[AMERICAN JOURNAL OF SCIENCE, VOL. 265, OCTOBER 1967, P. 641-684]

American Journal of Science

OCTOBER 1967

SOME PRINCIPLES OF GEYSER ACTIVITY, MAINLY FROM STEAMBOAT SPRINGS, NEVADA*

DONALD E. WHITE

U. S. Geological Survey, Menlo Park, California

ABSTRACT. Steamboat Springs, Nevada, is the third most active geyser area in the United States, after Yellowstone Park and Beowawe, Nevada. More than 20 natural geysers have been identified at Steamboat Springs; most are small and inconspicuous, but eruptions to 25 feet were observed during the present study; many shallow wells erupt intermittently as geysers.

Deep drilling for geothermal energy throughout the world has demonstrated that natural geysers are near-surface expressions of very large convection systems in which water of surface origin circulates down to surprisingly great depths—a minimum of 5000 to 10,000 feet in the deepest explored systems. Isotope study of oxygen and hydrogen of the waters demonstrates that at least 95 percent must be of surface origin. Volcanic steam cannot account for more than one quarter of the heat of geyser waters; therefore, the circulating water must be heated largely by thermal conduction through solid rocks. The source of the conducted heat consists of volcanic masses, perhaps still molten, and probably at depths of 2 to 4 miles beneath the Earth's surface.

If water circulates too rapidly through a convection system, or if the supply of heat is too small, the upward flowing water is not high enough in temperature to cause geyser activity. Subsurface temperatures at least as high as 150°C (302°F) and probably above 170°C (338°F) seem to be necessary in the circulation system (but actual temperatures in the immediate vicinity of a small geyser may be much less). Hot water at 150°C has more than adequate energy for a major geyser eruption. Less than 5 percent of the mechanical energy available in the expanding steam could sweep the associated water upward to a height of 200 feet.

A geyser is a special type of hot spring with channels and dimensions too narrow to permit convective loss of all excess energy under steady or constant conditions. Instead, the supply of energy periodically becomes so large that eruption occurs. This paper considers many different aspects of the eruption process which may explain the great differences in individual geyers as well as the changing behavior of a single geyser with time.

INTRODUCTION

Steamboat Springs, Nevada, is the third most active geyser area in the United States. Yellowstone Park is by far the most notable (Allen and Day, 1935), followed by Boewawe, Nevada (Nolan and Anderson, 1934). Small geysers have been identified at four other places in the United States: (1) Morgan Springs, Tehama County, California, a few miles south of Lassen Peak (Waring, 1915, p. 138; and personal observations); (2) a single vent in the mud-volcanoes of the Salton Sea geothermal area (White, 1955, p. 1121), since inundated by the rising level of Salton Sea; (3) Bradys Hot Springs, Churchill County, Nevada; and (4) Umnak Island, Alaska (Byers and Brannock, 1949). Most other "geysers" in the United States are drilled wells, which erupt as man-made geysers, or are misnomers.

---

* Publication authorized by Director, U. S. Geological Survey, Washington, D.C.

The natural geysers of Steamboat Springs are mostly small and inconspicuous, typically erupting to heights of only 1 to 3 feet. Twenty-one natural vents were identified as geysers during the years from 1945 to 1952 (White, in press, table 4). One or more were in periods of activity during about 75 percent of the total interval, but a visitor commonly could observe all the springs without seeing a single eruption. Four natural geysers erupted to heights of more than 4 feet, and the maximum observed was about 25 feet (White, in press, table 4, spring 12).

According to Allen and Day (1935, p. 173), a large geyser near the north end of the Main Terrace erupted to 25 feet in 1924 and 1925, and DeQuille (1876, p. 327-329) stated that one geyser erupted over 50 feet high in 1860 and another from 50 to 60 feet in 1862.

A review of the extensive geyser literature of the past 120 years is not attempted here. Most theories are reviewed by Allen and Day (1935, p. 208-231), Barth (1950, p. 76-88), Noguchi (1956), Iwasaki (1962, and Golovina and Malov (1960). Many authors emphasize one or more of the factors here considered important, but no previous general theory explains many aspects of why geysers erupt, how eruptions are initiated or terminated, how the energy is supplied, why geysers differ so much from one to another, or how a single geyser may come into existence and change its characteristics so much with time.

DEFINITIONS

Several different types of springs have been defined by White (in press). The following especially concerns geysers:

Geyser—a hot spring characterized by intermittent discharge of water ejected turbulently and accompanied by a vapor phase. The temperature of water at the ground surface is generally within a few degrees of the boiling point for pure water at the prevailing atmospheric pressure, but in many hot-spring areas the gas content of the water is high enough for a vapor phase to form at temperatures below boiling for pure water. All hot springs with intermittent turbulent discharge are here considered as geysers, with no defined lower limit in temperature.

Subterranean geyser—the intermittent turbulent ejection of steam and gases within the subsurface parts of a vent. Water is ejected above the general water table, but the latter is too deep for liquid water to appear above ground level; all erupted water that remains liquid eventually drains back into the system. The deeper the water table the more vigorous an eruption must be in order to eject water to the surface. Many geysers have one or more subterranean eruptions between intervals of surface discharge; each subterranean eruption can then be considered as a substage of the whole eruption cycle.

Aquifer eruption—the intermittent ejection of water from an individual aquifer or from a part of the total reservoir system. This intermittent ejection can be expressed as a true geyser or as a subterranean geyser; it can also be associated with continuous discharge from some other aquifer, resulting in pulsations or changes in the rate of discharge of an individual spring or well that is supplied by these multiple aquifers. The concept that two or more aquifers or separate parts of the total reservoir can each discharge continuously or intermittently, and at different rates and at different intervals, is helpful in understanding the complex behavior of many geysers, pulsating springs, and geothermal wells.

ACKNOWLEDGMENTS

I wish to express my deep appreciation to Dr. Philip F. Fix, who made important contributions early in the Steamboat Springs investigations (Brannock and others, 1948) and who helped to stimulate my interest in geyser phenomena.

Other associates on the Geological Survey have provided very helpful discussions, especially R. O. Fournier and L. J. P. Muffler. I have also valued greatly the several opportunities to observe Yellowstone's geysers with Dr. George Marler of the National Park Service, and I have benefited from Marler's keen observational abilities.

RECENT STUDIES OF THE LARGE NATURAL HYDROTHERMAL SYSTEMS

Geyser-bearing systems dominated by hot water.—Many advances in the understanding of hydrothermal systems have been made over the past 20 years, in part from exploration for geothermal energy and in part from general scientific study. Geothermal drilling in hot-spring areas has shown that some yield high-temperature steam with little or no accompanying liquid. Two of these, Larderello in Italy and "The Geysers" in Sonoma County, California, have been highly successful in yielding geothermal steam to generate electricity, but both areas lack natural geysers.

Hydrothermal systems dominated in explored parts by hot water are far more common than systems dominated by vapor (White, 1965). Of the seven known areas of natural geysers in the United States, *all* are of the water-dominated type. Five of these seven—Yellowstone Park, Beowawe, Steamboat Springs, Bradys, and Salton Sea—have now been drilled either for research purposes or for geothermal energy. In the drilled areas temperatures generally increase with increasing depth and are close to the theoretical boiling points for pure water at existing hydrostatic pressures. At some depth and temperature characteristic of each particular system (if drilling is deep enough), temperatures level off and show little further increase within explored depths.

Effects of pressure on boiling temperatures and the boiling-point curve.—In order to understand the significance of the relationships and their bearing on geyser activity, we must consider the effects of pressure on water and steam. The familiar temperature of 100°C (212°F) is the boiling point for pure water with no dissolved gases or salts, and at a pressure of exactly one atmosphere. In a well filled with water, the temperature of boiling increases with depth as the total pressure increases. Several kinds of reference curves have been calculated, showing the

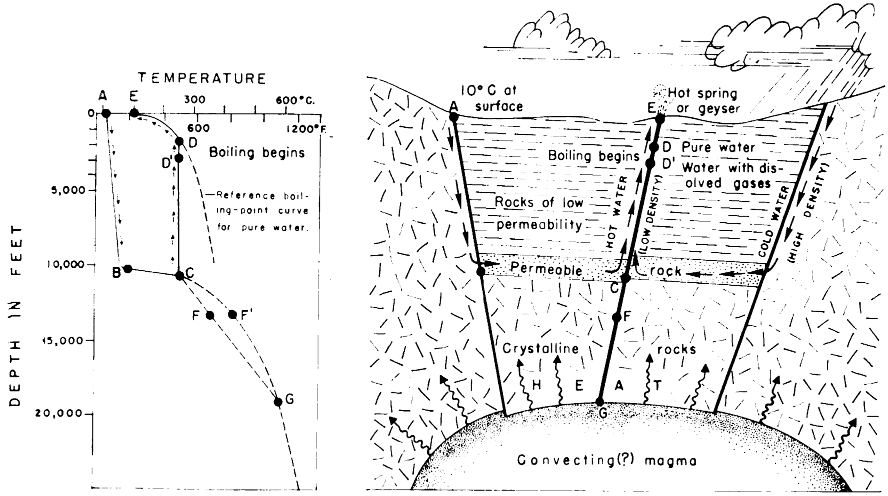


Fig. 1. Generalized model of a high-temperature spring system with deep convective circulation of meteoric water, heated dominantly by thermal conduction rather than by magmatic fluids.

changes in the temperature of boiling of pure water with depth (White, in press). The most useful reference curve for geothermal and geyser considerations involves the decrease in density of water as temperature increases, with external pressure equal to water-vapor pressure. At 100°C, for example, the relative density of pure water is 0.958 or 4.2 percent less than 1.000 at 4° C; corresponding densities at 200°C are 0.863 and at 300°C, 0.715. The reference boiling-point curve of figure 1, expanded to larger scales in other appropriate figures, was calculated by a method of successive approximations (White, in press).

Temperatures measured in wells in all explored geyser areas are strongly influenced by this boiling-point curve. The influence is particularly evident where special effort was made to obtain bottom-hole temperatures as drilling progressed (White, in press).

Subsurface temperatures and deep convection.—The explored geyser areas of the world include (with maximum recorded subsurface temperatures summarized by White, 1965, table 1): Ahuachapán, El Salvador (174°C); Hengill (Hveragerdi), Iceland (230°C); Krysuvik, Iceland (230°C); Otaki, Japan (185°C); Noboribetsu, Japan (160°C); Onikobe, Japan (185°C); Pathe, Mexico (155°C); Ixtlan, Mexico (150°C); Wairakei, New Zealand (268°C); Waiotapu, New Zealand (295°C); Orakei Korako, New Zealand (240°C); Kawerau, New Zealand (285°C); Steamboat Springs, Nevada (187°C); Bradys, Nevada (168°C); Beowawe, Nevada (207°C); Salton Sea, California (340°C); Norris Basin, Yellowstone Park (205°C); Upper Geyser Basin, Yellowstone Park (180°C); and Pauzhetsk, Kamchatka, U.S.S.R. (195°C). Of these 19 drilled geyser areas, the lowest maximum subsurface temperature is 150°C, and only 4 areas have maxima less than 170°C. Of these 4, all were drilled only to a few hundred feet in depth, and none is known to have attained its "leveling off" temperature. These data suggest that deep subsurface temperatures of at least 150°C and probably 170°C are necessary to support natural geysers near the surface.

The highest temperatures tend to occur within the central core of each system, surrounded by zones of lower temperature (Banwell, 1957; Elder, 1965, figs. 13-19). The distribution patterns of temperature cannot be explained by static systems but demand active convection, with hot water rising in the core of each system because of its lower density and mushrooming outward near the surface. Outside the central core, cooler water of higher density descends along any available channels. Because of these relationships, wells drilled near the margins of each system commonly attain their highest temperatures at intermediate depths, with decreasing temperatures at greater drilled depths. The geyser-bearing convection systems are large and deep—much deeper than was suspected prior to geothermal exploration. Ten of the 19 geyser areas listed above have been drilled to depths from 2000 to more than 8000 feet, with no evidence that the base of convective circulation was attained in any one of them!

This concept of upflow of hot water as parts of huge convection systems was appreciated by Benseman (1965) and was anticipated at least in part by Barth (1950) and Golovina and Malov (1960).

Several factors favor deep circulation in geothermally anomalous areas. The most important is density difference related to thermal expansion. The differences are not linear but increase greatly at higher temperatures. Relative to water of density 1.000 at 4°C, pure water at 45°C is 1 percent lighter; and, correspondingly, at 66°C, 2 percent; 111°C, 5 percent; 168°C, 10 percent; 250°C, 20 percent; and 344°, 40 percent. With a "normal" geothermal gradient, only a fraction of 1 percent of difference in density of water can be expected between the Earth's surface and 1000 feet in depth. But in an anomalous area where temperature differences can exceed 200°C, thermal expansion becomes a potent factor, increasing almost directly with total depth involved but exponentially with temperature difference.

A second factor is the decreasing viscosity of water with increase in temperature. The viscosity of pure water is 0.016 poise at 4°C but only 0.0016 at 170°C. Decreasing viscosity may therefore help to offset decrease in permeability which may occur at depth in a thermal system.

A third factor fostering deep circulation is the large increase in solubility of silica with increasing temperature (Fournier and Rowe, 1966). As cold water first circulates down on the margins of a developing convection system, movement along fractures, faults, and interconnecting pore spaces may be extremely slow, especially initially in the so-called "impermeable" rocks. But mass permeability of such rocks is not completely nil; as water moves through tiny channels and is heated, silica is dissolved from quartz and other minerals, thereby increasing the permeability of a given channel with time.

Most students of geyser activity have assumed that the water is dominantly of surface origin but of relatively shallow circulation (depth not specified), being heated from below by hot rising volcanic gases. If so, a hole drilled into rocks beneath the zone of shallow circulation should show continuously increasing temperatures down to the magmatic source. Moreover, such a hole should either intersect channels bearing the assumed volcanic gases or should yield interstitial fluids of some other origin but *not* liquid water almost identical in temperature, chemical composition, and concentration to the shallow water, assumed in such models to be the product of mixing of surface water and volcanic steam. Deep drilling in the cited geyser areas has consistently found only slight changes in temperature in deeper parts of the systems, and the deep waters are typically very similar chemically and isotopically to the near-surface geyser waters. The minor differences that do occur are consistent with the physical changes and hydrothermal reactions between the fluids and rocks of each system. No drill hole has yet encountered vapor with a temperature or a chemical composition consistent with a pure volcanic origin.

Origin of the water.—Isotopes of the waters of some geyser-bearing hydrothermal systems have been studied, including Wairakei, New Zealand, several Icelandic areas, Steamboat Springs, Salton Sea, and Yellowstone and Lassen National Parks (Craig, 1965; White, Craig, and Begemann, in press). These studies have consistently demonstrated that the D/H and O^{18}/O^{16} ratios of the cold surface waters of each area differ from each other depending on latitude, distance from the ocean, and other climatic factors. The D/H ratios of waters from hot springs, geysers,

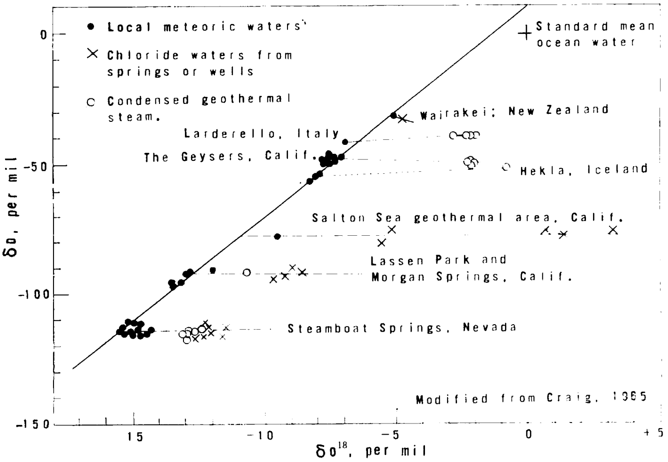


Fig. 2. Observed isotopic variations in near-neutral chloride-type geothermal waters and in geothermal steam.

and deep wells are almost identical to those of the cold surface waters of each particular area, but the thermal waters are generally enriched in O^{18} relative to O^{16}. Craig's plot of the data (1965, p. 33) is here reproduced as figure 2. The O^{18}/O^{16} shift with no significant change in D/H can be explained as mixing of meteoric and magmatic waters high in O^{18} only if *each* magmatic water happens to have the same D/H as the particular meteoric water with which it mixes. Because the meteoric waters are isotopically dependent upon their particular climatic environment, this possibility is extremely unlikely. A far better explanation, consistent with experimental and other evidence, is that each meteoric water circulates to great depths and, at high temperature, reacts with silicate and/or carbonate rocks, both of which are strongly enriched in O^{18} relative to the meteoric waters.

The isotope evidence does not exclude the possibility of a small contribution of water of volcanic or other deep origin, but in general we should see clearly evident isotopic effects if the mixing nonmeteoric water is as much as 10 percent of the total. The problem then becomes: How much volcanic water can a system contain and still escape detection isotopically? Craig concedes (1965, p. 34-35) that 1 or 2 percent could escape detection, and I have assumed an upper limit of 5 percent as a general rule.

Some details of a large convection model, with effects of dissolved gases.—On figure 1, let us now trace the path of meteoric water, initially cold, that circulates to depths of 1 or 2 miles in an area of very great heat supply. The model is greatly oversimplified, and it assumes a steady-state heat flow permitted by convection in a magma chamber (White, 1957; in press).

Tracing a single path of flow instead of the countless paths of a natural complex system, cold surface water recharges the system at point A (temperature shown on graph, with depths common to both parts of the figure); it migrates downward to point B because of its relatively high density. It then flows to point C, dissolving SiO_2 and other constituents as it is heated to 250° C by thermal conduction through rocks with very high temperature gradients, assumed for simplicity to be straight-line gradients, as in line GFC. For magma chambers and convection systems that are not horizontally infinite, actual temperature gradients are not linear. Also if some hot fluid is escaping from the magma chamber, gradients would be distorted from linearity, perhaps similar to the dashed curve GF′C.

Because pressures are so high at point C, the circulating fluid is a single liquid phase consisting of water with dissolved salts (dissolved gases are neglected for the moment). The hot liquid rises in the core of the system from point C, with a relative density close to 0.80. Only a slight change in temperature occurs between C and D if the rate of upflow is high, because a system with an age of 10,000 years or more (White, 1965) has already heated its wall rocks, and rocks are good thermal insulators.

At point D, the hydrostatic pressure has decreased to the vapor pressure of water at 250°C, so the first steam bubbles start to form. From point D to the surface at E, hydrostatic pressure declines continuously, so more steam forms by transfer of excess energy from the still-liquid water.

Now let us retrace the same circulation path, considering the effects of dissolved gases. Differences are negligible to point C, where volcanic gases may be flowing up from a magma. If the pressure at C is 300 bars and the temperature of mixture is 250°C, the data of Takenouchi and Kennedy (1964, p. 1059) suggests that about 4.2 mole percent or almost 10 weight percent of CO_2 could be dissolved in liquid water. The effect of dissolved salts in decreasing the solubility of gases is small if the salt content is less than 1 weight percent, as in most geyser waters (Takenouchi and Kennedy, 1965, p. 448). If volcanic water accounts for less than 5 percent of total water, as favored by the isotopic data and if CO_2 does not exceed 20 percent of the volcanic fluid (White and Waring, 1963), the content of CO_2 in the meteoric-dominated mixture is only about 1 percent, which is far below saturation.

Water with dissolved gases rises from C to D′ where the first vapor bubbles form. D′ is slightly deeper than D, the depth of first boiling for pure water, because of the vapor pressure of the dissolved gases; the depth of D′ below D depends on the composition and total content of gases as well as the temperature of the rising water.

If we assume a temperature of 250°C and 1 weight percent (or about ½ mole percent) of CO_2 in the water, D′ should be at a depth of about 3000 feet (pressure near 60 bars; Takenouchi and Kennedy, 1964, figs. 4 and 5) or about 1000 feet deeper than for pure water. The first bubbles contain about 20 percent of CO_2 and are therefore much enriched. As the water rises further and pressure continues to fall, more CO_2-enriched vapor is formed, accompanied by a decrease in temperature because of latent heat of vaporization and expansion of already-formed gas bubbles. As point D is approached, temperatures are slightly below the reference curve, more H_2O is vaporized, and the remaining liquid becomes increasingly impoverished in CO_2. This is essentially an isothermal polybaric distillation process, in which the composition of liquid and vapor phases changes continuously. The total amount of vapor that forms below 2000 feet is small compared to the amount that forms at shallower depths.

Actual temperatures measured in drill holes at Steamboat Springs, especially between depths of 200 to 400 feet, are completely consistent with the above explanation of control by the boiling point curve, modified by the effects of dissolved gases (White, in press, figs. 41 to 43). At shallower depths, conductive heat flow and other factors become important because drill holes are never sited in the vents of geysers or boiling springs but instead are sited on firm rock where surface temperatures are close to atmospheric.

Importance of geochemistry of SiO_2.—The geochemistry of silica has been clarified greatly in recent years, with important applications to geyser theory. All areas of natural geysers are characterized by water containing at least 250 parts per million of dissolved SiO_2. Fournier and Rowe (1966), Mahon (1966), and White, Brannock, and Murata (1956) have shown that the silica content of natural waters is closely related to temperature, especially above about 170°C where equilibrium with quartz seems frequently approximated. At lower temperatures, equilibrium with quartz becomes increasingly sluggish, and cooled thermal waters are commonly strongly supersaturated with respect to quartz; little or no deposition of SiO_2 may occur until the solubility curve for amorphous SiO_2 is attained. At 100°C, for example, the solubility of quartz is 50 parts per million, but that of amorphous silica is 385 parts per million (Fournier and Rowe, 1966). A silica content of 250 parts per million, which is near the minimum for natural geyser waters, suggests a subsurface temperature of at least 180°C in the system, assuming equilibration with quartz. In general, subsurface temperatures estimated from the SiO_2 contents of discharged waters agree closely with actual temperatures recorded in drill holes (Mahon, 1966).

However, SiO_2 is deposited in the upper cooler parts of the high-temperature systems. At the surface, amorphous opaline siliceous sinter forms; as the waters cool and as dissolved matter is concentrated by evaporation, the solubility of amorphous SiO_2 is exceeded. The beaded type of sinter called geyserite forms around the throats of geysers and perpetual spouters where complete or extensive evaporation occurs (White, Brannock, and Murata, 1956). Similar amorphous opal forms deeper in the throats of geysers, especially during the steam stages immediately following eruption when water levels are low and evaporation of surface films occurs (see later section). Amorphous opal is also a very common filling of pore spaces of near-surface rocks in hot-spring areas, as shown by study of drill cores and outcrops of eroded rocks. Porous sands, gravels, volcanic rocks, and previously deposited primary sinter are thereby converted into rocks of much lower permeability. Amorphous opaline sinter and silica glass both have about the same solubilities as silica gel (about 250 ppm at 70°C according to R. O. Fournier, oral communication, 1966). Zeolites, K-feldspar, and chalcedony are other hydrothermal minerals commonly deposited in pore spaces at depths and temperatures generally somewhat above those for amorphous opal. The near-surface effects of this mineral deposition is to decrease porosity and permeability, tending to confine the upflow of hot water into the most permeable channels. Even these channels are filled in time, eventually impeding circulation until new channels are formed or old channels are reopened by faulting or by buildup of hydrostatic pressure in the upper part of the system.

Origin of geyser tubes and associated cavities.—The nearly neutral to alkaline high-silica waters are likely to be saturated with respect to silica and silicate minerals; they are very unlikely to create solution cavities near the surface. Some acid waters that form by oxidation of

H_2S (White, 1957) can dissolve some minerals near and above the water table and can thereby create pore spaces, but careful study of these processes and of drill core gives no evidence that large chambers assumed for many geyser models do form by dissolution below the water table.

In addition, geyser tubes and large interconnected chambers are unlikely to be inherited from normal pre-geyser environments. The most appealing explanation for geyser tubes and adjacent chambers is that they form by physical ejection of rock fragments as a new geyser develops from a facture or some other interconnecting channel. Ejection of mud, sand, and rocks has been observed when a new geyser first develops, as in Seismic Geyser of Upper Basin, Yellowstone Park (formed in 1963 and increased greatly in vigor through 1966, according to Dr. George Marler, oral communication), or when an old geyser increases in vigor. The large voids are then *effects* rather than *causes* of geyser action.

Furthermore, many geothermal wells drilled randomly to depths of only 50 to 200 feet in high-temperature ground of limited permeability erupt intermittently as man-made geysers (White, in press; Vymorokov, 1960). In initial eruptions, such a well frequently ejects large quantities of rock and sand and is evidently forming "voids" by enlarging the drill hole in susceptible zones. If such wells are drilled deeper and encounter additional high-temperature permeable channels, the wells then erupt continuously rather than intermittently. With time, many continuously erupting wells evolve into geysers with intermittent discharge. In some of these, the change is clearly related to deposition of $CaCO_3$ or SiO_2 in the drill pipes or in aquifers adjacent to the wells.

The plumbing system of a geyser evidently consists of porous rocks adjacent to interconnecting channels. Because individual channels are normally filled in time by mineral deposition, all geysers should eventually change their behavior; some cease erupting and other new geysers come into existence.

Superheated and supersaturated waters.—Allen and Day (1935, p. 222) and Bloss and Barth (1949, p. 883-885) recognized the frequent occurrence of superheated water at and near the surface of many springs and geysers in Yellowstone Park; temperatures are commonly 1° to 3°C above the boiling point of pure water; such waters are metastable and are not in equilibrium with their environment. Superheated water is particularly common in the more alkaline geyser areas, where gas contents are low (Allen and Day, 1935, p. 90; unpub. data). In contrast, the waters of some other geyser areas are neutral to slightly acid and their content of gases, particularly CO_2, generally increases with increasing acidity (Allen and Day, 1935; Iwasaki, 1962). The dissolved gases lower the temperatures of vapor formation, in some instances by 35°C or more. These neutral or acid waters are also normally supersaturated or "superheated" with respect to their actual gas contents. But, in contrast to the alkaline waters, a temperature measurement alone does not prove supersaturation. Fortunately a simple field test can identify waters that are superheated by as little as 1°C, and this test is actually more diagnostic than temperature measurements unaccompanied by detailed data on the dissolved gases: If the introduction of a thermometer or a few grains of sand produces a local increase in "boiling" near the foreign objects, the water is supersaturated; the foreign objects provide sites for nucleation of excess gases or steam. The physics of the phenomenon has been considered by Marboe and Weyl (1948) and Marboe (1949). The boiling point of pure water is useful only in identifying the superheated waters that are lowest in gas content.

Supersaturation occurs commonly in thermal and mineral waters discharged at the surface, and it plays an important part in the activity of many geysers as will be seen. Thorkelsson (1940), Bloss and Barth (1949), and Barth (1950) maintained that dissolved gases were essential for geyser eruptions, rapid evolution of the supersaturated gases providing the trigger for initiating the eruption. Supersaturation and a triggering action by gases is indeed likely, but supersaturation also occurs in a pure-water system but at a slightly higher temperature.

Supply of energy for eruption.—The rising hot waters deep in the geyser-bearing thermal systems, at temperatures of 150°C and higher, contain abundant energy for geyser eruption. Such systems are considered worldwide to be prime prospects for geothermal energy (Banwell, 1963). A successful producing well is similar to a geyser except that it erupts *continuously*, with part of the water flashing into steam as pressure decreases. Formation and expansion of steam propel the steam-water mixture to the surface, where the steam is then separated and utilized in power plants to generate electricity. The potential mechanical energy stored above 100°C in liquid water at 150°, 200°, and 250°C is equivalent to 5350, 10,900, and 16,800 foot-pounds per pound of water (White, 1955, p. 1124).

Liquid water at 150°C, for example, must have a containing pressure of at least 69 pounds per square inch in order to remain liquid. But if the pressure is suddenly dropped to atmospheric, as in convection or ejection of such water to the surface, 9.4 percent of the liquid flashes into steam as the temperature drops to 100°C (White, 1955, p. 1124); the total volume of a pound of water suddenly expands from 0.017 ft^3 to 2.55 ft^3 of steam and water, an increase in volume of about 150 times. Most of the mechanical energy available from each pound of such water is utilized in expanding the steam against air pressure, but less than 5 percent of this energy could sweep the associated water upward to a height of 200 feet in a major geyser eruption. In a convecting boiling pool, in contrast, nearly all the mechanical energy is utilized in expanding steam against air and water pressure, and none is used in ejecting water above the water table.

Stable and unstable systems.—A high-temperature thermal system that discharges water at the surface is inherently unstable because relatively low temperature water (high in density) is situated above high-

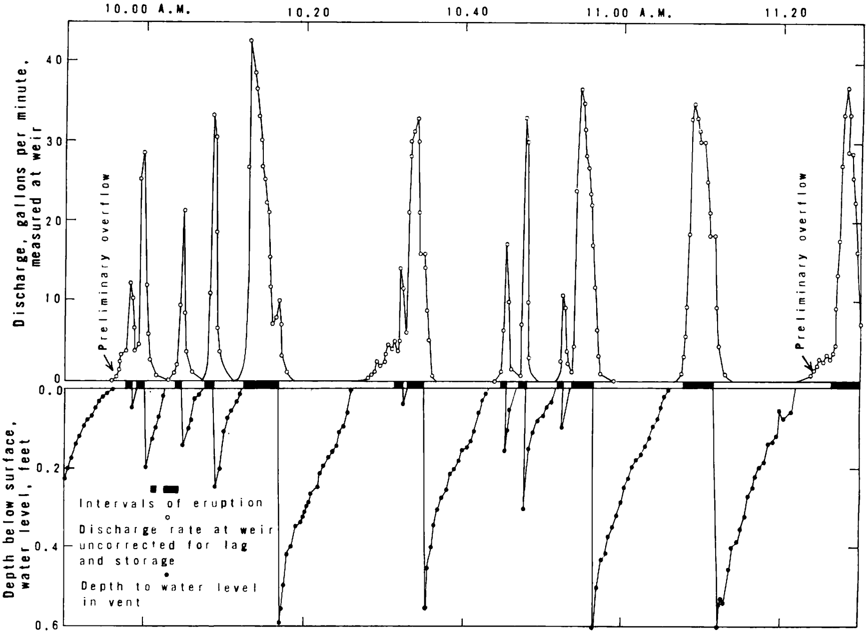


Fig. 3. Graph of discharge and water levels of Geyser 23-n, June 22, 1946, showing different eruption magnitudes.

temperature water (lower in density but higher in energy content). Thermal instability and density contrasts are especially pronounced in the upper 200 feet of the systems because of the convex shape of the boiling point curve with depth. Most hot spring systems evidently have many interconnected channels, and these channels are likely to enlarge upward, just as the porosity of sediments or of a fault breccia is likely to increase toward the ground surface. For these reasons, secondary convection cells are abundant near the surface wherever temperature contrasts and permeable interconnections permit; in such cells the excess energy can be dissipated in steam. An inherently unstable system is thereby stabilized, so that geyser eruptions either do not occur or the frequency of eruption is much less than if convection were absent.

NATURAL GEYSERS OF STEAMBOAT SPRINGS

As stated previously, 21 different natural vents at Steamboat Springs are known to have erupted as geysers (White, in press, table 4). Most of these were small, erupting to heights of only 3 feet or less, and most were active irregularly over periods of a month to several years, with other intervals of temporary or even permanent quiescence. Geysers were most numerous during periods of high water level and high discharge, especially during 1945 (7 known as active) and in 1950 (11 known as active). Some of these geysers erupted so infrequently that little was learned about their behavior. Three of the 21 natural geysers are described briefly here and the others are tabulated by White (in press, table 4).

Geyser 23-n.—Most geysers erupted so infrequently or the individual eruptions were so small that the discharge of the whole spring system was not affected. One exception was a small geyser, 23-n, near the crest of the Main Terrace. Figure 3 shows the differences in behavior that sometimes occurred during a single series, when total discharge of individual eruptions ranged from 7 to 90 gallons. The weakest eruptions were consistently followed by only slight declines of water level and short intervals to the next eruption; the strongest eruptions were followed by the greatest declines of water level and the longest intervals to the next eruption. The main uncertainty concerns the magnitude of the next eruption. The average rate of discharge during each cycle tends to be approximately constant.

In figure 4, curve B shows the *average* rate of discharge for each daily series of measurements, and a comparison with curve A demonstrates a marked influence of barometric pressure; at times of low pressure, discharge generally increased, and vice versa. The reasons for the barometric influence and the changes that occurred in December are considered in detail by White (in press).

This small geyser varied notably in activity during the period of figure 4. During 5 of the first 7 series of measurements through November 10, the spring discharged intermittently but with no turbulence or vigorous ejection of water. From November 16 through December 6 it

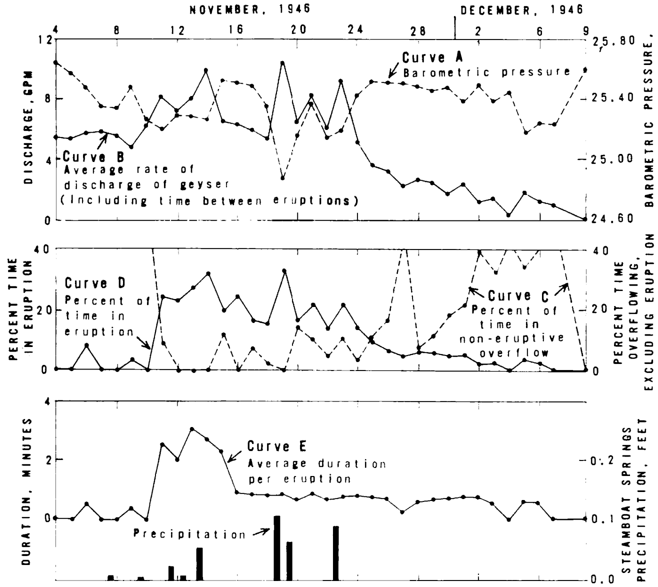


Fig. 4. Data from Geyser 23-n, Main Terrace, November and December 1946.

normally erupted for slightly less than 1 minute per eruption (curve E) to heights of 6 inches to 1 foot; intervals between eruptions generally ranged from 5 to 20 minutes. But from November 11 through 15 the eruptions were more vigorous, to heights of 1 to 2 feet, and frequently were not preceded by preliminary overflow (curve C). The mixing of different magnitudes, as shown in figure 3, generally did not occur during any one day, but some of the more vigorous eruptions started on a smaller scale before "shifting gears" to the more vigorous type, as if a deeper level had been triggered into action.

The average rate of discharge of a geyser presumably could increase in several different ways, as by increasing the vigor of eruption (rate of discharge per unit of time in eruption), by increasing the frequency of eruptions (shortening interval between eruptions), by increasing the duration of each eruption, or by increasing the overflow occurring between eruptions. For geyser 23-n during the graphed interval of figure 4, the average rate of discharge (curve B) correlates most closely with percent of time in eruption (curve D), with some influence by non-eruptive overflow (curve C). The longer eruptions evident in curve E are compensated largely by longer intervals between eruptions.

Precipitation influences the discharge of this particular geyser in a manner similar to that of the whole spring system (White, in press), but earth tides and wind velocity are probably minor factors.

Geysers 12 and 13.—Vent 13 was an active geyser during the first month of detailed study in 1945 and was active again at intervals from 1950 to 1952. Efforts were made in 1945 to measure the discharge through a V-notch wier, but occasional more vigorous unobserved eruptions washed out the wier. One such eruption left a high-water mark indicating a rate of discharge in excess of 100 gallons per minute. Duration of the usual minor eruption was about 40 seconds, the eruption interval was in the order of 10 minutes, and maximum height of eruption was only 1 to 1½ feet.

Early in 1950, vent 13 was again an active geyser, now erupting to 10 to 12 feet in height with a duration of about 1½ minutes, an eruption interval of about 30 minutes, and estimated discharge of about 100 gallons per eruption.

In January, 1950, vent 12 became the most vigorous natural geyser that I observed at Steamboat Springs, erupting to a maximum height of 25 feet. During the next two months it alternated with vent 13 in its active periods. At other times vent 12 was a continuously discharging spring or spouter with variable discharge. Systematic observations were not made, but interconnections between the two geysers and adjacent vents were proved by fluorscein as a tracer and by responses in water levels.

DRILLED WELLS THAT DISCHARGE AS GEYSERS

Many wells have been drilled in the Steamboat Springs thermal area to supply hot water for the resorts; others were drilled for geothermal energy or for research. Some wells and drill holes were observed closely as drilling progressed, to clarify subsurface conditions (White, Thompson, and Sandberg, 1964, table 3; White, in press, fig. 4). Of these, the following have discharged as geysers: Rodeo Well, Mount Rose (1947) Well, Steamboat Well 4 (on a few occasions), GS-3, and GS-4 diamond drill holes. At least 8 other drilled wells of the area have behaved as geysers.

Few geyser studies except those of Nomura (1954), Noguchi (1956), and Vymorokov (1960) have utilized detailed data from drilled wells, although the existence of geyser action in some at temperatures below the boiling of pure water was noted by Henderson (1938), Barth (1950, p. 72), White and Roberson (1962), Iwasaki (1962, table 1), Mal'tsev and Bubol'ts (1961), and Sappa (1955). Because the detailed plumbing system of natural geysers is unknown and access for temperature measurements is normally limited to the part of the geyser tube that is vertically below the vent, much can be learned from geysering wells. This is especially so if the nature of the subsurface rocks, the structural control, the total and the cased depths, and the natural ground temperatures are determined as drilling progresses.

Each well drilled in or near the more active parts of the Steamboat Springs area shows an upper zone of rapidly increasing temperature with depth; an intermediate zone, generally from about 100 to 300 feet in depth, where temperatures are very close to but a few degrees below the boiling-point curve for pure water; and a deep zone (if drilled deep enough), where temperatures level off at about 170°C, with no further increase at greater depth. Local reversals in temperature at higher levels are caused by inflowing tongues of dilute water or by local convection.

THE GEYSER WELL

Physical characteristics and setting.—An abandoned well previously described as number 32 Geyser Well (White, in press), but hereafter called the Geyser Well, was studied in considerable detail. The Geyser Well is shown on figure 5 relative to nearby fissures and natural vents. The well reportedly was drilled about 1930 to a depth of 70 to 74 feet, but the hole was crooked and was eventually abandoned (John Czykowsky, oral communication, November 1945). A 6-inch-diameter iron casing extended down from the ground surface to about 35 feet. The total effective depth through the period of study was 43.1 feet below ground surface. The bottom of the well, as probed by steel tape and heavy weight, was firm and almost horizontal, indicating the complete filling of the lower 30 feet of the well, as reportedly drilled. Frequently a thermometer or other instrument, while being raised or lowered into the well hung up at specific levels, so the hole was probed with a long stiff wire hooked on the end. The results are interpreted in figure 6, utilizing additional data from Steamboat Well 5, a 5-inch diamond drill hole inclined 32° from the horizontal toward the Geyser Well (shown in fig. 5), and drilled in 1950 to a depth of 46 feet (30 ft below ground surface).

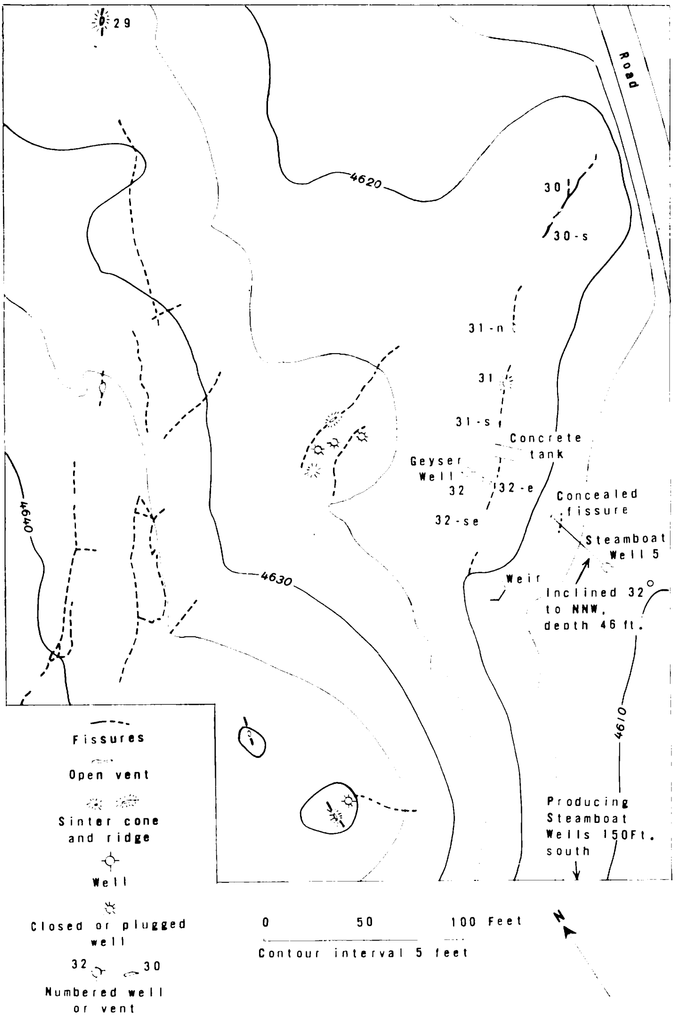


Fig. 5. Map of the Low Terrace near the Geyser Well, showing relation to fissures and natural vents.

General behavior June 1945 to March 3, 1946.—The Geyser Well was very active during 1945 and early in 1946, with major eruptions generally occuring at intervals of 5 to 7 days (table 1). The duration of eruption was usually from 30 to 40 minutes, and the maximum height was 60 to 75 feet. Water level immediately prior to eruption ranged from 0.5 to 2.6 feet below the top of the casing (ground level), with differences mostly related to changes in behavior of the Steamboat wells (White, in press).

Usually after an interval of 5 to 7 days following a previous eruption, the Geyser Well either erupted naturally, or, if disturbed even slightly, an induced eruption followed within a few minutes (see comments, table 1). For both natural and induced eruptions, after initial mild boiling the evolution of vapor bubbles increased, at first with no net change in water level but eventually with an abrupt rise until the top of the water column discharged at the ground surface. The geyser then surged rapidly upward into eruption, generally attaining its maximum height within 30 seconds after the initial overflow. The first 3 to 5 minutes of eruption were characterized by behavior here called the

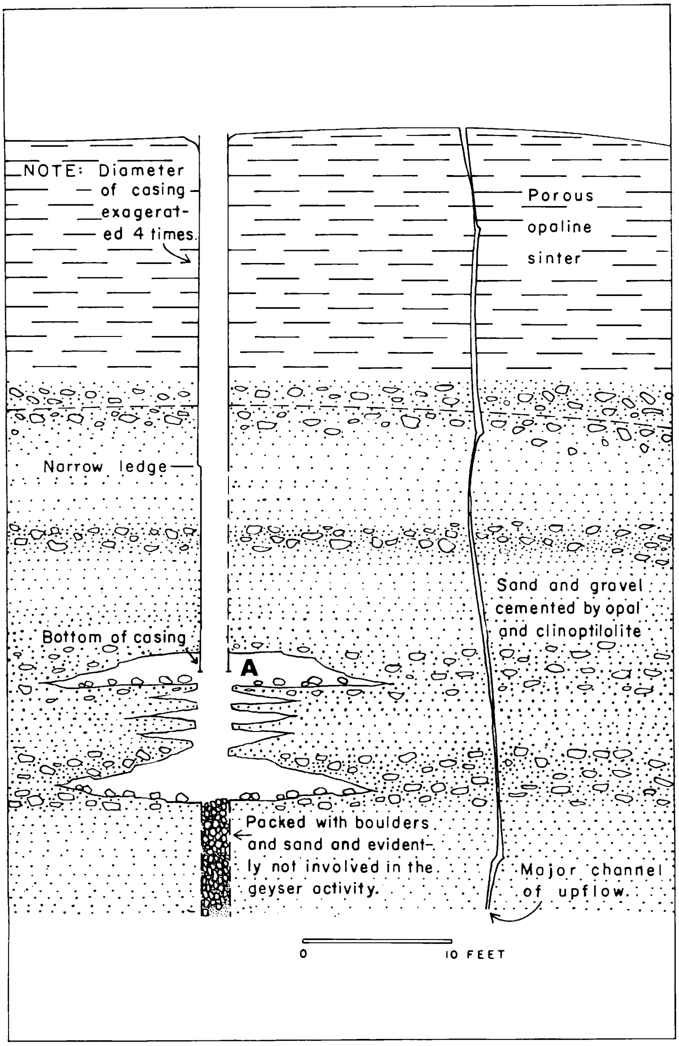


Fig. 6. Section through the Geyser Well, incorporating available data and hypothesizing some enlargement of cavities near effective bottom of the well.

TABLE 1

Representative eruptions of the Geyser Well, Steamboat Springs, Nevada

Eruption Date	Eruption Hour	Interval since last eruption, days	Water level prior to eruption, feet	Comments
1945				
June 8	11:50 a.m.	?	1.5	Induced unintentionally; ∼ 2 min of increasing boiling after temp and water sample; max height 65 ft; duration 39 min; violent steam stage 6 min, passive stage 21 min.
12	8:11 a.m.	4	1.6	Induced unintentionally by tapping casing with hammer to obtain level reference point; water started to boil, first gently, then vigorously, with eruption in ∼ 3 min. Max height 65 ft, max Cl 968 ppm.
	9:11 a.m. 10:01 a.m.	27:05 min 30:00 min	....	Passive steam stage that did not cease for 22 min; gradually rising surges finally overflowed, led to 2d eruption (induced by water-level measurements?); 3rd and 4th eruptions similar (max height of 4th, 60 ft; max Cl 1020 ppm. Successive durations of eruption 33, 20, 20¼, 21¾ min.
	10:55 a.m.	33:45 min	....	Might have erupted again but 1 gal cold water added.
20	Evening	8	1.6	Natural eruption.
23	9:48 a.m.	3	1.6	Induced by tapping casing to confirm June 12 result. Duration 33½ min; violent steam stage 5 min, passive stage 20¾ min.
July 7	P.M.?	7?	1.5	Natural.
11	3:00 p.m.	4	1.6	Induced unintentionally by drilling reference marker in sinter 5 ft from well.
27	8:45 a.m.	9	1.3	Induced by inserting thermometer.
Aug. 6	11:55 a.m.	5	1.2	Induced by thermometer.
9	9:15 a.m.	3	1.2	Induced; light tapping of casing with wooden stick followed in ∼ 3 min by eruption. Duration 29 min.
Sept. 23-24	3:45 p.m.	5?	0.5	Natural; series of eruptions like June 12; Steamboat Well 2 being cleaned.
Oct. 3	P.M.	9	1.3	Natural; Steamboat Well cleaned to bottom on Oct. 2.

	11	~ 4:00 p.m.	3	1.6	Induced by handful of gravel.
	16	12:58 p.m.	5	1.7	Induced unintentionally by thermometer at −20 ft; max Cl 1000 ppm (initially ~ 888 ppm). Duration 40 min.
	23	10:44 a.m.	7	1.9	Induced by thermometer at −5 ft; max height 75 ft, no wind. Duration 31 min.
	28	~ 3:00 a.m.	4¾	1.6	Natural; shortest natural interval observed; see fig 10.
Nov.	7	1:18 p.m.	10	1.5	Induced by thermometer at −5 ft; after removal, intermittent boiling, increasing in vigor, eruption after 3½ min; Cl content from 880 to 972 ppm. Bottom temperature 103.5°C 3½ min after eruption terminated. Long interval caused by heavy rain (fig. 10). Duration 41½ min.
	14	3:10 p.m.	7	1.4	Induced by thermometer at −1 ft; eruption 2 min after removal. Barometer falling rapidly, slight wind. Duration 44 min.
	20	12:11 p.m.	6	1.6	Induced. Total discharge gaged 985 gal, but at least 30 percent lost because of wind. Duration 40 min. See figs. 8 to 10 and table 2.
	26	11:51 a.m.	6	1.6	Induced by thermometer at −25 ft. Duration 36 min; violent steam stage 5 min, passive stage 15 min.
Dec.	21	~ 9:00 a.m.	17	1.0	Natural; long interval caused by precipitation of Dec. 23-25.
	28	9:22 a.m.	7	0.8	Induced by thermometer at −22 ft; eruption in 2 min after removal; 2 qts cold water failed to stop subsequent boiling. See effects of precipitation, fig. 10.
1946					
Jan.	13	Late P.M.	11	1.2	Natural. Increasing depths to water from random change unrelated to Steamboat Wells.
	27	3:18 p.m.	3½	2.2	Induced by thermometer at −43.1 ft; total discharge ~ 1350 gal; max Cl 1064 ppm. Duration 39 min. Total steam stage 25 min.
Feb.	6	8:45 a.m.	4	2.4	Induced after thermometer very gently introduced to −1 ft. Duration 35 min.
	12	9:28 a.m.	6	2.4	Induced; 2 min after removal 1 pt water sample. Duration 33½ min.
	23	12:45 p.m. 1:40 p.m.	4 ~ 20 min	2.5	Natural; at least 2 eruptions, possibly more. Steamboat Well 2 partly cleaned on Feb. 22.
Mar.	2-3	~ 11:00 p.m.	7	2.6	Natural; series of eruptions like June 12, 1946 for 16 hrs after Steamboat Wells 2 shut down on March 2; cleaned to bottom on 3rd. Exchange of function had occurred by March 4 with vent 31 boiling vigorously and Geyser Well lower in gas content (few bubbles induced by thermometer).

TABLE 1 (Continued)

Eruption		Interval since last eruption, days	Water level prior to eruption, feet	Comments
Date	Hour			
April 16	~ 1:00 p.m.	44	2.3	Natural.
June 18	8:35 a.m.	46	1.3	Induced by throwing in pebbles.
1947				
Feb. 16	11:10 a.m.	225	0.3	Induced; eruption typical for ~ 40 min, then spurting 6 to 10 ft for ~ 3 hrs. Steamboat Well cleaned Feb. 18-19; vent 32-e becomes small natural geyser.
April 20	~ 12:30 p.m.	65	0.9	Natural. See figs. 10, 11.
July 21	~ 8:30 a.m.	92	1.3	Natural. See fig. 10.
Aug. 11	11:20 a.m.	8	1.5	Induced to collect erupted sample for analysis; height 50 ft during bursting stage, then generally 10 to 20 ft; bottom temp 104.9°C just before end of eruption. Steamboat Well cleaned August 12 and 25. Duration 93 min; violent steam stage 7 2/3 min, passive, 22 min.
Sept. 7	~ 11:30 p.m.	27	1.5	Natural; triggered by strong local earthquake; eruption either very long or series of 2 or more. Steamboat Well 4 drilled Sept. and Oct., cleaned Feb. 12 and 17, 1948. See fig. 10.
1948				
Mar. 23	11:29 a.m.	27	1.2	Induced; soaping ineffective but rapid withdrawal of plunger caused overflow, eruption. Bottom temp 103.3°C at end of eruption. Duration 185 min; violent steam stage 11½ min, passive, 23½ min.
April 1948 to July 1949				Very long intervals, generally ~ 150 days.
August 1949 to June 26, 1950				Long intervals (>100 days) when water level low, 4 to 9 days if water level rose above −0.2 ft (starting to discharge through corrosion leak in casing −0.15 to −0.2 ft).
June 27, 1950 to Aug. 1952				Short intervals 1/3 to 2 days when water level high enough to permit overflow, alternating with long intervals when water level low (controlled by production cycles of Steamboat Wells; White, in press). See fig. 12.
Sept. 1952 and later				Not systematically observed. About 1955 Steamboat Well 6 drilled 100 ft nearer to Geyser Well than wells 2 to 4, causing cessation of activity.

PLATE 1.

Geyser Well in early bursting stage of eruption, June 29, 1950.

"bursting" stage (as in pl. 1), when water supply and back pressure were evidently so high that superheated water much above the surface boiling point of 95°C was ejected. With emergence of this superheated water from confinement in the well casing, rapid flashing of additional water to steam could occur. After this initial bursting stage, water and steam were then ejected vertically with only gradual horizontal expansion above the ground surface. At the end of the bursting stage the rate of discharge and height of eruption decreased, eventually tapering off in 30 minutes or so, with individual upward spurts finally failing to reach the ground surface. Discharge then ceased, terminating the eruption.

The steam phase which followed an eruption consisted of an early violent stage of about 5 minutes duration followed by a passive steam stage (terminology of Fix, 1939, is used wherever possible). When the well was probed by an instrument responsive to water movement, such

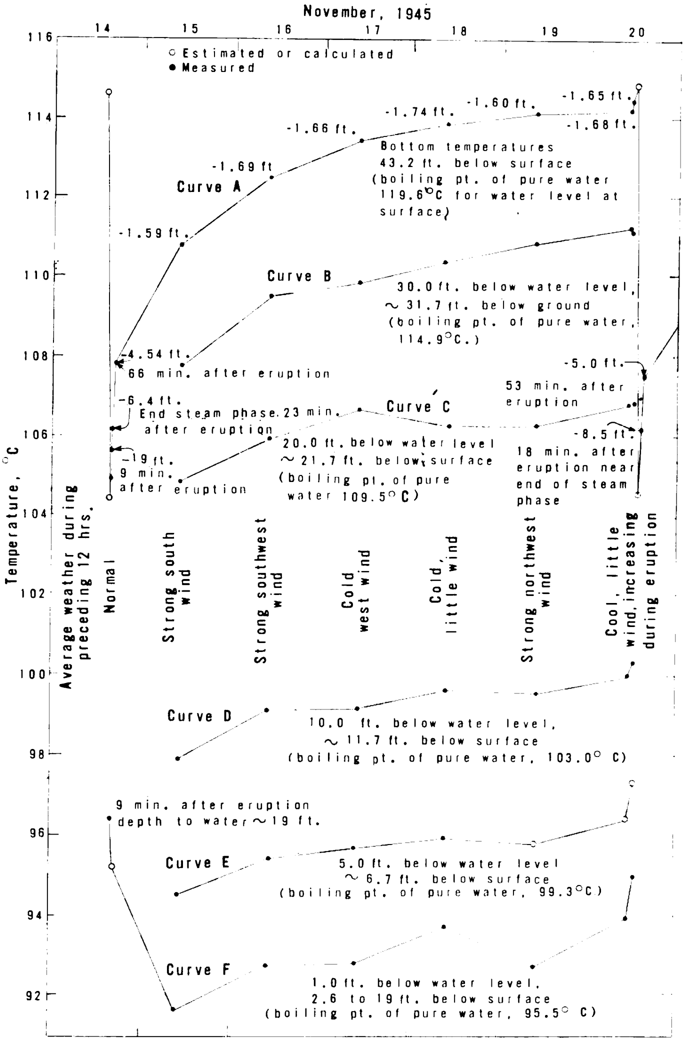


Fig. 7. Temperature-time curves for six different depths in the Geyser Well, November 14 to 20, 1945.

as a suspended water sampler, the violent steam stage was found to be characterized by vigorous upward surges, presumably identical to the final spurts of the eruption proper except that liquid water normally did not reach the surface; steam and water were evidently still turbulently intermixed, with steam as the continuous phase. About 5 minutes after eruption had ceased, subsurface noises rather abruptly decreased in intensity, and probing then identified a rather well-defined water level that surged moderately up and down a foot or so. Now, the continuous phase evidently was liquid water in which the rising bubbles of steam were dispersed. Duration of this passive steam stage was generally about 20 minutes (table 1), and its termination became evident by an abrupt cessation of noise. A nonsurging water level could then be identified, generally at a depth of 5 to 9 feet.

Detailed data of November 14 to 20, 1945.—Many temperature profiles and other data were obtained from time to time in the Geyser Well, but an extensive series was made during an eruptive cycle that started November 14 and ended November 20. The temperature data, obtained by a standardized maximum-recording mercury thermometer, are shown for six different depths in the well (fig. 7). After the November 14 eruption, bottom temperatures (curve A) recovered along a uniform curve, except for a final upsurge on November 20 which will be described. With decreasing depth, curves B through F became more irregular. Curve F from 1.0 foot below water level is affected by air temperature and wind velocity as is evident on November 17 and 19. Even when no gas bubbles were breaking the surface, close observation detected slight current movements, indicating convection in the well.

A double series of temperature measurements were made on November 17 and again on November 20, with results shown on figure 8. On November 17 when the first series (curve A) was measured from top to bottom, the rate of evolution of gas bubbles increased temporarily as the thermometer and steel tape were lowered to each deeper level. At 30 feet below water level, rather vigorous boiling occurred. Somewhat less sensitivity was found at greater depths, and at each new point the

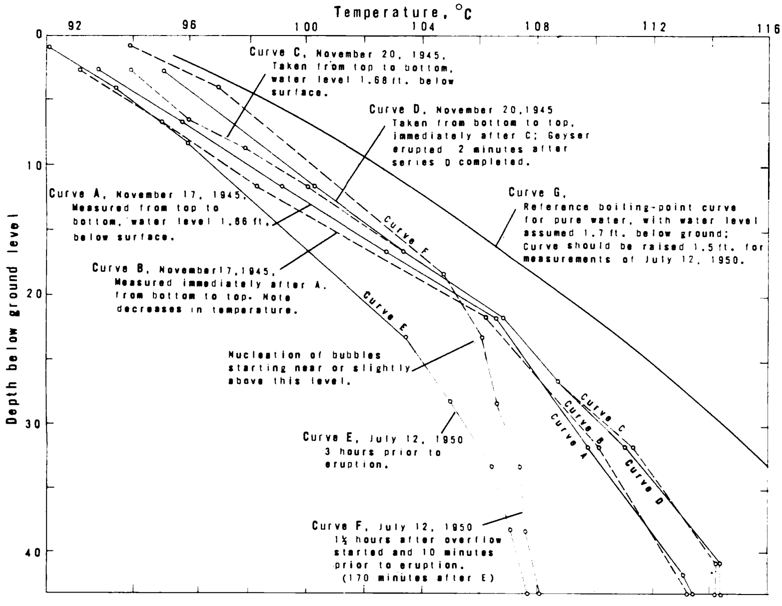


Fig. 8. Temperature-depth curves of November 17 and 20, 1945, and July 12, 1950, showing changes during short intervals of time in the Geyser Well.

evolution of gas decreased with time. During the second series (curve B), measured from bottom to top, notably less gas was evolved than during the first series; large bubbles were completely absent and no vigorous boiling occurred.

A comparison of curves A and B shows a significant decrease in temperatures in the upper part of the column. The *average* of curve B is about $\frac{1}{3}$°C lower than for A, and changes of as much as 1°C were measured near the surface (reproducibility with this thermometer was 0.1 to 0.2°C, so most of the indicated changes are significant). The changes in temperature from curves A to B are evidently related to the formation and escape of vapor bubbles and to convection. The water was initially supersaturated with respect to dissolved gases (principally CO_2, Brannock and others, 1948), and introduction of the thermometer and steel tape provided sites for nucleation of bubbles, even though all temperatures were at least 5°C below the curve for pure water.

Immediately prior to the November 20 measurements shown in figure 8, a few large bubbles were rising to the surface at intervals of 1 to 2 seconds. When the thermometer was first held 1 foot below water level, a few small bubbles formed around the thermometer. At greater depths additional large bubbles were evolved, resulting in vigorous boiling for a short time, after each deeper penetration. At depths of 5 and 25 feet, boiling became so vigorous that the thermometer was temporarily raised several feet. I had found by experience that continued boiling of such high intensity soon led to eruption, but if the thermometer were raised to levels already deprived of much of the excess gas, the series of measurements perhaps could be continued without eruption. This did indeed postpone eruption, so curve C was completed, and series D from bottom to top was started. During series D many large bubbles were again evolved each time the thermometer was lowered. At 10 feet below water level the boiling had become very vigorous, and, in contrast to earlier measurements, the intensity of boiling did not decrease after the thermometer was removed. This signified that an eruption was imminent and that time was not available for temperatures at both the 5- and 1-foot levels. The temperature at 1 foot was then measured as boiling increased. The general water level in the meantime had changed little, except for minor surges of perhaps 0.1 foot. But a few seconds after the thermometer was finally removed, the water level started to rise from 1.65 feet below the surface. After initial slight overflow the geyser immediately surged upward into full eruption.

Little or no significant change in temperature had occurred at depths of 15 to 30 feet between series C and D of November 20. But near the bottom and especially at depths of less than 15 feet series D increased in temperature by as much as 1°C (*all* measured temperatures were still 3°C or more below the boiling-point curve for pure water). The average temperature in the well increased between the two series in spite of the loss of considerable unmeasured heat in vapor resulting from the increased boiling. The change between the two series of November 20 was almost opposite in effect to that of November 17. The observed relationships on the 20th are satisfied only if heat was added from outside the geyser tube. Perhaps the increased boiling and convection in the tube promoted exchange with somewhat hotter water of the reservoir below the bottom of the casing.

On November 15 (5 days before the eruption described above) 20 grams of fluorescein dissolved in water was poured into the Geyser Well

TABLE 2

Chloride and fluorescein contents of samples from the Geyser Well, November 14 to 20, 1945

Date	Eruption Hr Min Sec	Chloride content, ppm	Fluorescein content, ppm	Comments
Nov.				
14	3:10 p.m.	880		Eruption
	3:54 p.m.	?		End of eruption
15	6:00 p.m.			20 grams fluorescein put in well
16	8:50 a.m.	912	4.0	
17	9:13 a.m.	880	0.5	
18	9:00 a.m.	882	0.25	Trace of fluorescein first appears in vents 30 and 31 to northeast
19	10:15 a.m.	880	0.2	
20	9:20 a.m.	876	0.01	0.03 to 0.04 ppm fluorescein estimated in vents 30 and 31
	12:11:00 p.m.			Eruption
	13:30 ±	936	0.02	In bursting stage of eruption; maximum measured rate of discharge 110 gpm at weir
	14:30			End of bursting stage; discharge decreased greatly
	16:30	964	0.03	Measured discharge 45 gpm
	19:00	964	0.04	" 30
	25:00	952	0.06	" 20
	32:00	960	0.04	" 19
	39:00	952	0.06	" 14
	46:00	940	0.05	" 17
	50:00	936	0.08	" 21
	50:40			End of eruption; total measured discharge 985 gal
	55:30			End of violent steam stage
	1:12:30			End of passive steam stage
	18:00	902	0.03	
23	8:50 a.m.		Trace	Trace of fluorescein in vent 30 but none detected in 31
26	8:30 a.m.		< 0.01	No fluorescein detected in any of the vents

to study volumes and movements of water involved in the eruption and recovery phases. As shown in table 2, water samples were collected at daily intervals until November 20 when ten samples were obtained. These samples were analyzed for chloride and fluorescein, determined both by colorimetric and by fluorescent (ultraviolet mercury vapor lamp) methods by comparison with prepared standards. The fluorescein contents shown in table 2 are averages of the two methods, which were in close agreement. The principal conclusions drawn from these data are:

1. Convection in the geyser tube is strongly confirmed; otherwise, all fluorescein would have remained near the surface of the well.

2. Fluorescein appeared visibly in vents 30 and 31 on November 18 (fig. 5) and by midmorning of November 20 the visual intensity of color was considerably greater in these two vents than in the Geyser Well. Intercommunication between natural vents and the well thus was proved conclusively, and a general northward migration of water was indicated as a part of the "reloading" process.

3. If thorough mixing is assumed with no fluorescein adsorbed on mineral surfaces or thermally decomposed, 100,000 gallons of water was involved by November 19 and 2 million gallons by the following day. The data prove a northward migration, but they also prove that mixing was not uniform; the total reservoir seems likely to involve as much as 100,000 gallons of water, but no weight can be given to specific figures. Because of the doubtful validity of the assumptions, similar quantitative measurements should be made in other geysers using tracers of known thermal stability.

4. On November 20 the content of fluorescein in the erupting water increased notably and attained a maximum immediately before the end of the eruption. This indicated that the direction of migration noted above had been reversed, and, more important, it proved that the water supply for sustaining the eruption was derived in large part from storage in the immediate vicinity.

5. Pronounced changes in chloride concentration occurred during eruption, equivalent to an increase of about 9 percent. Elsewhere (White, in press; White, Sandberg, and Brannock, 1953, p. 497) I have shown that this change results from conversion of water to steam, with concentration of involatiles in the residual water. Na, Li, F, and B show similar increases, but SO_4 is not so enriched. Noguchi (1956), Iwasaki (1962, p. 24), and Noguchi and Nix (1963) found that Cl increased during the eruptions of many geysers but that SO_4 frequently decreased, probably because of eruptive discharge of deeper water in which less S had oxidized to SO_4.

Bottom-hole temperature recovery curves.—Some recovery curves of bottom-hole temperatures are shown in figure 9. Five curves are included from the very active period of June 1945 to March 3, 1946. Three of the five, curves A, C, and D, show relatively rapid but differing rates of recovery. Curve A showed the most rapid recovery, constituting the shortest observed interval for a natural eruption listed in table 1.

Curves B and E are of special interest because of reversals in temperature 3 to 5 days after the preceding eruptions. Both reversals coincide with and are clearly related to very heavy storms; precipitation entered into and temporarily cooled the geyser system, thereby delaying temperature recovery and eruption by 3 to 4 days.

Total discharge per eruption.—A weir was installed in the gully about 50 feet south of the Geyser Well (fig. 5). Normally most of the discharge of an eruption drained off to the south, but during the eruption of November 20, previously discussed, a rather strong south wind diverted much discharge to the north. For this reason the total of 985 gallons measured for the November 20 eruption is too low.

The total discharge of water measured under more favorable conditions during the January 27, 1946, eruption was 1350 gallons. Additional water saturated the porous sinter around the well; some drained off to the north, and some escaped as steam. These unmeasured quantities probably account for at least an additional 20 percent or a total of about 1600 gallons. In comparison, the volume of the geyser tube is only 50 to 60 gallons, but the capacity of the whole reservoir, judging from the fluorescein data of table 2, may be in the order of 100,000 gallons. Benseman (1965) proposed the concept of a finite reservoir; when exhausted, the eruption ceases, but this concept probably is not valid for most geysers.

Relations to nearby fissures and vents.—Interconnections between geysers and other springs or geysers have been stressed by Marler (1951), Benseman (1965), and Bloss and Barth (1949). The fissure 16 feet east of the Geyser Well (figs. 5 and 6) is probably the principal channel of

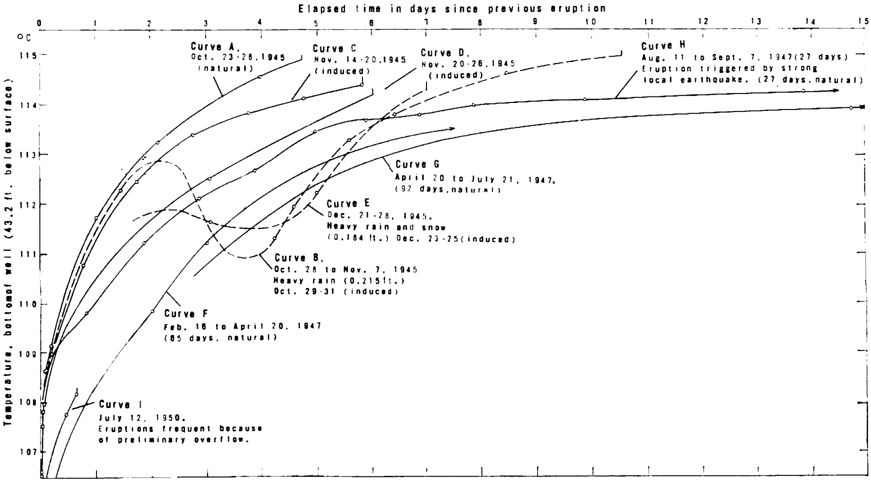


Fig. 9. Curves showing recovery of bottom-hole temperature in the Geyser Well.

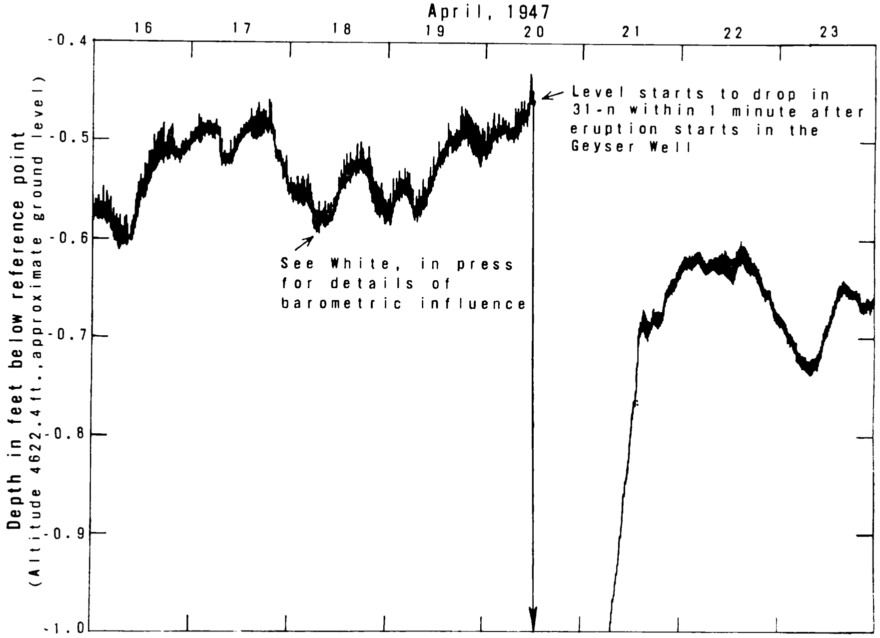


Fig. 10. Water-stage record from vent 31-n showing barometric fluctuations and effects of eruption of the Geyser Well.

upflow of hot water. At the same depth as the bottom of the Geyser Well (43 ft below the surface), the temperature of upflowing water in the fissure is probably somewhat higher than the maximum measured in the Geyser Well (115°C, but known to be affected by convection) and below the boiling point for pure water at this depth (121.4°C) because of dissolved gases. The most probable temperature is estimated at 117° ± 1°C.

The Geyser Well and nearby vents are connected by very permeable channels, as proved by the fact that water levels in vents 31 and 31-n (shown in fig. 5) started to decline within a few seconds after initiation of eruption in the Geyser Well and attained maximum rates of decline within 2 to 4 minutes. Response in vent 30 farther to the northeast was less rapid but was clearly detected 90 seconds after eruption and attained a maximum rate of fall in 10 to 14 minutes.

A water-stage recorder was installed at vent 31-n in October 1946, 7 months after the very active period being discussed here, but responses were probably similar. Records from November and December 1946, uncomplicated by eruptions of the Geyser Well, are discussed by White (in press, figs. 17 and 18). Figure 10 illustrates the response of vent 31-n to an eruption, showing many details not evident from periodic measurements.

The water supply for an eruption evidently is obtained from vents and open cavities on interconnected fissures and from cavities that may have formed near the bottom of the well by ejection of rocks and sand during its early eruptive history. An even more important part of the reservoir is probably the interconnecting pore spaces in the sand, gravel, and sinter deposits that underlie the area.

Termination of eruption.—Much confusion exists in geyser literature concerning termination of eruptions. Intrusion of cold surface water is frequently called upon (Allen and Day, 1935); Bloss and Barth (1949) conclude that cool water may terminate some but not all geyser eruptions, but Golovina and Malov (1960) find no evidence for a cold-water influx.

The fluorescein and chloride data of table 2 prove that normal saline hot-spring water dominates the Geyser Well system throughout its eruption. Chloride commonly increased from about 880 parts per million to about 960 parts per million during an eruption, and this change is produced by volatilization of about 8½ percent of the original water, concentrating the nonvolatile chlorides in the residual water (White, in press).

During the early bursting stage of eruption the supply of water and of energy was at maximum levels; rate of discharge at the well mouth probably exceeded 200 gallons per minute. The deeper plumbing system of the geyser, however, could not continue to sustain a discharge as little as the 20 gallons per minute (plus some unmeasured water and steam) that characterized the second half of an eruption. From this we must conclude that the initial high rate of discharge depends largely on water (and energy) previously stored in the system, as emphasized by Benseman (1965). As the reservoir becomes depleted, pressure gradients increase toward the well. Water levels in communicating fissures and vents of the surrounding area respond, and water stored in the porous wallrocks drains into the fissures or, deeper in the system, is forced into fissures by expanding steam in interconnecting pore spaces. Some cooler water from near the surface sinks down into hotter ground and is heated by energy previously stored in the rocks. The temperature of water flowing into the geyser tube, partly from higher cooler levels and partly from deeper hotter levels, declines during the eruption as water levels and pressures decline. An irregular zone of falling water levels forms around the geyser tube. This phenomenon is similar to the subsurface cone of depressed water levels or pressures that form around a water well when pumping starts.

Pressures throughout the reservoir system of the Geyser Well must have been at or near their maxima immediately prior to eruption when the plumbing system was largely filled with liquid water. Although pressures were not measured directly, temperatures and water levels were measured, and their effects on pressure can be interpreted. As the eruption proceeds the proportion of vapor to liquid in the tube increases (note characteristics of the bursting stage and rates of discharge of table 2), so we conclude that pressures in the reservoir system decline progressively. At any given depth below the ground surface, temperatures formerly near the boiling point for a water-filled column are now much above the temperature of boiling for a vapor-dominated column. A part of the liquid water in pore spaces flashes into steam, and in addition heat stored in the solid phases converts more pore water to steam. Abundant rapidly expanding steam is thus available initially to eject the associated liquid water, but both water and heat are discharged faster than they are being replenished from depth. As an effective cone of depression forms and deepens, the distance that water must be ejected to reach the surface increases, thereby increasing the demand on energy required for ejection.

When the decreasing supply of available energy can no longer meet the increasing demand, the eruption is terminated. At this critical point, the volume of steam that is escaping has become small enough so that it can again rise *through* the available water in the tube, without ejecting water to the surface.

Induced eruptions and their significance.—Many eruptions of the Geyser Well were induced, ranging in subtlety from the very careful introduction of a bare glass thermometer to a depth of only 1 foot (table 1, Feb. 6, 1946) to the forced unloading of many gallons of water from the top of the water column by rapidly lifting a plunger inserted repeatedly below water level. On the former occasion the temperature at the 1-foot level was only 91.4°C or 5.0°C below the boiling point of pure water at this depth (96.4° C). When the thermometer was first introduced, gas bubbles started to nucleate on the thermometer. The vigor of boiling gradually increased and failed to change even after the source of irritation, the thermometer, was withdrawn. Two or 3 minutes after the initial introduction the water level started to rise from 2.44 feet below the surface until the top of the column overflowed, and the well surged immediately into full eruption.

Prior to the thermometer-induced boiling, the water was supersaturated with respect to its dissolved gases and was in a delicately balanced unstable state. Why did boiling continue after the thermometer was removed? I suspect that the gas bubbles evolving around the thermometer set up a mechanical disturbance that induced nucleation, migrating down the gas-supersaturated column as a chain reaction. The same mechanical disturbance, greatly exaggerated, explains the rapid nucleation and growth of gas bubbles when a bottle of carbonated beverage is shaken violently.

Eruptions were also induced in other ways. Some were intentional, but most were indirect results of various measurements made in water strongly supersaturated with gases. Each was effective only when the system was sufficiently "loaded".

Changes in behavior after March 3, 1946.—On March 2, 1946, Steamboat Well 4, 350 feet south of the Geyser Well, was cleaned to the bottom (table 1; and White, in press). The Geyser Well soon erupted, as it frequently did in some complex response to the cleaning of the nearby wells. Multiple eruptions of the Geyser Well then occurred, one soon after another until midafternoon of March 3.

After this eruption series, the recovery in water level in the Geyser Well was much slower than usual. Eleven days after the eruptions, the water level was still 3.1 feet below the surface, and the bottom temperature was only 114.5°C, indicating a much slower thermal recovery than in curves A to E of figure 9, including those affected strongly by heavy precipitation. In addition, bubbles did not evolve in response to introduction of a thermometer, even when temperatures at the 1-foot depth were as high as 94.3°C.

Contemporaneous with this change, vent 31 became much more active, with vigorous boiling and a rise of about 1 foot in water level, relative to the Geyser Well. The next eruption of the Geyser Well was on April 16 after an elapsed time of 44 days instead of the usual 5 to 7. Measurements through the following two years showed that bottom-hole temperatures in the well were as high as 115°C after long intervals of quiescence, but the gas content and degree of instability of the water evidently were much less than prior to March 1946.

Curves F, G, and H of figure 9 demonstrate that bottom-hole temperature relationships did change significantly, requiring much longer times for recovery. Evidently an exchange in function, of the type described by Marler (1951), occurred, and circulation patterns had changed. The lower gas content and lower rate of recovery of temperatures in the well were probably caused by diversion of primary upflow in the controlling fissure into a channel near vent 31, where much excess gas and heat were lost. Some of this gas-deficient water may then have circulated into the Geyser Well.

During the 41 months from March 1946 through July 1949 the interval between eruptions ranged from 8 to 225 days (table 1), but no natural eruption occurred in less than 13 days. The eruption of September 7, 1947, is of interest because it was almost certainly triggered by local earthquakes (White and others, 1964, p. B53-B55). Marler (1964) described similar but much more extensive effects of the Hebgen Lake earthquake of 1959 on the hot springs and geysers of Yellowstone Park.

From March 1946 through July 1949, the depth to water immediately prior to eruption ranged from 0.3 to 2.3 feet. The differences in level were determined largely by changes in discharge of the Steamboat Wells (for details see White, in press, especially fig. 50).

After July 1949, the Geyser Well was inactive whenever water level was low, but when the water level rose within 0.2 foot of the ground surface, eruptions were frequent. At first the reason for this was not clear but then I found that corrosion had penetrated through the iron casing of the well at a depth of 0.15 to 0.18 foot, and slight leakage was taking place when water levels were high enough. Although these changes occurred in a manmade vent under special conditions, they provided a unique opportunity to study the effects of preliminary overflow in the reloading of a geyser system as compared to its behavior when preliminary overflow could not occur.

In June 1950, for example, when the water level was generally about 0.1 foot below ground, eruptions occurred at intervals of about 1 week. On June 29, 1950, I enlarged the hole in the casing slightly by removing a little iron oxide. Seepage through the casing had not previously exceeded 1/20 gallon per minute, but the enlarged hole permitted a discharge of about ¼ gallon per minute. With this increased rate of flow, the average interval between eruptions decreased abruptly to about 1 day, as illustrated for July 9 to 11 from responses in vent 31-n (first 3 eruptions of fig. 11).

The summer of 1950 was also a period of unprecedented high water levels in all nearby vents. As a result, vent 31-n filled and eventually overflowed after each eruption of the Geyser Well. On July 12, in further experimentation on the influence of preliminary overflow, the outlet of vent 31-n was dammed with sinter so that discharge ceased. This resulted in a rapid rise of about 0.2 foot as shown on figure 11. This in turn forced a corresponding rise in water level in all nearby related vents, including the Geyser Well. Preliminary seeping discharge through the corroded casing started 1½ hours later at 11:15 a.m., it increased in rate until 12:59 p.m. when boiling was first noted, and eruption occurred at 1:05 p.m. During the following 4 days until the dam at 31-n was removed, eruptions occurred at average intervals of 8 hours as shown in figure 11. This was three times the frequency of the previous two weeks.

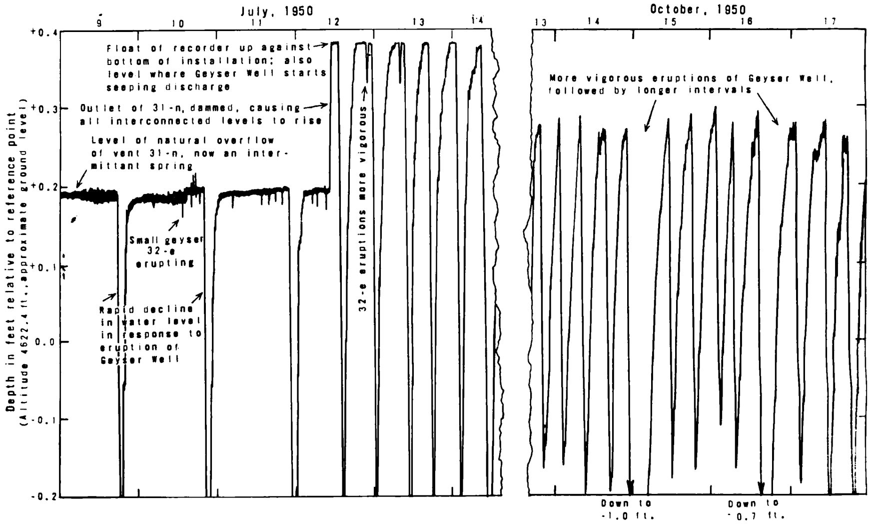


Fig. 11. Water-stage records from vent 31-n showing effects of preliminary overflow on the activity of the Geyser Well; also shows effects of increasing the rate of preliminary overflow and two different intensities of eruption.

The important changes in temperature that occurred during the 3 hours preceding the first eruption of July 12 are shown as curves E and F of figure 8 and curve I of figure 9. These curves differ greatly from those unassociated with preliminary overflow, especially in temperature relationships at depths greater than 20 feet. With marked preliminary overflow, the cooler water on the top of the column was discharged immediately from the system instead of recirculating by convection. Hotter water flowed in near the bottom and then upward, with inflow increasing less than ½°C in temperature during the 3-hour interval between the two curves of figure 8. In contrast, the upper part of the system was increasing rapidly in temperature.

No vapor phase was involved in the system until after considerable preliminary discharge had occurred, when small gas bubbles were first noted about 20 minutes prior to eruption. Curves E and F and field observations provide clear evidence that a vapor phase *first* started to form at the surface, and as temperatures increased *initial* nucleation of bubbles progressed to greater and greater depths in the upflowing column of water. Ten minutes prior to eruption, judging from the shape of curve F of figure 8, nucleation was probably starting about 21 feet below the surface.

Triggering of the July 12 eruption could not have occurred near the bottom of the well, where maximum temperatures were about 12°C below the boiling point of pure water and 7°C below likely temperatures of initial nucleation of vapor bubbles.

By October 1950 the well casing was almost completely corroded at a depth of about 0.15 foot, and still higher rates of preliminary discharge were occurring through holes in the casing. As a result, three or four eruptions took place per day (fig. 11), spurting with much less than former vigor to heights of only 20 to 40 feet. The duration of each eruption was generally only 20 to 30 minutes, and the water in vent 31-n fell only about 0.5 foot in response to each eruption. Occasional eruptions were more vigorous, with about double the decline in water level and the interval of time required to recover for the next eruption.

Systematic observations at Steamboat Springs were terminated in September 1952. In 1955 Steamboat Well 6, not shown in figure 5, was drilled 250 feet south of the Geyser Well, or 100 feet nearer than wells 2 to 4 which had been the principal earlier producers. This seems to have robbed the normal water supply of the geyser to such an extent that water levels declined and eruptive activity ceased.

The triggering mechanism.—Many students of geyser behavior, after failing to find temperatures above boiling except perhaps near the surface, have concluded with Allen and Day (1935, p. 198) that the triggering mechanism *must* occur beyond the accessible limits. The described eruption of July 12 clearly disproves such an argument. In other geysers and at former times in the Geyser Well the triggering may have occurred beyond the accessible limits.

Perhaps as long as boiling was restricted to the geyser tube, the gas bubbles could still rise *through* the water without causing much change in water level. But if the level of initial boiling progressed down the tube and eventually extended beyond the limits of the drilled well, perhaps to an enlargement such as near A of figure 6, previously filled with liquid water, expansion of gas collecting in the upper part of A might then force hot water into the well. The water level at the surface would then rise to level of overflow, initiating an eruption.

An alternative explanation, in part similar to the mechanism proposed by Vymorokov (1960) and Golovina and Malov (1960, p. 616), assumes that the triggering action occurs near the surface, as shown in a general model (fig. 12) useful in understanding the Geyser Well and natural geysers with irregular tubes. Individual bubbles start to form at deeper and deeper levels, as in A to B, as temperatures recover from the previous eruption. New bubbles help to increase the rate of convective upflow within the zone of streaming bubbles; convective downflow also occurs, generally near the sides of the well or vent. Bubbles in the upflowing zone enlarge greatly as they rise (B of fig. 12), in part because of expansion with decreasing pressure, but also by accretion as more gases and steam vaporize from the supersaturated water in the upflowing zone.

Some critical size or total volume of bubbles is eventually attained, differing with physical relationships in each geyser. Frictional resistance and viscosity inhibit downflow of water between the expanding bubble-strewn upflowing zone and the confining walls (C of fig. 12). Water swept upward with bubbles is no longer balanced by downflow; the column then expands until water overflows at the surface (D of fig. 12).

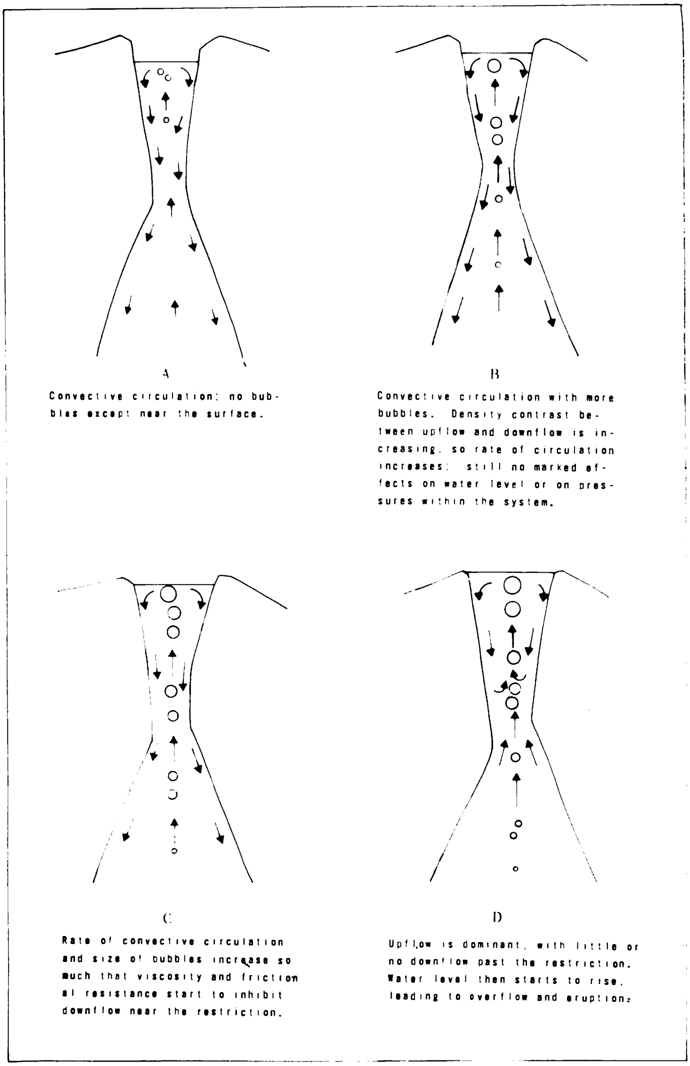
In a geyser with preliminary discharge prior to eruption, as for the Geyser Well eruptions of figure 11, a similar increase in size and number of bubbles is effective in increasing the rate of discharge until water flows out faster than it flows into the geyser tube from below.

This explanation for initiation of overflow and eruption of the Geyser Well prior to March, 1946, is supported by the fact that incipient eruptions could be stopped even after the water level had started to rise by rapidly pouring a quart or two of cold water in the well. The cold water greatly affected near-surface temperatures and relative pressures, but at the bottom of the well the only immediate effect was to increase the pressure by less than 1 percent.

GENERAL BEHAVIOR OF GEYSERS

Geysers characterize the hot-spring systems that are particularly unstable near the surface. Any one of many natural or human influences can initiate a chain reaction that leads to an eruption. Probably no two geysers are identical in behavior, and, as we have seen at Steamboat Springs, individual geysers can change greatly in their behavior with time. For this reason no universal cycle can be proposed.

I agree with Iwasaki (1962, p. 57) that the eruption mechanism is not explainable by a single simple theory. Nevertheless, I summarize here



A

Convective circulation; no bubbles except near the surface.

B

Convective circulation with more bubbles. Density contrast between upflow and downflow is increasing, so rate of circulation increases; still no marked effects on water level or on pressures within the system.

C

Rate of convective circulation and size of bubbles increase so much that viscosity and frictional resistance start to inhibit downflow near the restriction.

D

Upflow is dominant, with little or no downflow past the restriction. Water level then starts to rise, leading to overflow and eruption.

Fig. 12. Effects of increasing the number and size of bubbles in upflowing part of a convection cell as temperatures increase, preliminary to overflow and eruption. Effects are most pronounced near restrictions but can also occur in vertical smooth-walled tubes of constant diameter.

a cycle that characterizes many geysers, and I then consider additional facets that may apply to geysers differing from the simple pattern.

1.[1] The immediate plumbing system of a geyser is near the surface (probably within 1 to perhaps 200 feet) and is in the uppermost part of a huge convection system. Water that is dominantly of surface

---

[1] Elements that may be essential for all natural geyser systems.

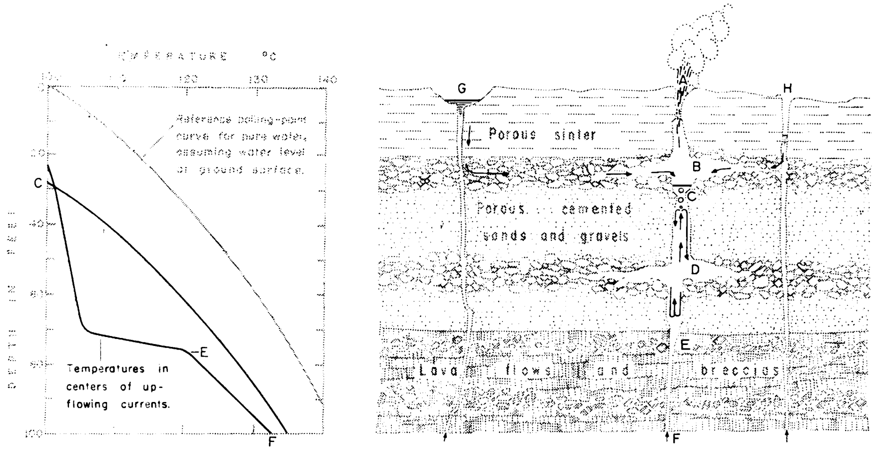


Fig. 13. Immediately after end of eruption and associated steam phase; water levels are at or near their minima, but temperatures are locally high.

origin (probably 95 percent or more) circulates to depths in the order of 5000 to 10,000 feet. Near the base of circulation, as in figure 1, the water is heated by conduction and probably in small part by volcanic fluids to 150°C or higher; lower temperature spring systems do not sustain natural geysers. The hot upflowing water (CDE of fig. 1) contains abundant energy for eruption. Near-surface heating by volcanic gases and chambers for storage of steam are not necessary and are not supported by the evidence.

2.[1] Such a system is fundamentally unstable because water of low density and high energy content is overlain by heavier water of lower energy content. The relative density of liquid H_2O at 100°C, for example, is 0.958 and at 150°C, 0.917.

3. The excess heat of many high-temperature systems is lost near the surface by several means, thereby explaining the absence or scarcity of geysers where they might otherwise be abundant. Excess heat can be lost by thermal conduction if rate of upflow is low; near-surface convection and boiling can occur.

4. Now consider an eruption cycle that starts, for convenience, at the beginning of the recovery phase immediately after an eruption and its associated steam phase have ceased. The water level in the geyser tube is typically at some low level such as near point C in figure 13. During the early part of the recovery phase, unequal levels such as at C and in vents G and H may tend to equalize as water from the surrounding ground drains into the geyser tube. Some steam may continue to escape from the tube, resulting from vaporization of water by heat still stored in the rocks above the level C. Temperatures near C are likely to be higher than equilibrium for pure water.

---

[1] Elements that may be essential for all natural geyser systems.

5.[1] As hot water is being supplied from below and, at first, from interconnected marginal areas, water levels rise, and open cavities and empty pore spaces of the adjacent rocks are generally filled with liquid water, as at B and D of figure 14. Some steam and other gases may be trapped in local pockets. The steam readily condenses as water levels and pressures rise; other gases dissolve more slowly or persist as vapor pockets if pressure does not increase sufficiently for complete resolution to occur.

6. At the end of the water-recovery stage (here considered separately from the overlapping temperature-recovery stage, with both together constituting the total recovery phase), water may start to discharge from one or more interconnected vents, as at G, H, and I (exceptions considered later). Even though the water rising from great depths may be 150°C or more, the first water to be discharged may be cooler than the boiling temperature of pure water. The rocks of a typical plumbing system have a much larger capacity for storing heat than does the water alone, if the effective porosity of the system is less than about 30 percent (White, 1965). The rocks near the chamber of B of figure 13, for example, had been cooled down to 100°C or less by loss of heat in forming steam when pressures were minimal during the preceding eruption and steam phase. In figure 14 the chamber near B has already been heated a few degrees by transfer of heat from the hotter rising water. Convection in the geyser tube may now be vigorous as in the two cells separated at point C.

7. The discharge of cooler water from the system may not be absolutely essential; some geysers may recover in temperature by means of convection and conduction, but preliminary discharge of the cooler water greatly hastens recovery for the next eruption.

8. Deep underground where pressures are sufficiently high, the gases of the natural geyser systems are dissolved completely in the rising hot

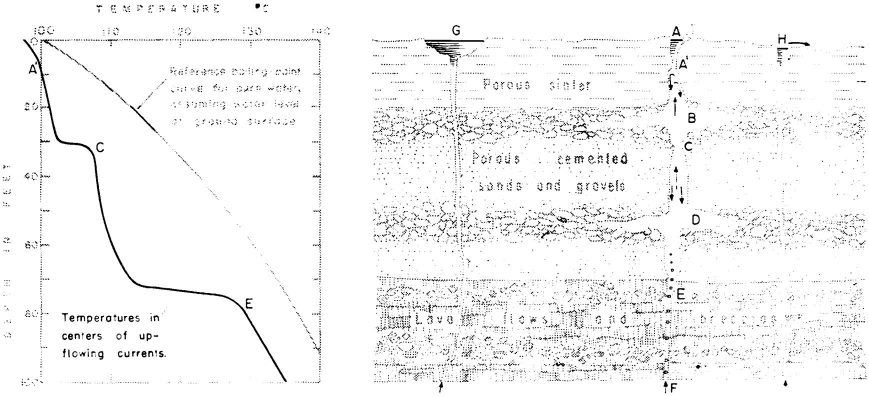


Fig. 14. End of water-recovery stage when one or more interconnected vents start to discharge again.

---

[1] Elements that may be essential for all natural geyser systems.

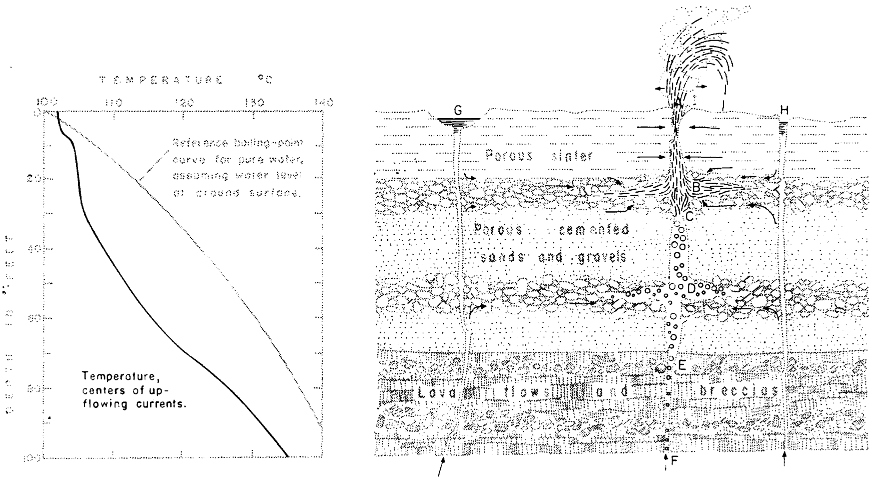


Fig. 15. Early bursting part of eruption phase; supplies of water and heat are abundant. A well-defined interface between liquid and vapor no longer exists, but a poorly defined zone in which vapor is the continuous phase migrates progressively downward during eruption and is here near point C.

water. With upward migration and decrease in pressure, dissolved gases start to exsolve. Temperatures are characteristically slightly below the reference curve for pure water. Thus in figure 14, vapor bubbles greatly enriched in CO_2 and other gases are rising from below point F. During the early parts of the temperature-recovery stage, if gas content is low enough, the bubbles may even redissolve at higher levels where temperatures are enough below the reference curve for pure water. Bubbles that are almost entirely steam without much CO_2 or other gases can rise into cooler water and condense suddenly, with an action like a water hammer. This phenomenon probably accounts for the earth tremors noted by Rinehart (1965) and for the oscillations in water levels prior to eruption noted by Allen and Day (1953) and by Golovina and Malov (1960). The minor fluctuations of figure 10 may have a similar origin. With higher gas contents, the bubbles may persist, decreasing in size in spite of decreasing pressure upward as water vapor condenses from the bubbles.

9.[1] As temperatures increase further because of the heat and water rising from below, reloading of the system for the next eruption is eventually completed. Actual initiation of an eruption can be triggered in many ways and at different sites from time to time in the same system. The most critical points may be similar to those near A, C, or possibly E of figure 14. The number and size of vapor bubbles has been increasing in the upflowing parts of individual convection cells. A few bubbles can rise in the upflowing current without disturbing the adjacent downflowing currents essential in maintaining a balanced convection cell.

[1] Elements that may be essential for all natural geyser systems.

10.[1] But as the number of bubbles increases greatly with increasing temperatures in the system, the rising column expands until frictional resistance becomes too great for convective downflow to balance upflow, as in C of figure 12. Water is then swept along through the bottlenecks with the rising gas; the column must then expand upward, until water starts to discharge over the lip of the geyser tube, either in quiet overflow or in more vigorous preliminary spurts. Cooler water of higher density is unloaded from the top of the column, thereby decreasing pressures below. The rates of boiling and upflow increase further as pressures decline throughout the system and as the chain reaction extends out to parts of the reservoir previously filled entirely with water. The large reservoirs, as near B and D of figure 15, then start to flash as the geyser surges into full eruption.

11.[1] During eruption there is no longer a single clear interface between liquid and vapor that can be called a water level. At depth (as below C in fig. 15) liquid water is the continuous phase, containing dispersed bubbles; in the eruption column, vapor is the continuous phase with dispersed droplets of liquid water. Because of the high stored energy of the system, and because of the large open tube that characterizes most geysers, water is initially erupted at much higher rates than can be replenished from below through narrow restricted channels. Pressures, temperatures, and water levels normally start to fall as eruption proceeds; heat stored in the *rocks* of the system now becomes available to form more steam from water, as in the pore spaces around chamber B in figure 15. This supply of energy stored in rocks can be negligible in feeble geysers, or it can account for more than half the steam of some geysers. Directions of flow in related vents may be reversed, as in G and H. Some interconnected springs may even respond by erupting as geysers if the stored heat of the adjacent rocks forms enough steam to trigger a secondary eruption (not shown in figure 15).

A cone of depressed water levels now develops around the geyser tube, extremely irregular in detail because of the inhomogeneity of the rocks and open spaces of the geyser system, but nevertheless comparable to the cone of depression that forms around a water well as pumping starts. Near the end of the eruption phase, because of the enlarging cone of depression, water is ejected from increasing depth, as near D of figure 16. On the other hand, temperatures and the energy available for eruption decrease as pressures decline and as the stored heat of adjacent rocks is expended in vaporizing water to steam.

12.[1] Finally, a point in time is reached where the steam can no longer eject the available water. Some final spurts may reach the surface as water periodically collects in sufficient quantity and is ejected in "slugs", but in general the rate of steam production has decreased enough so that vapor bubbles can again rise through the available water (terminations of different types will be described).

---

[1] Elements that may be essential for all natural geyser systems.

13. The steam phase that follows the eruption of many geysers commonly consists of two stages (Fix, 1939, p. 101). During the first or violent steam stage, steam and water are turbulently intermixed without a principal interface between the two. The relationships between water and steam that existed during eruption are continued except that the upward spurts no longer have sufficient energy for liquid water to be ejected continuously at the surface. Temperatures in the wall rocks continue to be somewhat above those of boiling of water at prevailing pressures, so steam continues to form.

14. The steam phase changes from the violent to the passive stage when an essentially continuous water interface first forms, as at C in figure 13. Prior to this time vapor was the continuous phase throughout the upper and larger open spaces, but liquid water has now become the continuous phase below the newly formed water level. This level may surge vigorously up and down and is disrupted by individual bubbles, but enough liquid water has accumulated so that water pressure has gained the upper hand over the steam that is still forming. Temperatures above the boiling point of pure water can frequently be measured near point C of figure 13 because of upward convection of superheated water, now exhausted of most of its original dissolved gases. As the water level rises above C of figure 13 because of drainback, seepage, and replenishment from other parts of the system, the steam phase is eventually terminated, and the system has completed a cycle of eruption.

Other important aspects helping to account for the individuality of numerous geysers that do not fit the simple pattern outlined above include:

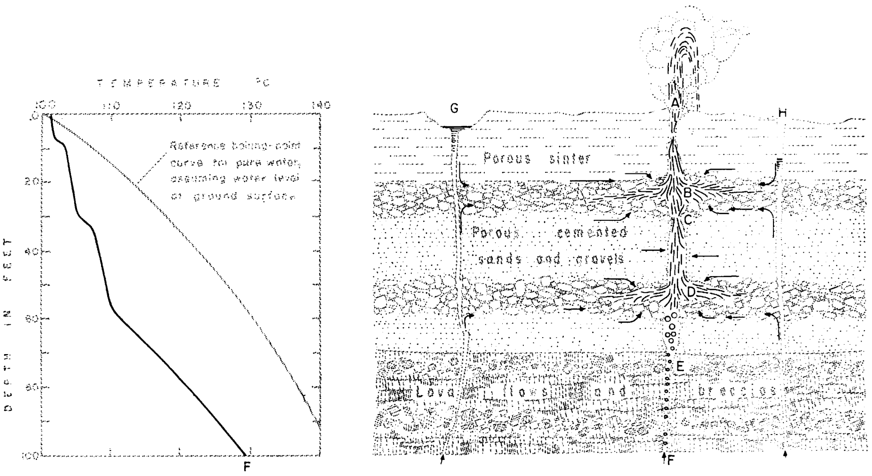


Fig. 16. Near the end of the eruption phase. Temperatures, pressures, and water supply are decreasing, and water must be erupted from increasingly greater average depth.

15. Some geysers seem to be isolated units, such as Old Faithful and Riverside (Marler, 1951), with no proven interconnections with other discharging springs. But even these geysers probably discharge some water from their systems below the surface through interconnected channels and pores, as we have seen for the Geyser Well from the fluorescein experiments. During the temperature-recovery stage, water cooled by loss of heat to colder rocks is gradually flushed out by subsurface discharge, either directly into an adjacent stream or to distant springs with undetected and greatly dampened influences. Subsurface discharge is particularly likely to occur if the geyser is on a terrace or cone considerably above the surrounding ground. Deposition of silica from the hot water tends to fill channels of subsurface leakage, but this sealing process is seldom complete.

16. During both stages of the steam phase hydrostatic pressure of the water is nearly balanced by the vapor pressure of steam and other gases; hydrostatic pressure of the water commonly gains the upper hand after a single eruption. But the balance is close enough so that many geysers erupt again in two or more stages. Each eruption is then a substage of the full multiple eruption. Some geysers show both types of behavior at different times, as demonstrated by the Geyser Well.

17. In some geysers the eruption terminates gradually as the water supply becomes exhausted. The ratio of water to steam decreases until the steam is dry with no suspended water droplets. The heat reservoirs of such geysers are large, relative to the liquid reservoirs, with stored heat in the rocks accounting for the excess steam.

18. In other geysers the heat available for eruption is small relative to water supply. The large pools and vents of some geysers may lose so much heat by convection and evaporation that eruption is greatly inhibited. In some geysers, such as Daisy in Yellowstone Park (George Marler, written communication, 1966), much of the erupted water flows back into the geyser vent. Effectively, this increases the supply of water relative to available energy, terminating the eruption earlier than if return flow were absent. Such eruptions are impeded by too much water of inadequate energy content; a steam phase is either minor or absent.

19. Eruption can be confined almost to a single central tube with narrow feeding channels (Vixen Geyser, Norris Basin, Yellowstone Park), or with a more complex plumbing system it can extend downward or horizontally to tap other parts of the reservoir.

20. When a new geyser first becomes active, the increasing vigor of eruption may *create* large voids not initially present, by forcefully ripping off and ejecting fragments of wall rocks. The vertical or subvertical central tube that characterizes most geysers is probably formed in this way during the early history of a particular geyser.

21. Some vigorous geysers commit suicide by enlarging their central tubes so much that convection becomes dominant, permitting excess

energy to be dissipated constantly by boiling; the local system no longer becomes periodically unstable enough to trigger a chain reaction.

22. The behavior of a geyser can differ from one eruption to another depending on the different parts of the plumbing system that are tapped. Eruption from a shallow part may decrease pressures enough in one or more adjacent or deeper parts to involve them in some eruptions but not in others, if not yet heated sufficiently to become unstable.

23. The site of greatest sensitivity and perhaps even the triggering mechanism may differ with time in a single geyser.

24. Nearly all natural geysers are associated with silica-depositing hydrothermal systems. The precipitated mineral is initially amorphous opal, which may in time change to β-cristobalite (by X-ray). Natural geyser waters have at least 250 parts per million of dissolved SiO_2, with 300 to 400 parts per million as more characteristic contents. In contrast, many wells drilled in areas of much lower silica content erupt intermittently as geysers. These facts suggest that deposition of silica may be essential to stabilize the plumbing systems of most natural geysers. Fournier and Rowe (1966) have shown that 250 parts per million of SiO_2 is not sufficient to precipitate amorphous silica at temperatures above 70°C; about 390 parts per million is necessary at 100°C and 475 parts per million at 120°C. Amorphous silica, therefore, is not normally deposited deep in a plumbing system. During an eruption or in the following steam phase, evaporation frequently concentrates dissolved matter by 5 to more than 50 percent in the residual water (depending on relative contribution of stored heat from the wall rocks). Opal may then be precipitated, perhaps stabilizing the plumbing system and decreasing the frequency of suicide.

25. In many geysers eruption is initiated only after the water level has risen to the ground surface and preliminary noneruptive discharge has occurred. This permits the flushing-out of cooler water from the upper part of the system and replacement by hotter water from depth. In geysers that require preliminary discharge, maximum hydrostatic pressures occur when overflow starts; as discharge continues, temperatures increase and water densities and pressures decline, thereby helping to increase discharge until eruption is initiated.

26. Geysers with large upward-flaring pools lose so much heat by convection and surface evaporation that they generally require preliminary overflow to reload their systems. For this reason, they commonly occur in relatively low-lying areas where the surrounding water table is at or near the ground surface.

In contrast, geysers with narrow vents have minimal convective and evaporative heat losses; where preliminary discharge is not essential, long-continued activity can build sinter cones far above surrounding water levels.

27. Detailed behavior can be affected by barometric pressure, earth-tides, earthquakes, wind velocity, air temperature, and many man-

related physical and chemical effects. Any factor that increases the rate of preliminary discharge of water, such as a decreasing barometric pressure on a pressure-sensitive geyser, or that increases the rate of formation of vapor can induce or hasten an eruption.

28. Change is normal. Subsurface chemical precipitation, earthquakes, and perhaps other causes not yet well understood can cause a shift in activity from one vent to another.

REFERENCES

Allen, E. T., and Day, A. L., 1935, The hot springs of the Yellowstone National Park: Carnegie Inst. Washington Pub. 466, 525 p.

Banwell, C. J., 1957, Borehole measurements, *in* Physics of the New Zealand thermal area: New Zealand Dept. Sci. Indus. Research Bull. 123, p. 39-71.

———— 1963, Thermal energy from the earth's crust; introduction and pt. 1: New Zealand Jour. Geology Geophysics, v. 6, p. 52-69.

Barth, T. F. W., 1950, Volcanic geology, hot springs and geysers of Iceland: Carnegie Inst. Washington Pub. 587, 174 p.

Benseman, R. F., 1965, The components of a geyser: New Zealand Jour. Sci., v. 8, no. 1, p. 24-44.

Bloss, F. D., and Barth, T. F. W., 1949, Observations on some Yellowstone geysers: Geol. Soc. America Bull., v. 60, p. 861-886.

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, p. 211-226.

Byers, F. M., Jr., and Brannock, W. W., 1949, Volcanic activity on Umnak and Great Sitkin Islands, 1946-1948: Am. Geophys. Union Trans., v. 30, p. 719-734.

Craig, Harmon, 1965, The isotopic geochemistry of water and carbon in geothermal areas, *in* Nuclear geology on geothermal areas: Consiglio Naz. delle Ricerche, Spoleto, Italy, 1963, p. 17-53.

DeQuille, Dan, 1876, History of the Big Bonanza: Hartford, Conn.; repub., Wright, W., 1947, The Big Bonanza: New York, A. Knopf, 439 p.

Elder, J. W., 1965, Physical processes in geothermal areas, *in* Lee, W. H. K., ed., Terrestrial heat flow: Am. Geophys. Union Mon. 8, p. 211-239.

Fix, P. F., 1939, Nomenclature of geyser eruptions: Jour. Geology, v. 47, p. 99-104.

Fournier, R. O., and Rowe, J. J., 1966, Estimation of underground temperatures from the silica content of water from hot springs and wet-steam wells: Am. Jour. Sci., v. 264, p. 685-697.

Golovina, I. F., and Malov, N. N., 1960, On the theory of geysers: Akad. Nauk SSSR Izv., ser. geofiz., no. 7, p. 616-620 (English trans.).

Henderson, J., 1938, Te Aroha thermal water: New Zealand Jour. Sci. and Technology, v. 19, no. 12, p. 721-731.

Iwasaki, Iwaji, 1962, Geochemical investigation of geysers in Japan: Tokyo Inst. Technology Bull. 46, 60 p.

Mahon, W. A. J., 1966, Silica in hot water discharged from drill holes at Wairakei: New Zealand Jour. Sci., v. 9, p. 135-144.

Mal'tsev, L. M., and Bubol'ts, A. L., 1961, Well-geyser on the Boya-Dag: Akad. Nauk Turkmen SSR Izv., Ser. Fiz-Tekh., Khim-i Geol. Nauk, no. 1, p. 68-73.

Marboe, E. C., 1949, Gas evolution from supersaturated liquids: Chem. and Eng. News, v. 27, no. 31, p. 2198-2202.

Marboe, E. C., and Weyl, W. A., 1948, Release of gas supersaturation from liquids by solid surfaces: Soc. Glass Technology Jour. Trans., v., 32, p. 281-290.

Marler, G. D., 1951, Exchange of function as a cause of geyser irregularity: Am. Jour. Sci., v. 249, no. 5, p. 329-342.

———— 1964, Effects of the Hebgen Lake earthquake of August 17, 1959, on the hot springs of the Firehole Geyser Basins, Yellowstone National Park, *in* The Hebgen Lake, Montana, earthquake of August 17, 1959: U.S. Geol. Survey Prof. Paper 435, p. 185-197.

Noguchi, Kimio, 1956, Geochemical investigation of geysers in Japan: Pacific Sci. Cong., 8th, Philippines, Proc., v. 2, p. 228-233.

Noguchi, Kimio, and Nix, Joe, 1963, Geochemical studies of some geysers in Yellowstone National Park: Japan Acad., Tokyo, Proc., v. 39, no. 6, p. 370-375.

Nolan, T. B., and Anderson, G. H., 1934, The geyser area near Beowawe, Eureka County, Nevada: Am. Jour. Sci., 5th ser., v. 27, p. 215-229.

Nomura, Yukichi, 1954, Studies on geysers at Onikobe [Japan]: Tohoku Univ. Tech. Rept., v. 19, p. 45-62.

Rinehart, J. S., 1965, Earth tremors generated by Old Faithful Geyser: Science, v. 150, no. 3695, p. 494-496.

Sappa, Mercurino, 1955, Fenomeni geyseri formi dell' isola d'Ischia [Geyseriform phenomena of the island of Ischia]: Bull. volcanol., ser. 2, v. 16, p. 225-250.

Takenouchi, Sukune, and Kennedy, G. C., 1964, The binary system H_2O-CO_2 at high temperatures and pressures: Am. Jour. Sci., v. 262, p. 1055-1074.

——— 1965, The solubility of carbon dioxide in NaCl solutions at high temperatures and pressures: Am. Jour. Sci., v. 263, p. 445-454.

Thorkelsson, Thorkell, 1940, On thermal activity in Iceland and geyser action: Vísindalfélag Islendinga [Soc. Sci. Islandinga], no. 25, 139 p.

Vymorokov, B. M., 1960, How geysers work: Priroda, no. 11, p. 97-99.

Waring, G. A., 1915, Springs of California: U.S. Geol. Survey Water-Supply Paper 338, 410 p.

White, D. E., 1955, Violent mud-volcano eruption of Lake City hot springs, northeastern California: Geol. Soc. America Bull., v. 66, p. 1109-1130.

——— 1957, Thermal waters of volcanic origin: Geol. Soc. America Bull., v. 68, p. 1637-1658.

——— 1965, Geothermal energy: U.S. Geol. Survey Circ. 519, 17 p.

——— in press, Hydrology, activity, and heat flow of the Steamboat Springs thermal system, Washoe County, Nevada: U.S. Geol. Survey Prof. Paper 458-C.

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

White, D. E., Craig, Harmon, and Begemann, F., in press, Isotope geology of the Steamboat Springs area, Nevada, *in* Nuclear geology on geothermal areas: Consiglio Naz. delle Richerche, Spoleto, Italy, 1963, in press.

White, D. E., and Roberson, C. E., 1962, Sulphur Bank, California, a major hot-spring quicksilver deposit, *in* Engel, A. E. J., James, H. L., and Leonard, B. F., eds., Petrologic studies: Geol. Soc. America, Buddington V., p. 397-428.

White, D. E., Sandberg, C. H., and Brannock, W. W., 1953, Geochemical and geophysical approaches to the problems of utilization of hot-spring water and heat: Pacific Sci. Cong., 7th, New Zealand 1949, Proc., v. 2, p. 490-499.

White, D. E., Thompson, G. A., and Sandberg, C. H., 1964, Rocks, structure, and geologic history of Steamboat Springs thermal area, Washoe County, Nevada: U.S. Geol. Survey Prof. Paper 458-B, p. B1-B63.

White, D. E., and Waring, G. A., 1963, Volcanic emanations, *in* Data of geochemistry: U.S. Geol. Survey Prof. Paper 440-K, p. K1-K29.