Other domes to the north and northeast (locs. 7 and 8, fig. 1) consist of devitrified rhyolite and, locally, perlite. Although they are chemically similar to domes of the Steamboat Hills Rhyolite, their degree of devitrification and erosion and their relations to the Kate Peak and Coal Valley ("Truckee") Formations indicate that these northern domes are older (Thompson, 1956; Bonham, 1969).

The dome in Steamboat Hills (fig. 2) was extruded into a crater from which pumice and older rock fragments had been ejected; this dome is also capped by the eroded remnants of a small crater from which some pumice was erupted. Pumice fragments tentatively correlated with the earlier pyroclastic eruption are locally abundant in pediment gravels beneath flows of the basaltic andesite (Thompson and White, 1964; White and others, 1964), and pumice lapilli correlated with the last explosion are widely but sparsely scattered over the surface of the same flows. Thus, the basaltic andesite flows and the rhyolite domes had been considered almost contemporaneous in age. This conclusion now requires modification on the basis of potassium-argon dates.

POTASSIUM-ARGON AGES

SAMPLES

Samples from most of the pumiceous rhyolite domes in the region were collected for dating. The rocks consist of small crystals of sanidine, sodic plagioclase, quartz, sparse biotite flakes, and spherulitic crystallites in a pumiceous glassy matrix. The rhyolites of Mustang dome (No. 7, fig. 1) and the Washington Hill dome (No. 8, fig. 1) are much denser than the pumiceous rhyolites (probably because the pumiceous mantle was eroded) and consist of small phenocrysts of plagioclase and spherulites in a devitrified groundmass. Sparse small biotite crystals occur in the Washington Hill dome, and vesicles in the rhyolite of Mustang dome are partly filled with tridymite.

Feldspar mineral separates were prepared from all the rhyolite samples. A biotite separate was also obtained from the Washington Hill dome, and obsidian samples were collected from the domes on the flanks of the Virginia Range (Nos. 5 and 6, figs. 1 and 2).

The basaltic andesite contains abundant plagioclase, common olivine and ilmenite, and very sparse pyroxene phenocrysts, and a few partly resorbed quartz crystals. The largest plagioclase phenocrysts commonly contain cores or intermediate zones showing rounded resorption boundaries and a distinctive grained texture (White and others, 1964, p. B36–B37) that is recognizable even in rocks completely replaced by opal or by hydrothermal potassium feldspar. Groundmass feldspar is cloudy with inclusions and consists of at least two species, one of which is anorthoclase overgrowths on plagioclase crystals. The anorthoclase may grade into sanidine in the groundmass (Schoen and White, 1967). Feldspar separates of different densities were prepared from the basaltic andesite by techniques described at the end of this report, and a whole-rock sample was also prepared for dating. A sample of altered basaltic andesite from a depth of 30 m in GS–6 drillhole (loc. 2, fig. 2), almost completely replaced by potassium feldspar with minor celadonite and iron oxide after olivine, was also prepared for dating. The altered basaltic andesite, recognizable from its characteristic texture and high TiO_2 content, was intersected between 28 m and 40 m (cross section *I–I′*, pl. 2, White and others, 1964).

BASALTIC ANDESITE

The indicated ages of the three feldspar separates of the basaltic andesite, which differ in K_2O content by a factor of 4, agree well at approximately 2.5 m.y. (table 1, loc. 1). The whole-rock age is slightly lower, possibly because argon was lost from the fine-grained interstitial potassium-rich material, a fraction of which is undevitrified glass.

RHYOLITE DOMES

Ages of feldspars from the pumiceous rhyolite domes near the thermal area all agree within experimental uncertainty at approximately 1.2 m.y. The Sutro dome (loc. 6, fig. 1) has concordant obsidian and plagioclase ages of 1.5 m.y. and thus is slightly older than the other three age-dated domes of the Steamboat Hills Rhyolite (table 1, locs. 3–6). Two separate samples of obsidian were collected from the dome at locality 5 (fig. 2). One was an obsidian marble (marekanite) from the top of the dome (sample SR–5A, table 1), and the other was a nonperlitized residual of similar shape collected from a trench cut in perlitic obsidian (941–B). Both yielded ages of 3.0 m.y., in marked disagreement with the sanidine age of 1.2 m.y. from pumiceous rhyolite from the top of the same dome. It is possible that the obsidian marble was ejected during a previous eruption and was carried up to its present location by extrusion of the dome. The age of the other sample, however, cannot be explained by this process. Close agreement of the ages of feldspars from all of the domes in the vicinity of the thermal area suggests that those domes were extruded at approximately the same time.

Three possible explanations could account for the discordance. The first is that the obsidian age represents the true age of extrusion (and cooling) of this dome, and that the feldspar age has been reset through argon loss during a later thermal episode. A thermal event of this magnitude that could affect the feldspar phenocrysts of undevitrified obsidian erupted at the surface, without affecting the argon content of the obsidian, seems highly improbable.

A second possible explanation entails the presence of extraneous argon-40 in the obsidian sample. If argon incorporated from the environment was present in the rhyolite magma (Damon and others, 1967), it is possible that rapid cooling, indicated by the presence of the obsidian, did not permit outgassing or loss of this argon.

Experimental studies by Fyfe, Lanphere, and Dalrymple (1969) have demonstrated that argon is soluble in hydrous granitic melts and that a significant quantity of it is retained in the glass resulting from quenching these melts. Anomalously old ages, attributed to the presence of extraneous argon-40, have been reported from nonhydrated volcanic glass from a vitrophyre in welded ash-flow tuff in southern Nevada (Marvin and others, 1970); the amount of excess argon-40 necessary to explain the discrepancy of this example is on the order of 0.5 to 1×10^{-11} mole/g. If extraneous argon-40 in the glass accounts for the discordance between the sanidine and obsidian ages of the Truckee Meadows dome, approximately 1.3×10^{-11} mole/g of extraneous argon must be present, an amount approximately twice that reported by Marvin, Mehnert, and Noble (1970). The environment of generation of the magmas was probably similar as well. Both were rhyolitic and presumably resided in, and were erupted from, shallow levels in the crust.

A problem with this explanation is the low volatile content, expressed by H_2O in the obsidians from this dome. An obsidian marble (marekanite) collected from the dome had a content of only 0.25 percent H_2O (D. E. White, unpub. data). Friedman and Smith (1960) indicate that the best explanation for the low H_2O contents of nonperlitic glassy obsidian is loss of water during and before extrusion, which presumably would be accompanied by exsolution of other volatiles, including argon. However, in the experimental study by Fyfe, Lanphere, and Dalrymple (1969), in which granitic compositions with added water contents between 0 and 37 percent were melted in a closed system and then quenched, the amount of argon remaining in the melt after quenching was roughly in inverse proportion to the water content, being controlled by its partial pressure relation to H_2O pressure. However, the emplacement of rhyolite domes at surface pressure is an open system (as evidenced by the generation of pumiceous mantles), so that the experimental work is not directly applicable to the natural extrusion of the rhyolite dome. There is evidence, however, of extraneous argon in nonhydrated glass from both field and experimental work, and this explanation for the apparently discordant ages should be considered.

TABLE 1.—*Analytical data and radiometric ages for volcanic rocks from the Steamboat Springs area*

Location No.	Sample No.	Rock and locality	Mineral	K_2O (percent)[1]	Ar^{40}(rad) (mole/g × 10^{-11})[2]	Ar^{40}(rad)/Ar^{40}(total)	Age[3] (million years)
1	128–0 A	Basaltic andesite, Steamboat Hills.	Whole rock	2.25 2.27	0.717	0.30	2.15±0.10
	128-0 B		Plagioclase, An_{58-62}.	.845 .837	.314	.17	2.52±0.19
	128-0 D		Alkali feldspar and plagioclase.	2.68 2.70	1.014	.27	2.55±0.11
	128-0 E		Alkali feldspar.	3.65 3.67	1.369	.27	2.53±0.10
2	GS-6-100	Basaltic andesite altered to adularia, Steamboat Hills.	Whole rock	10.07 10.07 9.93 9.92	1.453 1.792	.08 .06	1.10±0.11
3	SR-1	Rhyolite dome, Steamboat Hills.	Sanidine	9.06 9.02	1.525	.48	1.14±0.05
4	SR-4	Rhyolite dome, Truckee Meadows.	do	10.53 10.53	1.893	.33	1.21±0.06
5	SR-5	do	do	10.53 10.54	1.801	.51	1.16±0.05
	SR-5A		Obsidian Whole rock	4.40 4.38 4.96 4.97	2.094	.78	3.03±0.12
	941-B		do	4.85 4.80 4.63 4.63	2.074 2.078	.45 .61	2.97±0.09
6	SR-3	Sutro rhyolite dome, Virginia Range.	Alkali feldspar.	2.27 2.24	.502	.14	1.51±0.22
	SR-3A	do	Obsidian	4.43 4.29 4.58 4.38	.987	.27	1.51±0.06
7	MR-2	Mustang rhyolite dome, northeast of Reno.	Plagioclase	1.039 1.039	1.261	.34	8.2 ±0.3
8	C-3	Rhyolite dome, Washington Hill.	Biotite	7.69 7.72	12.75 12.22	.34 .43	10.9 ±0.3
	C-3A	do	Plagioclase	.812 .813	1.168	.42	9.7 ±0.3

[1]Potassium was analyzed using a lithium metaborate fusion flame photometry technique with the lithium serving as an internal standard (Ingamells, 1970); mineral standards used for calibration. Analytical uncertainty is approximately ±1 percent of content reported. Analysts, L. Schlocker and J. A. Christie.

[2]Argon was extracted from purified mineral separates or crushed and sized whole-rock samples (see text for details of sample preparation) in a Pyrex high-vacuum system using external R. F. induction heating. During fusion a calibrated spike of purified argon-38 is introduced. Reactive gases were removed by an artificial molecular sieve, copper-copper-oxide and titanium furnaces. Mass analyses of the purified argon were made with a Neir-type 60° sector 15.2-cm-radius mass spectrometer operated in the static mode. Analytical uncertainty of argon measurement is approximately 2 to 10 percent, depending on amount of atmospheric argon (Dalrymple and Lanphere, 1969). Analyst, M. L. Silberman.

[3]Analytical uncertainty estimated at one standard deviation; includes uncertainties in measurement of argon and potassium.

Another problem with this explanation is the concordance in age of alkali feldspar and obsidian from the Sutro dome (loc. 6, fig. 1). The obsidian in this sample was collected from rhyolite breccia, including obsidian fragments, from the west flank of the dome. The obsidian at this locality has no extraneous argon relative to the feldspar. The obsidian does contain a higher proportion of crystallites and may have lost most or all or its volatiles, perhaps owing to slightly slower cooling than the sample from locality 5, which has few crystallites and consists mostly of glass.

The third explanation, favored by White, is that the dome at locality 5 actually includes rhyolite of two different ages, with the analytically determined ages being approximately correct. Although we have not recognized petrographic or other evidence for distinguishing two mappable units in this dome, the age data are consistent with the geologic evidence for two distinct extrusive events, thus accounting for rhyolitic pumice both above and below the 2.5-m.y. basaltic andesite. Within the geologic framework, this explanation is the most easily acceptable one.

The Mustang dome (loc. 7, fig.1) yielded an 8.2-m.y. age from plagioclase. This dome was considered partly correlative in age with the Coal Valley ("Truckee") Formation of late Miocene to middle Pliocene age, because it and similar rhyolites intrude the lower part of the formation, and debris from the domes is incorporated in clastic sediments of the upper part (Bonham, 1969). The 8.2-m.y. age places this rhyolite in the late Miocene according to the time scale of Berggren (1972), although the age is Pliocene according to the time scale used by Bonham (1969).

The rhyolite dome at Washington Hill (loc. 8, fig. 1) yielded ages of 10.9±0.3 m.y. (biotite) and 9.7±0.3 m.y. (plagioclase). These ages suggest that this dome is also late Miocene. Thompson and White (1964) interpreted this rhyolite as contemporaneous with the upper part of their Truckee Formation (Coal Valley Formation of Bonham, 1969). The Coal Valley ("Truckee") overlies and intertongues with the upper part of the Kate Peak Formation (Thompson and White, 1964), from which a potassium-argon age of 12.4±0.2 m.y. was reported (Silberman and McKee, 1972). The radiometric ages of these domes are in agreement with their stratigraphic relations.

ALTERED BASALTIC ANDESITE

The altered basaltic andesite from drill hole GS–6 (loc. 2, fig. 2; Schoen and White, 1967) was dated as a whole-rock sample. The rock is almost completely replaced by potassium feldspar with some quartz, and minor celadonite, illite, montmorillonite, iron oxides, and sulfides. Temperatures measured as drilling progressed indicate a present temperature of about 53°C at the sample depth, which is too low to drive argon from the feldspar. However, temperatures are certain to have been much higher in the past, when Sinter Hill was an active sinter-depositing part of the whole system. If we assume that about 15 m of sinter has been eroded and that former temperatures and pressures had adjusted to the then-existing reference boiling-temperature curve, sample GS–6–100 was altered at a temperature near 150°C. Still higher temperatures may have been caused by a shallow rhyolitic intrusion under Sinter Hill (White and others, 1964). These possibilities suggest that the indicated age of 1.1 m.y. for the hydrothermal alteration of this sample may be a minimum figure.

INTERPRETATION

The radiometric ages and the geologic evidence indicate that thermal spring activity has occurred over a time span of about 3 m.y., in part predating the 2.5 m.y. basaltic andesite (fig. 3). Rhyolitic tephra also predates the basaltic andesite, perhaps correlating with the enigmatic potassium-argon ages of about 3.0 m.y. obtained from the two obsidian samples of the dome at locality 5.

A number of small pumiceous rhyolite domes were extruded in the region from 1.2 to 1.5 m.y. ago, and these domes show clear relations to the thermal activity. The thermal area occurs on a northeast-trending line that connects the four domes of Steamboat Hills and Truckee Meadows; vertical uplift under Sinter Hill is likely to have been caused by a shallow intrusion correlative with these domes; and the age of hydrothermal alteration of basaltic andesite (1.1 m.y.) also suggests approximate contemporaneity.

A second cluster of domes near the Truckee River 12–20 km northeast of Steamboat Springs yields considerably greater ages, from 8 to 10 m.y. These domes are not related to the Steamboat thermal system and, instead, are probably late stages of the volcanism associated with generation of the Kate Peak Formation and the epithermal silver-gold deposits of the Comstock lode.

IMPLICATIONS OF THE AGE OF THE STEAMBOAT SPRINGS THERMAL SYSTEM

White (1968, p. C102) estimated that a magma chamber of at least 100 km^3 underlying the Steamboat Springs area must have cooled and crystallized to maintain the hydrothermal system at its present rate of activity for 100,000 years. The total time span from earliest activity to the present, in the light of our new age data, is about 30 times more than had been assumed. If heat was supplied at a constant rate over this span of time (a doubtful assumption), a batholithic volume of 3,000 km^3 is required.

Smith and Shaw (1975) estimated the volume of the magma chamber below the Steamboat Springs thermal system as between 45 and 180 km^3, with a best estimate of 90 km^3. Their estimate is based on a model that relates volume of chamber to surface distribution of exposed silicic volcanic rocks. R. L. Smith (oral commun., 1975) recognizes the inconsistencies between the two sets of estimates and suggests that a large magma chamber may underlie the area but at great depth, with abnormally low representation by extruded rhyolite. The Steamboat Springs system plots in the "low geothermal potential" field of their age-volume diagram (Smith and Shaw, 1975, p. 74). Smith and Shaw assumed an age of 1.2 m.y. rather than 3 m.y. as the time span of the system, but this makes little difference in the position of the point on their plot. An assumed volume of 3,000 km^3 makes the Steamboat Springs system much more consistent with other magma-sustained geothermal systems.

Many large volcanic systems are now known to have complex histories that may span as much as 8 to 10 m.y., with the silicic events generally occurring late in this history and involving two or more cycles of ash-flow tuff eruptions and caldera collapse. The Yellowstone Park volcanic system of Wyoming has undergone three such cycles, with culminations occurring 2.0, 1.2, and 0.6 m.y. ago (Christiansen and Blank, 1972). The Valles Caldera of New Mexico had two major events 1.4 and 1.0 m.y. ago (Doell and others, 1968). These clear indications of major pulses of rhyolitic magma suggest that the magma source for the Steamboat thermal system has also had major pulses. An assumption of a steady state for as much as 3 m.y. should now be questioned seriously.

Smith and Shaw (1975) developed another principle that implies variability, the "shadow-zone" concept for utilizing basaltic volcanism to indicate the physical state and crude dimensions of an underlying silicic chamber. Briefly stated, this concept assumes that basalt from the mantle can be erupted outside of and contemporaneously with a molten silicic magma chamber, but that basalt cannot be erupted through silicic molten magma without mixing. A very important corollary is that if basalt is erupted through a silicic magma chamber, the chamber must have been sufficiently crystallized and cooled to sustain fracturing and formation of a conduit for the basalt. These concepts, as applied to the Steamboat system and in view of our age data, indicate the presence of a molten rhyolitic magma chamber of unknown volume and distribution about 3 m.y. ago. This silicic chamber could not have maintained the same dimensions in a molten state 2.5 m.y. ago, when flows of the basaltic andesite were erupted only 1½ km southwest of the thermal area (Thompson and White, 1964). In contrast, a single molten silicic chamber probably did develop under Steamboat Hills and adjacent parts of Truckee Meadows and central Virginia Range roughly 1 to 1.5 m.y. ago, engulfing the channels of the earlier basaltic andesite.

Approximately steady-state discharge of heat and dissolved constituents was favored by White (1968, p. C1) for 10,000 or even 100,000 years: "During periods of deep erosion and low water table, springs did not always emerge at the surface, but for at least parts of these periods thermal waters continued to circulate below the surface***." Now, in view of the indicated long time span of perhaps 3 m.y., we must suspect that periods of surface discharge and sinter deposition alternated with periods of no activity, as indicated in figure 3.

STEAMBOAT SPRINGS AS RELATED TO OTHER THERMAL SYSTEMS AND HYDROTHERMAL ORE DEPOSITS

It is useful to compare the age of hydrothermal activity of the Steamboat Springs thermal system with other hydrothermal systems, both active and fossil. The similarities of ore-forming systems and active thermal spring systems indicate that their timespans may also be comparable. Table 2 lists the estimated timespans of thermal spring systems (perhaps including major interruptions) and epithermal ore-depositing systems. It is difficult to determine accurately the timespan of hydrothermal systems unless isotopic dating studies are made for that specific purpose. The best information on both fossil and active systems comes from the isotopic ages of alteration and vein minerals, as interpreted by geologic relations. In most places, such minerals are either not dated or dating is inadequate to estimate the timespan and continuity. Generally, geochronological studies in thermal areas emphasize the volcanic history. It is then necessary to estimate the age of hydrothermal activity from the available data. These ages were estimated for several of the systems listed in table 2 using geologic and stratigraphic relations of sinter deposits, or sediments and sedimentary rocks that have sinter associated with them, that are bracketed by or interbedded with volcanic rocks of known age. In many cases, the ages can be estimated with some precision.

According to Renner, White, and Williams (1975), most deeply eroded volcanic systems are associated with extensive hydrothermal alteration, and most of this alteration involves high proportions of meteoric water. Smith and Bailey (1968, p. 642) state that hot springs and solfataras are probably active throughout most cauldron cycles, especially during the late stages of each cycle. The hydrothermal stage may overlap all others, but it becomes uniquely characteristic only after the major eruptions have ceased, when it constitutes the terminal phase of waning volcanic activity. The few available potassium-argon ages from epithermal ore deposits (table 2) show that hydrothermal alteration and mineralization start near but not quite at the end stages of volcanic activity and generally continue for some time after the end of the volcanism (Silberman and others, 1972; Ashley and Silberman, 1976).

In many calderas, a caldera lake forms after collapse. During subsequent resurgent doming and ring-fracture volcanism, the late volcanic rocks are commonly hydrothermally altered, with the lake water as an obvious immediate source of the alteration fluid. If other evidence is lacking, the ages of volcanic rocks associated with resurgent doming or ring-fracture volcanism are used to estimate the lifespan of the hydrothermal alteration reported in table 2. This method of estimating the age of hydrothermal activity is not considered as reliable as the others discussed above.

These data show that timespans on the order of 0.1 to 1.5 m.y. have been recorded for thermal spring activity. Steamboat Springs, with an indicated timespan of about 3 m.y., may represent an upper limit for thermal springs that are still active. For the fossil systems, the ages are probably minimum estimates because it is not always possible to determine the earliest and latest phases of alteration or veining.

The results of the long-lived hydrothermal activity at Steamboat Springs, and the shorter but still relatively long activity indicated for some other active

TABLE 2.—*Timespans of geothermal and ore-forming epithermal systems*

Locality (active systems)	Timespan	Evidence	References
Valles Caldera, N. Mex.	1.2 m.y. to present (probably not continuous, with present geothermal system active perhaps 100,000 yr).	Altered postcaldera rhyolite domes, possibly erupted into a 1.2-m.y. caldera lake. Later ring fracture domes erupted between 1.1 and 0.4 m.y. are not altered; present active system probably younger.	Smith and Bailey (1968), Doell and others (1968).
Long Valley, Calif	Maximum intensity ~0.3 m.y. ago, perhaps intermittent to present.	Altered lacustrine sediments interbedded with dated rhyolite.	Bailey and others (1976, p. 741).
Steamboat Springs, Nev.	~3 m.y. to present (probably intermittent).	Sinter older and younger than dated basaltic andesite, 2.5 m.y.	This work.
Yellowstone Canyon hydrothermal system, Wyoming.	Perhaps 0.6 m.y. to present.	Chalcedonic sinter associated with lake sediments older than 0.266 m.y. pumice and less than 0.6 m.y. rhyolite flow.	Richmond (1976, p. 8); R. L. Christiansen (written commun., 1975).
The Geysers, Calif	>57,000 yr to present.	Calculated minimum time required to account for stored heat, assuming present heat flow.	D. E. White (unpub. data, 1975).
Sulphur Bank mine, Clear Lake, Calif.	~27,000 yr to present.	Nearly continuous sequence of Hg anomalies in ^{14}C-controlled lacustrine sediments 3 km offshore from mine.	John D. Sims (written commun., 1977).
Wairakei, New Zealand.	~0.5 m.y. to present	Hydrothermal eruption debris in Huka Falls Formation; approximate age, not reliably controlled.	Grindley (1965, p. 85–87, 0.5 m.y.); 0.3 m.y. now considered more probable (written commun., 1976).
Orakeikorako, New Zealand.	>20,000 yr to present (perhaps not continuous).	^{14}C dating of hydrothermal eruption and of 20,000-yr-old tephra overlying old sinter, probably disconformably.	Lloyd (1972, p. 44); E. F. Lloyd and E. E. White, unpub. data.
Ore deposits			
Bodie, Calif	8.6–7.1 m.y.	K-Ar ages of host volcanic rocks and wallrock alteration and vein hydrothermal feldspar in epithermal precious-metal deposit.	Silberman and others (1972).
Goldfield, Nev	21.0–20.0 m.y.	K-Ar ages of wallrock alteration minerals and hydrothermal vein alunite in epithermal gold deposit.	Ashley and Silberman (1976).
Creede, Colo	24.6±0.3 m.y. in 26.4±0.6-m.y. volcanic rocks; probably discontinuous.	K-Ar dating of fresh and altered rocks in epithermal base-metal–silver deposit.	Bethke and others (1976).
Silver City, DeLamar, Idaho.	15.6 to 15.7±0.3-m.y. rhyolite containing veins 14.8 to 15.2±0.6 m.y.	Vein adularia in epithermal gold-silver; veins contained in rhyolite.	Pansze (1975).
Tui mine, New Zealand.	<7 m.y. >2.6 m.y.; 2.6-4.0 m.y. considered most probable.	K-Ar ages of altered host rocks of epithermal base-metal-silver deposit contained in ~16 m.y. andesite.	Adams and others (1974).

hydrothermal systems, suggest that timespans on the order of 1 to perhaps 3 m.y. commonly may be involved in the formation of epithermal ore deposits.

METHODS OF SAMPLE PREPARATION

Samples of the Steamboat Hills Rhyolite were prepared by crushing, grinding, and sizing quantities of rhyolite to between 0.25 to 0.1 mm. The sized rock material was washed, and feldspar concentrates were prepared by passing the material through a Frantz magnetic separator and heavy liquids. Samples were optically examined with a binocular microscope and then under a petrographic microscope as grain mounts in immersion oils. X-ray diffraction patterns were made of the separates for structural determination of the sanidines. The samples were determined to have sanidine structure by the three-reflection method of Wright (1968).

The drill-core sample of altered basalt was prepared by crushing the core segment (about 5 cm long) and sizing to 2 to 4 mm. Splits of this size were crushed using a diamond crusher and ceramic pulverizer to −150 mesh (less than 0.1 mm) and analyzed for potassium. Two splits of the 2- to 4-mm fraction, each of about 10 g, were analyzed for argon.

The fresh basaltic andesite was the critical sample that provided geologic limits on the minium age of hot-spring activity. First, a whole-rock sample was prepared by hand crushing in a steel mortar and sieving to −170 mesh (88 μm). This sample was split and analyzed for argon and potassium without further treatment.

Petrographic and X-ray examination indicated several different feldspars: (1) plagioclase phenocrysts with cores of calcic oligoclase (An_{26-30}) with overgrowth of andesine to labradorite (An_{42-62}), (2) smaller plagioclase phenocrysts that seem to grade into tiny groundmass laths (White and others, 1964, p. B36), and (3) overgrowths of anorthoclase that may grade into (4) fine-grained sanidine in the groundmass (Schoen and White, 1967). All of these varieties contained potassium, which is probably also abundant in the minor interstitial glass, and all of these may differ in their argon-retention characteristics. Thus, feldspar concentrates of different densities were prepared from the basalt.

Each sample was crushed by hand in a steel mortar and sieved to 88–44 μm size. The sample was elutriated for 1 minute in a 20-cm-tall beaker of water. The process was repeated until the water was clear of small particles; the settled material was then dried.

Magnetite was removed with a hand magnet. The sample was then passed several times through a Frantz separator at 1.2 amps. The non-magnetic fraction consisted of alkali feldspars and plagioclase. This fraction was then centrifuged in bromoform, diluted to a density of 2.691, to separate the bulk of the plagioclase from the fraction rich in alkali feldspar. The plagioclase was washed and dried, constituting sample 128–0 B, which was dated.

The light fraction was centrifuged in bromoform, diluted with acetone to a density of 2.586, to separate lighter material (largely α-cristobalite). The heavy concentrate was centrifuged in a series of mixtures of bromoform and acetone of densities 2.663, 2.640, and 2.617, resulting in alkali-feldspar and plagioclase mixtures of density 2.663–2.691 (128–0 C), 2.640–2.663 (128–0 D), and 2.617–2.640 (128–0 E).

The concentrates represent a series of samples of increasing proportion of alkali feldspar relative to plagioclase. Samples D and E, along with B (pure plagioclase) were split and analyzed for potassium and argon. The analytical data are included in table 1.

A final, nearly pure concentrate of potassium-feldspar (>95 percent) with density between 2.586 and 2.617 was also prepared but was insufficient for dating. The slightly younger indicated age of the whole rock is, in our opinion, likely to be due to argon loss from interstitial glass.

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