Hydrology, Activity, and Heat Flow of the Steamboat Springs Thermal System, Washoe County Nevada

GEOLOGICAL SURVEY PROFESSIONAL PAPER 458-C

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By DONALD E. WHITE

GEOLOGY AND GEOCHEMISTRY OF THE STEAMBOAT SPRINGS AREA, NEVADA

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The present physical activity and detailed behavior of a notable hot-spring system, where water of surface origin circulates deeply in rocks of low bulk permeability

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HYDROLOGY, ACTIVITY, AND HEAT FLOW OF THE STEAMBOAT SPRINGS THERMAL SYSTEM, WASHOE COUNTY, NEVADA

By DONALD E. WHITE

ABSTRACT

This report is the third of a series of detailed studies of Steamboat Springs. Isotope ratios of D : H and $O^{18} : O^{16}$ of the spring waters indicate a system consisting almost entirely of meteoric water. Recharge is not directly from Steamboat Creek, the master stream of the area, but is largely from streams from the Carson Range, west of the springs; some recharge, however, is from streams from the Virginia Range, east of the springs.

The driving force for deep circulation is in part related to the differences in altitude between recharge areas and the springs, but the differences in density between cold downflowing water and hot upflowing water is of equal or greater importance. The difference in average density is likely to be 9 percent or more. If water circulates to depths on the order of 10,000 feet, as hypothesized, the driving force related to density differences may be equivalent to 9 percent of 10,000 feet, or 900 feet of water. Circulation below the shallow sedimentary and volcanic cover is in fractured and faulted Mesozoic granitic and metamorphic rocks with small but significant mass permeabilities. Major continuous channels of high permeability are probably lacking. In such rocks, deep circulation is inhibited by low permeabilities that presumably decrease with increasing depth, but this is evidently offset, at least to depths of a few thousand feet, by increase in driving force that is a direct function of differences in water temperature and density with depth; other important factors that favor deep circulation is the large decrease in viscosity and increase in solubility of silica in water, with increase in temperature.

Flowing springs are localized near Steamboat Creek on and near the Low and Main Terraces. In higher ground in the western part of the thermal area, springs were very active through much of the Pleistocene, but water levels are now generally 40 to more than 100 feet below the surface. Convection subsystems occur in the High Terrace, Sinter Hill, and elsewhere. Thermal water flows up the controlling structures of these subsystems and then outward below the surface into the more permeable wallrocks, eventually escaping unseen directly into Steamboat Creek. Total discharge from the visible springs is only about 65 gpm, or 6 percent of the total computed upflow of the Steamboat Springs thermal system.

The discharge of springs and water levels of the system are influenced by precipitation, changes in barometric pressure, earth tides, earthquakes, discharge from geothermal wells, and other short-term random changes. Other factors not considered in detail control long-term changes over hundreds and thousands of years; these include climatic changes, erosion and sedimentation of Steamboat Creek, changes in the magnetic hearth that supplies the excess heat of the system, structural events that create new channels, and vein filling that decreases permeability in old channels.

The Low and Main Terraces behave in some ways as separate subsystems. The springs and nonflowing vents nearest each terrace crest are highest in temperature and the most responsive to changes in barometric pressure; water levels, especially in nonflowing vents near each terrace crest, behave as inverted water barometers. Some vents respond with barometric efficiencies of 60–85 percent of perfect water barometers and even exceed 100 percent for short intervals. Barometric response seems to be determined largely within the upper 350 feet of each subsystem, where temperatures are close to the boiling-point curve with depth and a vapor phase is forming in the rising water because of the upward decrease in pressure. Continuous instrumental records of water-level fluctuations, corrected for barometric changes, show earth-tidal responses in the Main Terrace, but earth tides are not detected in the Low Terrace. The differences in response are explained by differences in altitude; the channels of low-altitude springs must be restricted, and water is driven through these channels by an excess of pressure equivalent to at least 45 feet of water. The highest water levels of the Main Terrace are at altitudes of about 4,668 feet and are the most responsive to minor changes. For comparison, the highest and most responsive water levels of the Low Terrace are at altitudes near 4,623 feet.

The springs respond to yearly and seasonal changes in precipitation. Individual storms with precipitation of one-half inch or less generally produce no response, but precipitation of more than one-half inch per storm saturates the porous sinter above the water table and seeps downward, diluting and cooling the saline thermal water body; heavy precipitation in the small basins west of the active springs also influences the spring system to a minor extent.

Local earthquakes affect different parts of the system in different ways, but detailed effects are not predictable.

The first geothermal wells were drilled about 1920 to provide a dependable supply of hot water for the resorts of the area. In recent years their average discharge has been about 300 gpm. Much of this discharge has evidently been diverted from the natural springs. Temperatures in wells in the central part of the thermal area increase very rapidly within the upper 300 feet and closely approach the theoretical boiling-point curve for the prevailing water pressures. Maximum temperatures in the Main Terrace approach 172°C near 350 feet in depth and then

level off or decrease slightly with greater depth. This phenomenon, called a "leveling-off" or base temperature, has been observed in many geothermal explorations throughout the world. It is related to the rates of flow of heat and water through each individual system. In a single-celled convection system, liquid water near the base of circulation is heated and then rises in the core of the system, losing little heat through most of its rise because of the high insulating properties (low thermal conductivities) of ordinary rocks. In systems with temperatures above the surface boiling point, the water rises until hydrostatic pressure is low enough for a vapor phase to form. The deep Steamboat Springs water at about 172°C rises within about 350 feet of the surface, where decreasing pressure first permits vapor bubbles enriched in CO_2 and H_2S to form. With further rise in position and decrease in pressure, more water is converted to steam. Water near the base of this boiling zone is high in temperature and enthalpy but lower in density ($\sim$ 0.90) than the cooler water near the surface (density near 0.99 at 95°C). Such a situation is very unstable and is the fundamental cause of eruption of geysers and geothermal wells.

Some shallow wells erupt periodically, much as natural geysers. Other wells drilled to greater depths and higher temperatures erupt continuously once the process has started, provided that sufficiently permeable aquifers have been intersected. At Steamboat Springs, $CaCO_3$ precipitates in pipes and casing because of chemical changes induced by loss of CO_2 to the vapor phase. As $CaCO_3$ is deposited, discharge of the well decreases to such an extent that the well must be shut down and cleaned. Erupting wells that are left uncleaned commonly change from continuous eruption to periodic eruptions, much like natural geysers. This detailed study of wells and drill holes of the Steamboat area contributes much to the understanding of geothermal phenomena elsewhere, including problems of exploration for geothermal energy.

The total surface discharge of hot water in the immediate thermal area, including springs, wells, and unseen discharge directly into Steamboat Creek, is about 600 gpm. A nearly equal quantity flows northward beneath the surface for distances of 1–5 miles, eventually escaping in part as visible springs of moderate temperature, but mostly as unseen discharge directly into the creek. The total quantity of thermal water, visible as well as unseen but detected from discharge and chloride measurements of the creek, is about 650 gpm of the total of 1,130 gpm calculated for the whole system.

Temperature relationships and circulation patterns within individual systems of the area are indicated by data from the wells and drill holes. The Main Terrace system is probably the best understood, but many details are lacking.

The total heat flow from a system of boiling springs, steam vents, and hot ground with great local differences in temperature and discharge of fluids is extremely difficult to measure directly. A much more reliable and time-saving method developed in this report can be used provided that (1) a "levelling-off" or base temperature of upflowing water is indicated by drilling, (2) the thermal water contains a high content of chloride, boron, or some other soluble constituent that is present only in low concentrations in the normal surface and ground waters of the area, and (3) all the liquid water of the system is eventually discharged into a stream that can be monitored to detect chloride or other tracers contributed by the thermal water. Before boiling, the water of the Steamboat system has a temperature close to 175°C, and its chloride content is 800–820 ppm. The total flow of heat contained in this water is nearly 12×10^6 cal per sec, which is equivalent to the conducted heat flow from 780 km^2 of "normal" crust (assumed to be 1.5 cal per cm^2 sec, or 1.5×10^4 cal per km^2 sec).

The surrounding basin areas are examined to determine whether their heat flow is low and is possibly channeled by convection into the Steamboat system. No reliable heat-flow data are available from these basin areas, but geothermal gradients in the best known parts are very high and indicate that the heat flow is at least two times the crustal average. Very scanty data from the Virginia and Carson Ranges suggest that these areas, too, have some excess above normal heat flow.

The total heat flow from Truckee Meadows basin is probably at least 20×10^6 cal per sec, and the local anomaly above the regional average is probably 15×10^6 cal per sec. With reasonable assumptions for temperature, heat capacity, and heat of crystallization, the probable minimum of excess heat flow is equivalent to a supply of 0.001 km^3 of granite magma per year. The age of the Steamboat Springs system is at least 100,000 years and is probably nearer 1,000,000 years. A magma supply of at least 100–1,000 km^3 is required through the life of the system.

A batholith that is intruded into the shallow crust and then remains static as it cools and crystallizes is not a satisfactory model, unless the fissure system controlling the circulating water can gradually extend deeper into the batholith as stored heat is removed at higher levels by circulating water. An attractive alternative for the heat-flow problem involves convection within the magma chamber to maintain magmatic temperatures near the base of hydrothermal circulation.

INTRODUCTION

LOCATION

The Steamboat Springs area is in southern Washoe County near the western border of Nevada (pl. 1). The most intense thermal activity straddles the common boundary of the Virginia City and Mount Rose quadrangles (Thompson and White, 1964, pls. 1, 2). The most active part of the area is just west of Steamboat Creek, a tributary of the Truckee River that heads in Washoe Lake about 9 miles south.

The springs emerge near the northeastern end of Steamboat Hills, which trend northeastward, transverse to the regional northerly trends; the hills lie near the axis of a chain of basins between the north-trending Virginia and Carson Ranges. On a larger scale, these ranges split northward from the northwestward-trending Sierra Nevada Range, whose main crest at the latitude of the hot springs lies 20 miles west of the Carson Range, separated by the Tahoe basin.

PURPOSE AND SCOPE

Steamboat Springs has provided an opportunity for fundamental research on hydrothermal activity, ore-metal transport, and ore deposition (White, 1955a, 1967a). A very wide variety of geothermal processes are taking place here under natural conditions that cannot be duplicated in the laboratory. Spring systems of this type are a phase of volcanism in which the cooling of a magma body of batholithic proportions (White, 1957a, p. 1642) is probably accompanied by separation of a vapor phase at high temperature and pressure. The present rate of heat flow from the Steamboat hydrothermal system is 12×10^6 cal per sec (calories per second), which is equivalent to the heat flow from about 780 km^2 of "normal" area; heat is likely to have been flowing at this rate for at least 100,000 years, probably closer to 1,000,000 years. The convecting water that transports this heat is overwhelmingly of surface origin, but some H_20, $C0_2$, H_2S, S, B, and other substances are probably separating from a magma body at depth, perhaps as a vapor phase at high temperature and pressure.

Results of the Steamboat Springs studies are being published as a series of professional paper reports. A paper by Thompson and White (1964) provides the regional geologic setting. The second of the series (White and others, 1964) is a detailed geological and geophysical study of the thermal area. Another report (White and others, 1966), not in this series, concerns the isotope geology and geochemistry of the thermal and cold waters of the area.

The present report deals with the physical aspects of hydrology and thermal activity of the spring system. Many of the geologic conclusions developed in the earlier papers are essential to an understanding of the hydrology of the spring system and are herein summarized.

FIELDWORK, METHODS OF STUDY, AND ACKNOWLEDGMENTS

Fieldwork was carried on from 1945 to 1952. The thermal area is shown in figure 1, and the geology is generalized on plates 1 and 2.

The Low and Main Terraces are shown in detail on plate 3. Topography was mapped by Robert G. Reeves, Hale C. Tognoni, and Donald E. White; geology was mapped on the topographic bases by White.

Systematic measurements of spring activity were made from June 1945 to August 1952. The interval between measurements was generally 1 week, but daily or biweekly measurements were also made for short periods to provide additional detail.

The rate of discharge and the vent temperature were measured in most individual springs that were discharging at rates of more than one-quarter gallon per minute. Discharge was measured either by V-notch wier or by determining by stopwatch the filling time of a can of known volume. The latter method was generally used for springs of small discharge, diverted through an iron pipe. Temperature was measured by maximum-recording mercury-glass thermometers. All thermometers had been standardized at two or more points from 0° to 100°C. Unstandardized maximum-recording thermometers may be in error by as much as 2°C or more, but discharge temperature and other temperatures given in this report are probably seldom more than 1°C in error.

A water sample was collected at the time the systematic physical measurements were made for each discharging spring. In laboratory space provided at the School of Mines, University of Nevada, Reno, these water samples were analyzed for chloride content; pH and specific conductance were generally determined monthly. Chloride was found to be the most significant and reliably determined indicator constituent in the water. Cl was titrated by a standardized $AgNO_3$ solution, using potassium chromate as an indicator under yellow light in a darkened room. A standard solution containing 1,000 ppm (parts per million) Cl was titrated at the start of each series. Some samples analyzed in Reno were also analyzed by W. W. Brannock, of the U.S. Geological Survey, Washington, D.C. The percentage difference between check analyses of normal Steamboat Spring waters (600–980 ppm Cl) was nearly always less than 1 percent of the content reported, and generally was within about 4 ppm.

Whenever an opportunity was provided to obtain data at depth, detailed measurements were made as drilling progressed in new wells and drill holes in and near the thermal area. Most commercial thermal wells do not exceed 200 feet in depth and were drilled by cable-tool (churn) equipment. Cable-tool drilling introduces no foreign water except when a hole is too "dry." For this reason, this method is superior to all other methods in providing opportunity for excellent bottom-hole temperatures and water samples as drilling progresses; the original nature of the rocks must be interpreted from broken rock fragments. Normal practice is to drill 8 hours per day, 5 days per week; the present study demonstrates that thermal equilibrium in churn-drill holes is generally approached within several degrees Centigrade after a layover of 16 hours; closer approaches to thermal equilibrium were indicated by repeat measurements made over weekends without drilling.

A weighted maximum-recording mercury thermometer suspended on steel tape or wire line was worked down as deeply as possible into the bottom-hole sediment. To insure recording of the maximum temperature attained, only standardized thermometers with a very narrow restriction at the separation point in the mercury column were used. If a restriction is too large, mercury escapes downward through the restriction to the bulb as the thermometer is being raised through

FIGURE 1.—OBLIQUE AERIAL PHOTOGRAPHS OF STEAMBOAT

A. View west-southwest along axis of Steamboat Hills with Carson Range in distance.

higher and cooler parts of the hole. In several drill holes, temperature decreased locally with increasing depth; a thermometer lowered into such a hole records the maximum temperature attained, which may not be at the bottom of the hole. If a temperature reversal is suspected for any reason, the maximum-recording thermometer must be insulated and left in the hole long enough to insure a good measurement, and then raised rapidly through any zones of higher temperature. A seven-unit thermistor cable was used in some measurements, but temperature-stable materials were not available at that time (1950), and the equipment soon failed.

In 1950 the Geological Survey contracted for the diamond drilling of eight holes with a total footage of 3,316 feet. Recovery of core exceeded 90 percent of drilled footage. Although this method is superior to others in providing core for chemical and petrographic study, introduction of cold water requires some time for attainment of thermal and fluid composition equilibria. Thus, temperatures measured in diamond-drill holes are somewhat less reliable than those measured in churn-drill holes.

SPRINGS. PHOTOGRAPHS BY ERNIE MACK, RENO, NEV

B. View west over the central and northern part of the thermal area.

The objective of all these physical and chemical measurements was to learn as much as we could about this spring system in four dimensions, including time. In many but not all respects, this study is probably the most exhaustive yet made of a hot-spring system, but many questions still have no clear answers.

Detailed measurements of spring activity were made first by White and later by Robert G. Reeves, Hale C. Tognoni, Douglas Baker, William Ebert, Robert Horton, William Reinkin, James Scott, and R. K. Vassar. The assistance of all these men is much appreciated. S. F. Turner, formerly of the Water Resources Division, U.S. Geological Survey, assisted in initiating the hydrologic studies, and the preliminary results were reviewed in 1949 by C. V. Theis, also of the Water Resources Division.

I am particularly grateful to Vincent P. Gianella, Professor Emeritus of the University of Nevada, who contributed much to this work, especially during the early stages. Gianella furnished notes made over many years and also supplied unpublished material by L. H. Taylor and J. C. Jones prepared in 1916. The Taylor

TABLE 1.—*Summary of observations of Steamboat Springs, Nevada, before 1945*

Reference and date of publication (Date of observation in parentheses, if known)	Activity of springs and geysers	Depth to water in fissures, Main Terrace (feet)	Elements noted	Spring deposits	Other observations and remarks
Laur, P. 1863	No details.		Fe, Mn, Cu, Au, S.		Earliest technical reference; ore deposits stated to be in process of formation.
DeQuille, Dan 1876 (1860–62 and later) Republished (p. 327–329) 1947	Many boiling springs on sides of terraces and ends of fissures. 1 geyser active in 1860, stated to erupt 3-ft column more than 50 ft high. Another in 1862 spouted 50–60 ft high through a 3-in. orifice.		S.		Crevices more than 1 ft wide with surging boiling water at depth. May be the earliest account of geyser action in United States.
Blake, W. H. 1864	Hot water and steam at extreme ends and flank of spring mound. Some spout up and overflow at intervals; 1 is intermittent, with about 4-min. interval.	"About 10".	Mn.	Chiefly amorphous SiO_2, crusting springs and fissures.	2 or 3 fissures 3–12 in. wide; open fissures believed to result from widening of fissures in bedrocks.
Stretch, R. H. 1867 (1866)	Numerous springs, some from fissures, others from isolated vents. Most noticeable spring has intermittent discharge, 6¼-min interval, from basin 2–3 ft in diameter.				Terrace 1 mile long, ¼ mile wide. Bedrocks, basalt on granite.
Browne, J. R., and Taylor, J. W. 1867	Some springs have "tidal action," followed by subsidence that may last for months or years.				
Browne, J. R. 1869 (1867?)	Springs very active, highest temperature 204°F (95½°C). Springs formerly more extensive (recognized old sinter of Sinter Hill and High Terrace?). Springs discharge intermittently from fissures. One intermittent spring can be soaped into erupting 6–8 ft high.				Springs reportedly more active when discovered 20 yrs earlier. Fissures 6–12 in. wide, apparently caused by upheaval from below.
Phillips, J. A. 1871	Some fissures overflow, waters slightly alkaline.		S, Fe oxides, CO_2, H_2S, Mn, Cu.	Banded siliceous incrustations on fissures are like veins as much as several feet wide.	Probably most remarkable hot springs yet discovered. 5 fissures, some with violent boiling. Character of water and gases determined; High Terrace is an old spring system, still has some steam, CO_2.
Phillips, J. A. 1879 (Same visit as 1871?)			Hg discovered (after visit).	Fissure fillings alternately amorphous and crystalline; some quartz(?).	Fissures formed by repeated widening.
Whitehill, H. R. 1877 (1875–76)			HgS, S.		Thomas Wheeler discovered cinnabar (silica pit area) Dec. 3, 1875, in "granulated clay and sand" with HgS and S. Hot vapors depositing native S; some HgS later than S.
Hague, Arnold, and Emmons, S. F. 1877	Several springs in conical mounds; only 1 or 2 small basins continuously filled with water. Several intermittent springs and geysers erupt 10–20 ft high.	10 or 15.	S, sulfur gases.	Siliceous sinter from springs, analyzed.	A number of parallel fissures 2–12 in. wide, generally open but some choked by debris; violent boiling at intervals.
LeConte, Joseph 1883	Only a few active vents east and north of main fissures; water hot and alkaline. Feeble geysers, "once more violent." Water in fissures, violently agitated.	7 or 8.	Hg, S, Fe.	Sinter believed 15–20 ft thick. Springs cover and choke their vents.	Northern part Main Terrace not very active. Fissures tortuous, ragged, probably 20–30 ft deep. No deposits on fissure walls. Mercury mine (silica pit) in rocks decomposed by acid from H_2S with no sinter present.
Becker, G. F. 1887 (1883–85)	Springs extremely hot, some at boiling point.		Au, As, Sb, Hg, S.		Ore minerals being deposited; laboratory investigation of sulfide transport. Induced precipitation of metastibnite on glass siphon at spring 8(?).
Becker, G. F. 1888 (1883–85)	Springs numerous; discharge reportedly higher in years of heavy rainfall and in spring. Some from pipelike vents at crests of smooth mounds; some discharge "fitfully," some are true geysers with periodic discharge. Eruptions to several feet reported, but not seen by Becker.	No data.	Hg, Sb, As, Pb, Cu, Co, Au, Ag, Zn, S. (No crystalline sulfides identified.)	Largely siliceous, minor carbonate deposits probably 50 ft thick. Sinter claimed in mercury mine area (silica pit).	Individual springs short lived, changing. Fissures are cracks in earlier formed sinter, 1 in to 2 ft wide. Much of Main Terrace inactive. Metals believed derived from granite, transported in alkaline sulfide solutions. Small enclosed basin of silica pit area formed by acid leaching and subsidence. First geologic map of thermal area. No samples or observations precisely located.
Pošepný, Franz 1902			Hg, S, CO_2.	Sinter deposit incorrectly claimed to be principally $CaCO_3$, 15 meters thick.	Starting in 1878 a fissure was opened by adit 15 meters below surface; vein matter mined as quicksilver ore (probably refers to silica pit area but may be old adit on east flank of Main Terrace).
Lindgren, Waldemar 1903 (1885 as Becker's 1906 assistant; 1901).			Sb, Fe, As, S.		In 1901, 40-ft shaft sunk at present Steamboat Resort. Gravel coated with stibnite needles, pyrite, marcasite, and opal.
Lindgren, Waldemar 1933 (1885, 1901, and later)			Same as Becker.	Impure SiO_2 or mixture with $CaCO_3$; older deposits of chalcedony, quartz.	Sb as amorphous metastibnite in quantities large enough to color some sinter red. In shaft, Sb as stibnite with black opal, pyrite, or marcasite.
Jones, J. C. 1912 (1912?) 1914	Stibnite-depositing spring 86°–93° C. Water analyzed by Cullen and Jones; may be spring 34, pl. 3.		Sb.		Crystalline stibnite found on edge of a pool, deposited at surface, not transported.
Jones, J. C. (1) (1916)	Discharge ~20 gpm from Low Terrace at 65°–85°C and about 180 gpm (Taylor's measurements?) from Main Terrace, with temperatures as high as 95°C. Springs on north ends of fissures and on east slope.	3–6.		Low Terrace, 1–40 ft thick; Main Terrace, about 100 ft thick, largely pure silica.	Shaft 50 ft into High Terrace with warm water in bottom; at south end, 100-ft well with water level <60 ft. Low Terrace discharge ~20 gpm.

1 1916, unpublished.

TABLE 1.—*Summary of observations of Steamboat Springs, Nevada, before 1945*—Continued

Reference and date of publication (Date of observation in parentheses, if known)	Activity of springs and geysers	Depth to water in fissures, Main Terrace (feet)	Elements noted	Spring deposits	Other observations and remarks
Taylor, L. H. (2) (1916 and earlier)	Discharge of single springs as much as 52 gpm (measured) with total of 179 gpm from Main Terrace. Discharge from Low Terrace estimated 50 gpm. Geyser at north end of Main Terrace (at or near vent 21-n).				First detailed topographic map showing fissures and individual springs. Estimated total discharge of system 900–1,400 gpm. Shaft described by Lindgren (1903) deepened to 70 ft, yielding 300 gpm by pumping.
Allen, E. T., and Day, A. L. 1935 (1924, 1925)	In 1924 and 1925, noted 1 small intermittent spring, 2 small geysers; a large geyser at north end of fissures erupted 25 ft high (identified by Gianella, oral commun., 1945, as 21-n).				
Gianella, V. P. 1939 (1930 and later) 1941			Hg, Sb, Cu, Fe, As, S.	Terrace largely siliceous; minor carbonate.	Erupting wells deposit $CaCO_3$; metastibnite deposited from erupted water at Steamboat Resort, matted stibnite and pyrite ejected from a well in gravels north of Main Terrace (Reno Resort of pl. 1). Many sulfate minerals in silica pit area.
Bailey, E. H. and Phoenix, D. A. 1944			Hg, S, SiO_2.	Older spring deposits contain Hg, none seen in recent sinter.	Most complete history of efforts to mine Hg and S; HgS in hydrothermally altered rocks and in old sinter; no evidence found for present-day deposition of Hg.

[2] 1916, unpublished data; see text.

data and map, the only quantitative record of discharge at Steamboat Springs that antedates the present work, are invaluable for comparison.

I am also greatly indebted to my colleagues Philip F. Fix, George Thompson, C. H. Sandberg, A. H. Lachenbruch, L. J. P. Muffler, R. O. Fournier, Philip Cohen, and J. D. Breschoeft for their contributions and stimulating discussions concerning geothermal phenomena.

PREVIOUS WORK

Numerous brief accounts of Steamboat Springs were published from 1863 through the following 25 years (table 1), ending with G. F. Becker's extensive study of 1888. Although Becker's report seems to have been accepted as exhaustive, his treatment was in fact general in nature and nearly devoid of specific data on mineralogy, hydrothermal alteration, and thermal activity. Le Conte's brief paper (1883) contains some specific details, and De Quille's popular account of geyser activity in 1860 and 1862 (1876) is of special interest because it seems to be the first published record of geyser observations in the United States.

Lindgren (1903), Jones (1912), and Gianella (1939) recognized the specific minerals containing antimony and proved that the stibnite and metastibnite (the amorphous red antimony sulfide) were being actively deposited.

The most useful records for comparing activity of the past with that of the present are unpublished material by J. C. Jones and by L. H. Taylor, quoted in part under "Previous observations" (p. C15–C17).

DEFINITIONS

Terms as used in this report are here defined.

Vent. Any megascopic opening between the general level of the ground surface and the water table. If the water table coincides with the general ground surface, a vent constitutes the discharge conduit of a hot spring or geyser. If the water table is below the ground surface, a vent may or may not be utilized by rising gases and steam. With time and appropriate changes in water level and temperature, an inactive vent may change to any of the different types of springs defined below.

Spring. A natural vent from which water discharges at the surface of the ground.

Intermittent spring. A spring characterized by intervals of discharge that alternate with intervals of no discharge. As will be seen, probably all springs of the area are intermittent if the period of observation is sufficiently long. A spring may discharge for days, weeks, or years and then cease to flow for an equally uncertain period of time and for a variety of causes. The interval between discharge of an intermittent spring, as this term is used in the present report, is ordinarily a week or less.

Pulsating spring. A spring with continuous discharge that changes in rate through a period generally ranging from seconds to minutes in a more or less cyclic manner. Vigorous, rapidly pulsating springs are also known as continuous or perpetual spouters.

Geyser. A hot spring characterized by intermittent discharge of water that is ejected turbulently and is accomplished by a vapor phase. The temperature of water ejected at the ground surface is generally near the boiling point of pure water at the atmospheric pressure for the altitude, but in some hot-spring areas, the dissolved gas content of the water is so high that a vapor phase exists at temperatures much below those of boiling of pure water for the prevailing pressures.

Geysers that have eruption temperatures 10°–20°C or more below boiling of pure water have been described briefly by Henderson (1938) and White and Roberson (1962, p. 414, 419). All hot springs with intermittent turbulent discharge are considered as geysers in this report without precisely defining temperature limits. At Steamboat Springs the average barometric pressure is 12.47 psi (pounds per square inch) or 645 mm of Hg, and the theoretical boiling point of pure water at this pressure is 95.36°C. Most of Steamboat's "boiling" springs and geysers are within a few degrees centigrade of this temperature at ground level, but vent 10 on the Main Terrace commonly discharges intermittently as a feeble gassy geyser at recorded temperatures not exceeding 79°C.

Subterranean geyser. The intermittent turbulent ejection of steam and gases in the subsurface parts of a vent; water is ejected above the general water table but the water table is too deep for liquid water to appear at ground level; all erupted water that remains liquid eventually flows back into the system. With increasing depth of water table below the ground surface, the vigor of eruption must increase in order to discharge water at the surface.

Fumarole. Any vent that discharges steam and other gases but not liquid water. Temperature ranges from much below the boiling point of pure water up to magmatic temperatures. Temperatures of fumaroles at Steamboat Springs most commonly range from about 70°C to 96°C, which is slightly above the average boiling point of pure water (95.36°C). Low-temperature fumaroles are here considered synonymous with solfataras.

Aquifer eruption. The intermittent ejection of water from an individual aquifer. This intermittent ejection can be expressed at the ground surface or water table by geyser or subterranean geyser activity, or it can be combined with continuous discharge from other aquifers to produce changes in rate of discharge of a spring or well tapping these aquifers. The concept that two or more aquifers can discharge continuously or intermittently and at different rates and cyclic intervals is helpful in understanding the complex behavior of many geysers, pulsating springs, and geothermal wells. Steamboat well 4, described in this report, provides a clear example of the phenomena.

GENERAL RELATIONSHIPS

SOME CONCEPTS OF GEOMETRY AND HYDRODYNAMICS OF HOT-SPRING SYSTEMS

A hot-spring system that consists entirely or predominantly of meteoric water is a huge convection system. Water from rain or snow seeps underground somewhere on the borders of the system and circulates downward along interconnected channels to a region of higher temperatures at depth. As the water increases in temperature it expands and becomes lighter despite increasing pressure at depth. This heated water is driven onward along the channels, being displaced by cooler and heavier water entering the system. Hot water is eventually driven out as discharging hot springs, which generally emerge in or near topographic lows. A simple system is illustrated by Darton's section (1906) through Thermopolis, Wyo., here reproduced as figure 2.

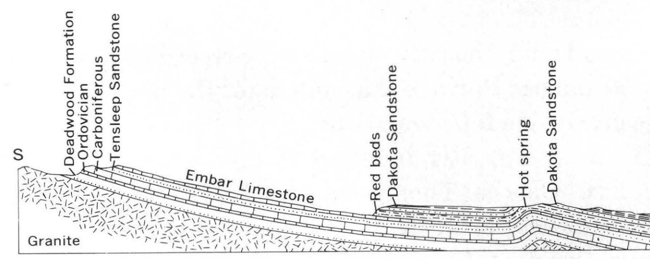

FIGURE 2.—Structure section through the hot-spring system of Thermopolis, Wyo. After Darton (1906).

The driving force of a hot-spring system is best considered as two separate factors related to (1) Difference in altitude between recharge and discharge areas. The recharge area is ordinarily the higher in altitude with a positive driving force, as in figure 2, but in special cases the recharge altitude can be lower than the discharge altitude. (2) Differences in density of water in the cold heavy downflowing part of the system and that in the hot light upflowing part of the system. Actual density distributions of water in the system depend on temperatures, pressures, salinities, and proportions of vapor phase throughout the system. Figure 3 illustrates a system slightly more complex than that at Thermopolis; two simple flow paths are shown, with a graph of temperatures and depths that are reasonable for one of the paths. In most but not all systems, temperature is much more important than pressure and salinity in determining density differences.

The Steamboat Springs system must be far more complicated than shown in figure 3. Crystalline granitic and metamorphic rocks of very low permeability underlie the whole region at all depths, except in a shallow cover that is generally less than 2,000 feet. Regional mapping (Thompson and White, 1964) failed to

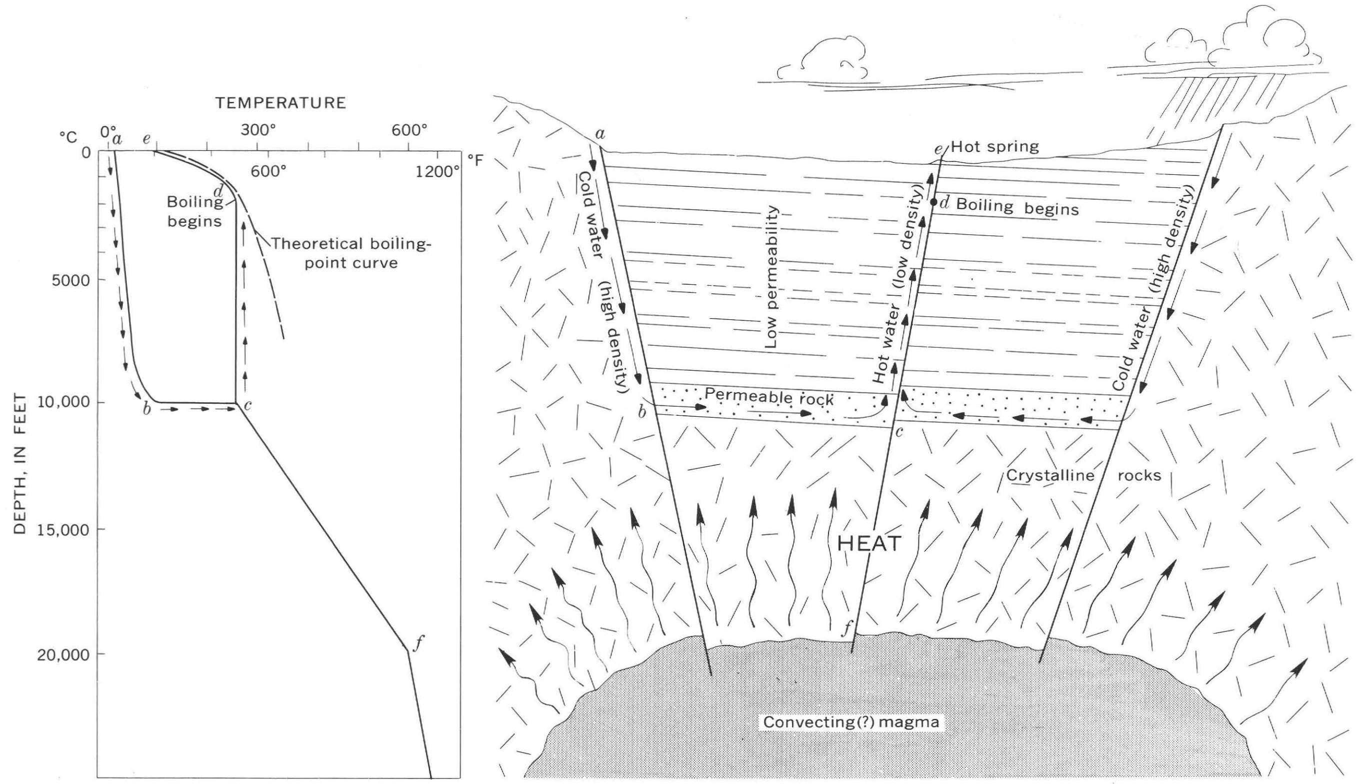

FIGURE 3.—Simple high-temperature hot-spring system with deeply circulating meteoric water assumed to be heated entirely by conduction.

reveal major faults that intersect in any simple system of interconnected channels. A much more complex system, with many different flow paths, that is considered more representative of the Steamboat Springs system is shown in figure 4.

Laboratory-determined permeabilities of small specimens of these crystalline rocks would be exceedingly low, but bulk permeabilities of these same rocks in place with interconnected fractures should have positive values differing greatly from one location to another.

The permeability of a rock is generally considered to decrease with increasing depth and pressure. But in a convection system where temperatures are high, the density differences related to thermal expansion provide a powerful pressure drive that increases with increasing depth of circulation, tending to counterbalance decreasing permeability. Another factor favoring deep circulation (for a given mass permeability under standard STP conditions) is the decreasing viscosity of water with increasing temperature. Viscosity of water at 45°C (the average temperature here assumed for the downflowing column) is 0.0060 poise but is only 0.0016 poise at the temperature of 170°C assumed for the upflowing column.

A third factor that may help to explain deep circulation is the large increase in solubility of silica with increasing temperature throughout the range considered here. As water is heated, it may dissolve quartz and perhaps silica from other minerals, thereby increasing the permeability of a given channel with time.

In the Steamboat Springs system, any recharge from the base of the Virginia Range occurs at an altitude about 120 feet higher than the water table in the crest of the Main Terrace, a difference equivalent to a pressure drive of about 50 psi. Recharge near the base of the Carson Range has an altitude advantage of about 1,300 feet, or nearly 600 psi.

The driving force related to differences in density in the Steamboat system is due almost entirely to differences in temperature. In the upper 10,000 feet of the earth, hydrostatic pressures in a convection system will seldom exceed 4,500 psi. Density differences due to pressure are only a little more than 1 percent at all temperatures as high as 100°C, and even these differences are almost entirely counterbalanced on the two sides of a convection system. The waters of highest salinity in the Steamboat system have about 2,500 ppm of dissolved constituents and a density that is only about 0.2 percent higher than pure water at the same temperature. In contrast, the relative density of pure water is 1.000 at 4°C, 0.958 (4.2 percent less) at 100°C, 0.917 at 150°C, 0.863 at 200°C, and 0.794 at 250°C. Within this range

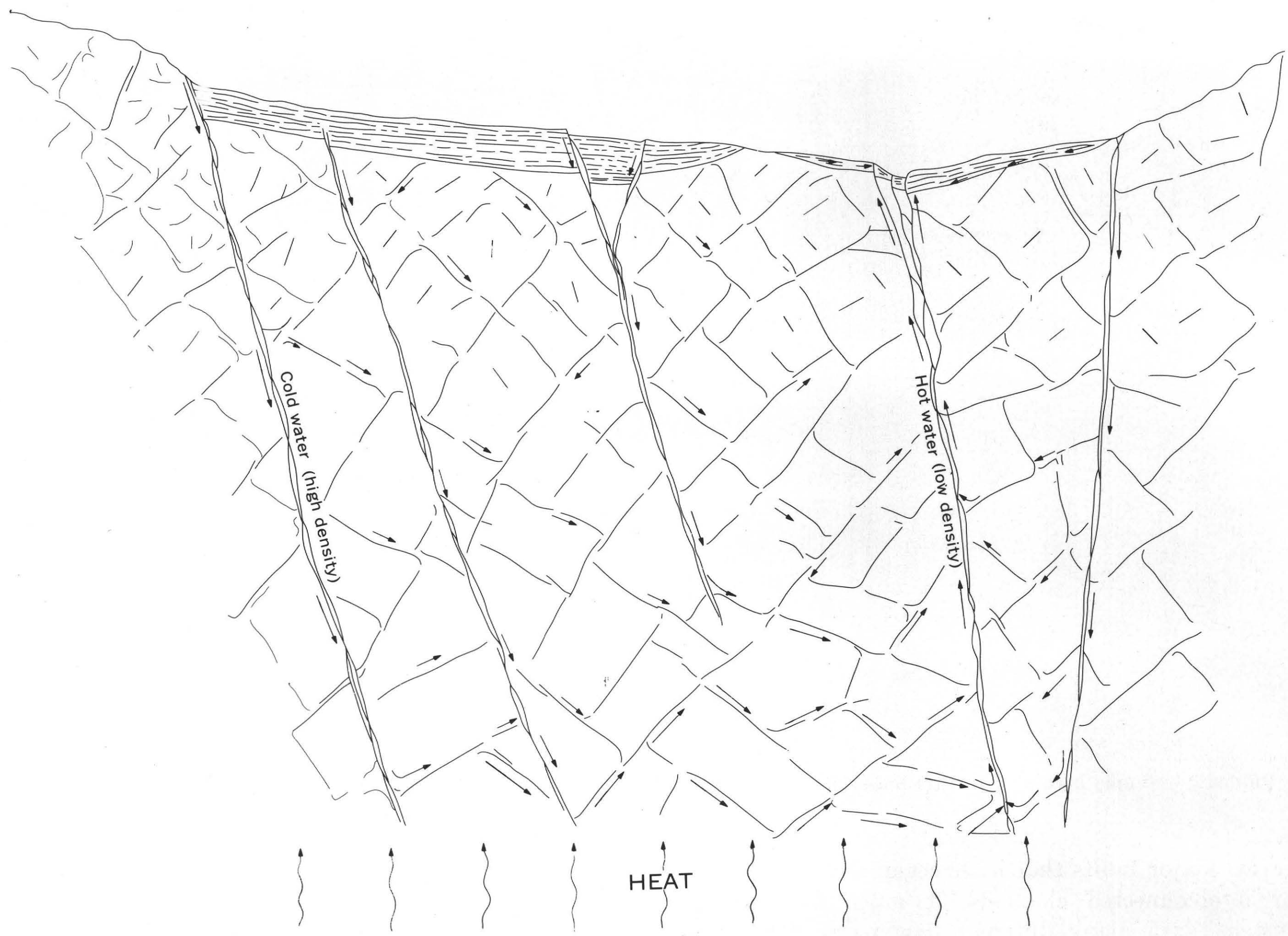

FIGURE 4.—Diagrammatic representation of a thermal-spring system largely in fractured crystalline rocks. Permeability decreases downward but is more than offset by decrease in viscosity and increase in driving force related to thermal expansion (see text).

in temperature, differences in density related to thermal expansion can be more than 20 percent as compared to salinity effects of about 0.2 percent and pressure effects that probably are much less than 1 percent.

The circulation of water in idealized homogeneous porous media that is related to effects of thermal expansion has been treated mathematically by Donaldson (1962). Wooding (1956) derived partial differential equations for liquid flow in saturated homogeneous permeable solids caused by thermal expansion and, with simplifying assumptions, applied the results to Wairakei, New Zealand. Bredehoeft and Papadopulos (1965) considered vertical rates of ground-water velocity in homogeneous media as related to temperature differences, but their model did not consider effects of thermal expansion.

In a few hot-spring systems that have been thoroughly explored for geothermal energy, the pattern of temperature distribution is reasonably clear within the very limited areas and depths of each total system that constitute the main economic interest. In figures 2 and 3 as examples, the main economic interest in most geothermal drilling to date can be considered as extending from surface expressions of thermal activity down to depths of 1,000 to seldom more than 3,000 feet. The permeable reservoir hypothesized at a depth near 10,000 feet in figure 3 would not have been discovered; distribution of temperature and pattern of circulation in the large noneconomic part of the total is not known for a single hot-spring system in the world! In addition, probably every hot-spring system is far more complicated than any simplified model that can be treated mathematically. These models, however, are still very useful in describing some of the problems within broad limits.

Temperatures in the downflowing and upflowing parts of a system are strongly dependent upon rates of flow of water and heat through the system, as shown by Donaldson (1962) and Bredehoeft and Papadopulos (1965). In figure 3, flow rates are assumed to be such that little change in temperature occurs from *a* to *b* or from *c* to *d*.

If the cold downflowing part is assumed to have an average temperature of 45°C with a density of 0.990 (relative to 1.000 for pure water at 4°C) and the hot upflowing part has an average temperature of 170°C with a density of 0.900, the pressure drive related to thermal expansion is equivalent to about 90 feet of cold water, or nearly 40 psi for each 1,000 feet of depth assumed for these temperature and density differences. If the depth of circulation is 10,000 feet, as suggested by White (1957a),[1] the driving force related to thermal expansion is equivalent to a head of 900 feet of cold water, or nearly 400 psi.

In each of the many explored hot-spring systems that show nearly constant temperature of upflow within a reservoir deep enough to exclude a vapor phase resulting from near-surface boiling, the rate of upflow must be relatively high. On the other hand, if the cross-sectional area of recharge is large, as in figure 4, and the rate of downflow is very low, an assumption of equally rapid rates of downflow and upflow is not valid. A temperature-depth curve for the average downflowing water, such as in figure 3, would tend to increase directly from point *a* to *c*, by passing point *b*. In many spring systems the average density difference between the downflowing and upflowing parts is probably considerably less than the maximum permissible from the temperature differences between recharge and upflow, and commonly may be nearer to one-half or two-thirds of this density difference. In the complete absence of data on rates of downflow for a single system, further speculation is unwarranted.

From the relations considered above, we see that the driving forces can differ greatly throughout a system and from one system to another. Potential recharge areas at higher altitudes are favored over those at lower altitudes if other factors are equal, but recharge from altitudes below those of the spring outlets is also possible. Countless numbers of interconnected channels no doubt exist, as suggested in figure 4, each with its own characteristics of permeability and heat supply. A critical factor of each potential channel is its most restrictive "bottleneck" that provides the greatest resistance to increased flow.

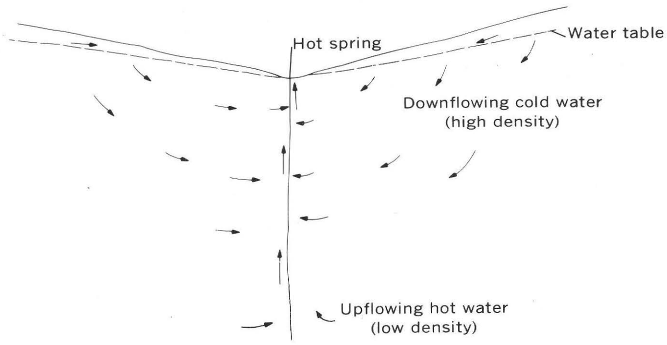

FIGURE 5.—A thermal-spring system that discharges in a topographic low, which also coincides with the local base level for the water table of the area. Cold water can mix with rising hot water at all depths.

A thermal-spring system that discharges in a topographic low and also coincides with the local base level for the water table of the surrounding area is represented in figure 5. In this system the differences in density provide a positive energy drive that permits cold water to enter the channels of thermal upflow at all depths, including those just below the ground surface.

Some spring systems do not discharge at points of lowest local altitude, either because of structural control or because the springs have formed silica or calcium carbonate deposits that have raised the outlet altitude above that of the surrounding ground. Figure 6 represents such a system. In this example pressures in the upper parts of the system exceed hydrostatic pressures related to the surrounding water table at lower alti-

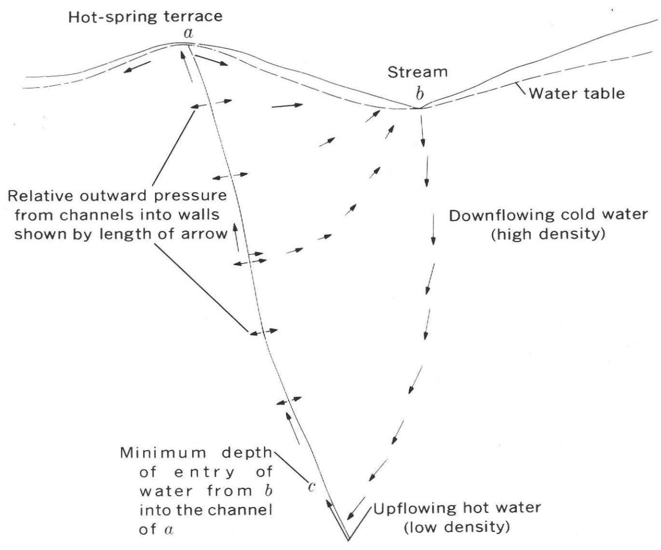

FIGURE 6.—A thermal-spring system that mounds the water table above that of the surrounding area.

[1] Such a depth would provide sufficient hydrostatic pressure to permit a vapor phase at or near the critical pressure of water. Alkali chlorides are characteristic constituents of most of the subsurface waters of volcanic areas; their solubilities are very low in ordinary low-pressure steam, but dense high-pressure steam has the solvent properties of liquid water. Thus, if meteoric water circulates to sufficient depth, alkali chlorides can be transferred continuously from magma to the surface. The evidence for some vapor transfer of alkali chlorides as opposed to other possible sources was reviewed briefly by White (1957a) and will be considered in more detail in forthcoming publications. A small proportion of magmatic vapor under high pressure would dissolve completely in liquid meteoric water near the base of the convection system.

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tudes. If the rock permeability is not nil, some water escapes into the walls. Depending on the differences in altitude between points *a* and *b* of figure 6 and the density differences between upflowing and downflowing columns of water, there is some minimum depth *c* above which cold water from point *b* cannot enter the hot-spring system.

At Steamboat Springs we can compute the depth of *c* with the following assumptions: (1) the difference in altitude between *a* (water level, crest of Main Terrace) and *b* (surface of Steamboat Creek) is 110 feet; (2) the average density of downflowing water from *b* is 0.990 and upflowing water to *a* is only 0.900. Solving the algebraic equation $x \cdot 0.99 = (x+110) \cdot 0.90$, the minimum depth of entry of water from *b* is found to be 1,100 feet below *b*, or 1,220 feet below *a*.

As will be seen, a little dilute water does enter the upper part of the Steamboat system, presumably from the small drainage basins in the highest western part of the thermal area, but the principles of figure 6 apply to all potential recharge areas at altitudes lower than the springs.

Topography, water table, temperature relationships, structure, and permeabilities of the rocks are all important factors in determining the characteristics of the upper part of a spring system. Additional complications can be caused by the presence of a vapor phase and of dissolved matter in the water. The vapor phase tends to decrease the average density of the fluids, whereas dissolved salts have an opposite effect.

If a geothermal system is structurally complex in its upper part, as is the Steamboat Springs system (pls. 1–3), thermal water can rise along more than one fault. If all potential channels of upflow had highly permeable interconnections, all thermal water would discharge from the fault that reaches the surface at lowest altitude. But at Steamboat Springs, thermal chloride water of about the same chemical composition is found at altitudes that range from 4,575 feet at Steamboat Creek to 4,680 feet in the Main and High Terraces. This chemical similarity is evidence that the structures are interconnected at depth. The differences in altitude of water levels is evidence that no single structure is permeable enough to discharge all water of the system; the upflow is distributed among many structures, some of which have no immediate discharging springs because the water level is below rather than at the surface (note water levels in the sections of pl. 2). Some of these thermal subsystems, such as that of the High Terrace, are geothermally active in spite of the absence of springs. Presumably, their wallrocks are permeable enough for water to flow up fractures and into the walls below the water table. Subsurface discharge of this type is very important at Steamboat Springs. As we shall see in another section, only about 5 percent of the discharge of the whole system appears at the surface as springs or geysers. Most of the water escapes unseen below the surface and is discharged directly into Steamboat Creek.

SUMMARY OF GENERAL GEOLOGY

The general geology of the region (Thompson and White, 1964) and of the thermal area (White and others, 1964) has been described in detail and is summarized here.

The springs emerge from the northeastern end of Steamboat Hills, which form a small structurally positive area within a chain of structural basins that lie between the north-trending Virginia and Carson Ranges. Metamorphic and granitic rocks of Mesozoic age form the "basement." A cover of middle and late Tertiary volcanic rocks and volcanic-derived sedimentary rocks is perhaps as thick as about 2,000 feet in the surrounding area (Thompson and Sandberg, 1958, p. 1274), but it is very thin or absent within most of the thermal area. A basaltic andesite flow of the Lousetown Formation, and possibly a concealed shallow intrusion of Steamboat Hills Rhyolite, are locally of early Quaternary age and are the youngest near-surface volcanic rocks. A large still-hot magma chamber must underlie the area to account for the heat flow and some of the mineral constituents of the hot-spring system (White, 1957a).

A complex history of erosion alternating with alluviation throughout the Quaternary is evident from the surface geology and data from drill holes. In general, the hot-spring deposits are local facies of the sedimentary formations deposited during each period of alluviation. Each of these periods in turn may correlate with a Sierran glaciation (table 2). The hot-spring deposits consist almost entirely of siliceous sinter and very minor calcium carbonate. All primary sinter consists of opal, but most of the older deposits have been reconstituted into chalcedonic sinter.

Three well-defined systems of faults have been recognized in and near the thermal area (pl. 1). An east-northeast system is parallel to the axis of Steamboat Hills. Post-Lousetown movement is as much as 100 feet, and pre-Lousetown movement may be considerably greater. Northwest-striking faults control Pine Basin in the west-central part of the thermal area and are approximately contemporaneous with the east-northeast system. The third system consists of numerous faults that strike nearly north; some are relatively old, but many displace alluvium and sinter of middle-Pleistocene age and are therefore the youngest in the area. No fault

TABLE 2.—*Tentative correlation of volcanic and sedimentary deposits with soils and Sierran glaciations* [1]

Sierran glaciations and interglaciations	Volcanic rocks	Stream deposits	Hot-spring deposits	Soils [2]	Lacustrine deposits [2]
			Recent		
"Little Ice Age," A.D. 1750±		Alluvium	Opaline sinter	Post-Lake Lahontan and post-Tioga soils; submature.	Shallow post-Lake Lahontan lake deposits.
			Pleistocene		
Tioga Glaciation Interglaciation		Alluvium contemporaneous with Lake Lahontan	Opaline sinter (No deposits?)	Middle Lake Lahontan soil; mature.	Lahontan Valley Group: Sehoo(?) Formation.
Tahoe Glaciation Interglaciation		Alluvium Mud-volcano breccia	Opaline sinter (No deposits?)	Pre-Lake Lahontan soil; very mature.	Lahontan Valley Group: Eetza(?) Formation.
Sherwin Glaciation Interglaciation	Steamboat Hills Rhyolite Steamboat flows of Lousetown Formation	Pre-Lake Lahontan alluvium Pediment gravels Early deposits	Opaline sinter Post-Lousetown chalcedonic sinter		
McGee(?) Glaciation ?		Pre-Lousetown alluvium	Pre-Lousetown chalcedonic sinter		

[1] See White and others, 1964, p. B27, for original table, description of lithologic units, and sources of data.

[2] From Morrison (1961).

displacement is clearly younger than the youngest Pleistocene sediments.

The Steamboat Springs fault zone, the largest of the north-striking system, provides the structural control for the Low and Main Terraces. The dominant evidence favors eastward-dipping normal faults (White and others, 1964, p. B48–B50). Total movement may exceed 1,000 feet; nearly all movement was earlier than the local basaltic andesite flow of the Lousetown Formation. Fractures cut the opaline sinter deposits, but movement on the fractures is negligible. Open fissures of the Main Terrace are formed from fractures by acid leaching and disintegration of sinter adjacent to the fractures above the water table; the open fissures do not form by physical separation of the walls, as formerly supposed (see "remarks", table 1).

The Steamboat thermal area exists because of a combination of favorable circumstances. These include a long history of volcanism in the area; a large magma chamber calculated to have a volume of at least 100 km^3 that has evolved heat and perhaps water and mineral matter for at least 100,000 years; and favorable topographic, structural, and water-table relations.

ISOTOPE GEOCHEMISTRY

The isotope geochemistry of waters of the area has been described briefly by Craig, Boato, and White (1956) and White, Craig, and Begemann (1967). Steamboat Creek, the master stream of the area, heads in Washoe Valley and flows northward through the chain of basins and along the base of the hot spring terraces. Two main streams of the Carson Range, Galena and Whites Creeks, flow eastward toward the springs and discharge throughout the year. The principal streams of the Virginia Range, on the other hand, have little or no surface discharge during the long dry summers. Summer temperatures and rates of evaporation are relatively high in the chain of basins and along Steamboat Creek but are much lower in the adjacent ranges.

Variations in the isotopic composition of Steamboat Creek water are primarily due to evaporation during the summer months. The heavy isotopes, D and O^{18}, start to increase markedly in June, attain maximum values in August, and are again nearly "normal" during and after October. The evaporational trend line for Steamboat Creek is shown in figure 7, along with the unweighted average of all isotope analyses of the creek.

Water from depths of more than 150 feet in the South Steamboat well near the southern limit of plate 1 consists, from chemical and physical evidence, of meteoric water migrating into the upper part of the hot spring system (White and Brannock, 1950). Isotopically, water from this well is intermediate between the runoff of the Carson and Virginia Ranges and is very unlike average Steamboat Creek water (fig. 7).

The hot springs are nearly identical in deuterium content to water from the South Steamboat well, but they range from 2.0 to 3.5 per mil higher in O^{18}. This major shift in O^{18} content is similar to the shifts that have been observed in some other high-temperature hot-springs systems (fig. 7; Craig and others, 1956) and is best explained by exchange of oxygen between circulating meteoric water and silicate minerals that are

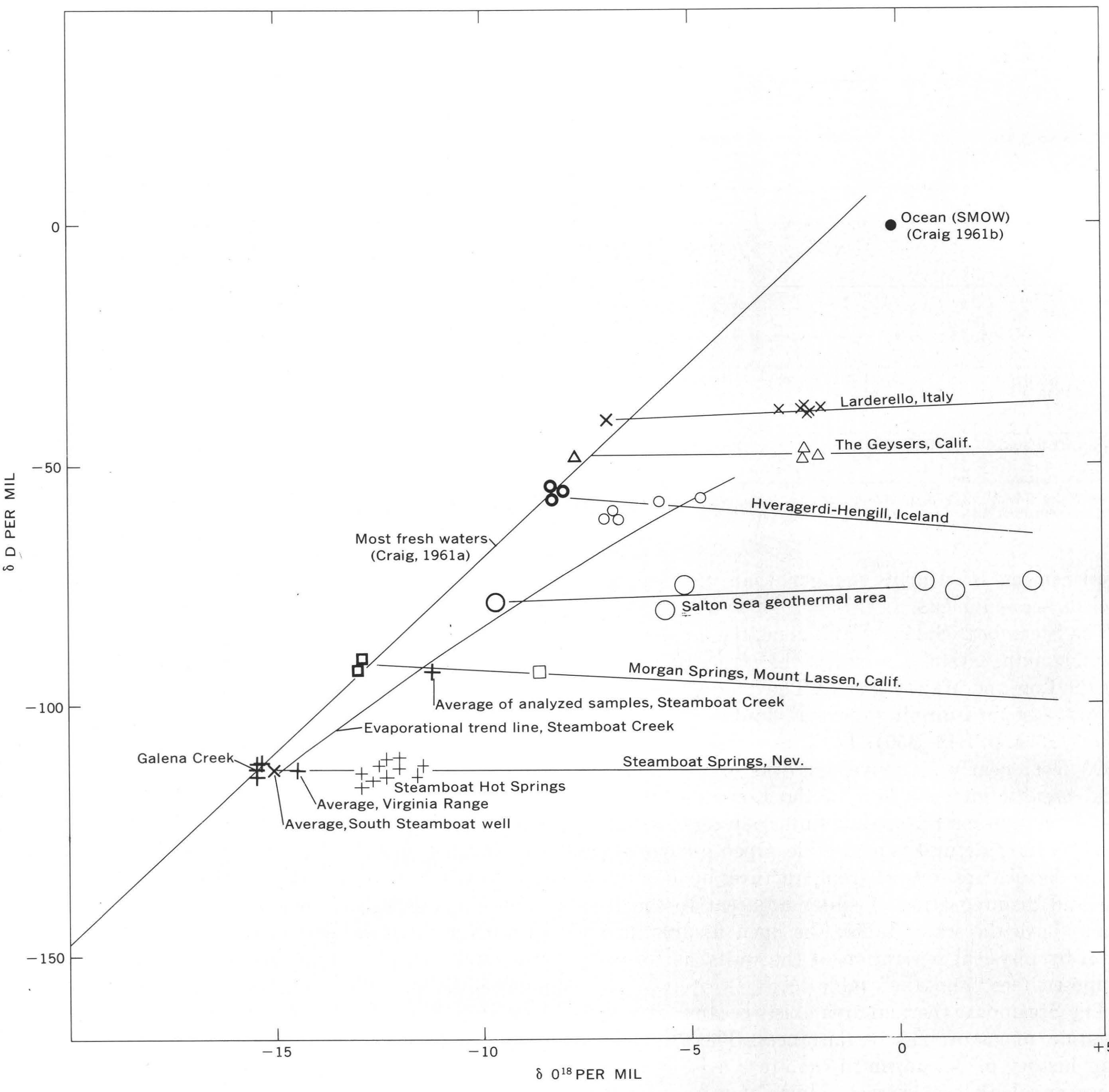


FIGURE 7.—Isotopic compositions of alkaline and neutral hot-spring waters and condensed steam, with associated surface waters (broad-line symbols). SMOW means "standard mean ocean water."

being hydrothermally altered. Other small variations have several different explanations, such as direct near-surface dilution of the deep thermal chloride water by water similar to that of the South Steamboat well (particularly evident, as will be seen, in the southern part of the Low Terrace, and by precipitation that falls in a small drainage basin, shown in the western part of plate 1, but seeps underground and into the Main Terrace system (evident in records from the Rodeo well).

The prebomb tritium content of Steamboat Creek is not known, but it was probably between 4 and 8 T units (1 T unit=1 tritium atom per 10^{18} H atoms). Water samples collected from South Steamboat well, spring 50 on the Low Terrace, and spring 16 on the Main Terrace, each prove the entry of some meteoric water of high T content and very short subsurface travel time (1 month or less); the greatly dominant component of each water is concluded to be of meteoric origin with an age of at least 50 years since isolation from direct contact with the atmosphere.

Chemical, isotopic, and physical evidence, some of which is described in detail in the present report, favors recharge of small proportions of young meteoric water from at least three specific local sources: (1) direct precipitation of rain and snow on the spring terraces; (2) precipitation in the small drainage basins just west of the spring terraces; and (3) in the Low Terrace, but probably not elsewhere, direct shallow inflow from Steamboat Creek.

Most of the total water is also of meteoric origin with an age of at least 50 years. The isotopic evidence does not prove the existence of any water of direct magmatic origin (Craig and others, 1956); an upper limit of the quantity that could reasonably escape isotopic detection is probably about 5 percent, but the actual amount present may be as little as 1 percent. As reviewed by White (1957a), a magma chamber is the most reasonable source for the huge quantities of CO_2, B, Li, Cs, As, and Sb that have been discharged during the indicated life of at least 0.1 million year of the spring system. If these substances were derived from a magma chamber, they must have been accompanied by some magmatic water or steam that is diluted beyond isotopic recognition by the dominant meteoric water of the system.

The water of South Steamboat well, and perhaps also of the whole spring system, may consist of a mixture from both the Carson and Virginia Ranges; recharge evidently occurs near the base of each range, before evaporation. Steamboat Creek, once viewed as the most reasonable source for recharge of the hot-spring system, is clearly eliminated on isotopic evidence as a major source.

DETAILED ACTIVITY IN THE THERMAL AREA

PREVIOUS OBSERVATIONS

Most of the early observers of Steamboat Springs were interested in qualitative rather than quantitative aspects of the activity. Table 1 contains a summary of the published observations.

Professor J. C. Jones and L. H. Taylor reported in 1916 on a plan to pipe hot water to Reno for space heating. The venture was not completed, but their material was left in the possession of Professor Vincent Gianella, who kindly permitted its use in the present study. Jones and Taylor's observations are significant because the data are more quantitative than any of the published records, and because their work was done shortly before the first geothermal wells were drilled to supply hot water for the resorts of the area. As will be seen in another section, the effects of the discharging wells on the spring system can be evaluated only in part. The natural activity as it existed in 1916 is therefore a very important part of the whole study. For these reasons, the most pertinent parts of their observations are quoted below.

Jones reported:

there is no water flowing from the fissures in the western area [High Terrace of present report, pl. 1], although a little steam may be occasionally seen on cold days rising from the fissures. A shaft has been sunk some fifty feet on one of the fissures [shaft at 2650W, Traverse 8 (White and others, 1964)] and warm water is standing at the bottom * * *. About a hundred yards to the south a well was drilled about one hundred feet deep * * *. Hot water was struck that heated the drill too hot to bear on the hand. Water now stands within sixty feet of the top of the hole * * *.

In the eastern area of sinter from which the springs are flowing, the spring deposits cover an area approximately 5,000 feet long by 1,000 feet wide [Low and Main Terraces of pl. 1] * * *. In the southern half (Low Terrace) the sinter deposit is from one to forty feet in thickness, resting on the disintegrated granite that underlies the spring area. A few short and irregular fissures and pools are found on the surface and the combined flow of water from them is about twenty gallons per minute. The temperature of the water ranges from 150° to 180° [F] and the water is piped from one of the larger pools to the bath house near the railroad station [Steamboat Resort of pls. 1, 3].

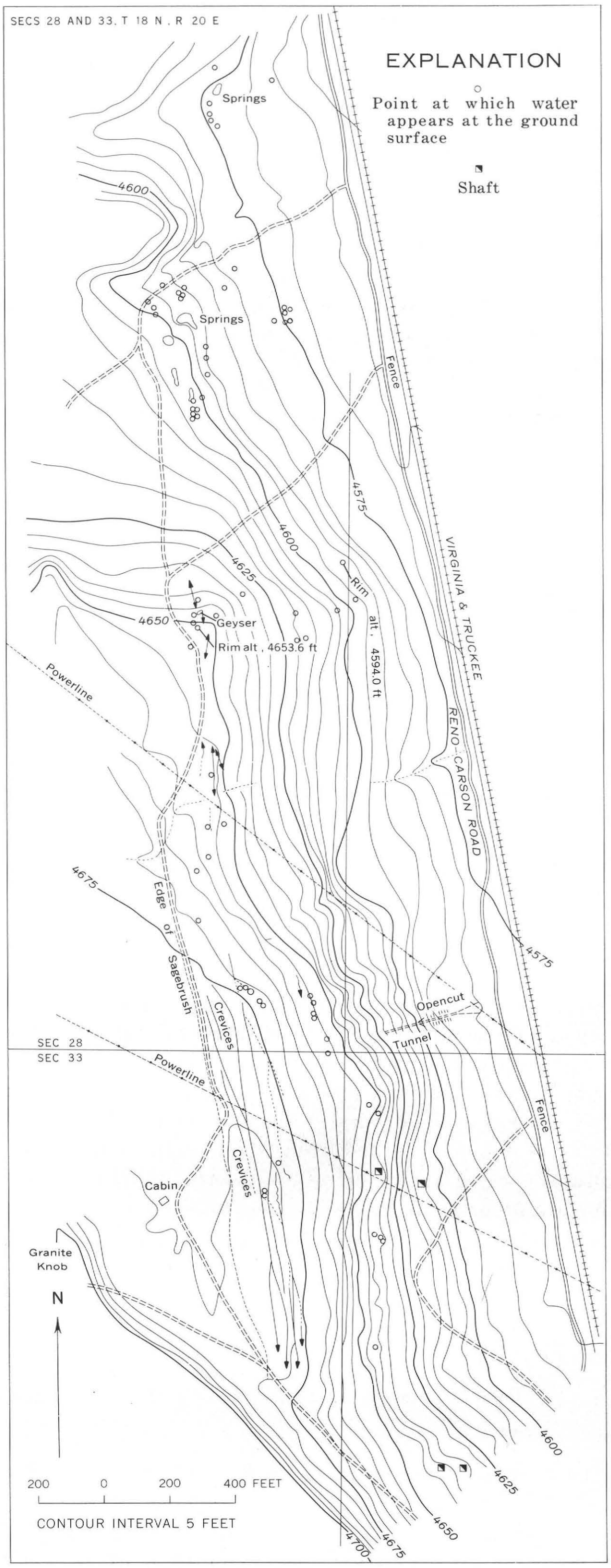
The northern half of the area (Main Terrace) comprises a terrace about 2,500 feet in length by a thousand in width and averaging 100 feet in thickness. This terrace has been entirely built up by the deposits of silica from the springs and rests on the decomposed granite below. On the upper surface of the terrace are several open fissures extending the length of the terrace. The boiling water may be seen at a depth of three to six feet below the surface and the fissures drain out at the northern ends, forming a considerable stream. Numerous craterlets and pools of boiling water are located towards the eastern border of the terrace and from their general alignment are probably located along one or two fissures running parallel to the main fissures.

Some years ago an attempt was made to drain the waters of the terrace by driving a tunnel at the base of the terrace about midway of its length [old caved adit of pl. 3]. At a distance of 100 feet a small crevice carrying hot water was struck and the operations brought to a standstill.

L. H. Taylor's work, dated October 30, 1916, contains the following:

The northern portion of the * * * eastern area of sinter [Main Terrace of present report], in my judgment, furnishes the most dependable source for a hot water supply. Here the boiling water is already issuing from the ground at many places, and in considerable volume.

The accompanying topographic map [reproduced here as fig. 8] shows this area with the points at which water appears at the surface of the ground indicated by small circles. The numbers on the map [unfortunately not shown on Gianella's copy] in-



This map, prepared by M. P. Hayes, accompanied an unpublished report by L. H. Taylor, dated October 1916

Figure 8.—Topographic map of the Main Terrace showing location of flowing springs, October 1916.

dicate where measurements of the flow were made, with results as follows reproduced here at table 3]:

Table 3.—*Spring discharge observed by L. H. Taylor, Main Terrace, September 25, 1916*

Taylor No.[1]	Discharge (gpm)	Type of observation
1	2.73	Measured.
2	.27	Estimated.
3	.55	Do.
4	2.27	Measured.
6	.63	Do.
8	42.64	Do.
9	3.64	Do.
10	13.45	Do.
11	2.00	Estimated.
12, 13	51.80	Measured.
14	25.90	Do.
15	5.00	Do.
16	2.50	Estimated.
17	10.00	Do.
18	6.00	Do.
19 [2]	10.00	Do.
Total	[3] 179.38	

[1] Not identified on only existing copy of L. H. Taylor's map, reproduced as fig. 8 of present report.
[2] Estimated discharge from small springs in southern part of fig. 8.
[3] Of total discharge reported by Taylor, 148.06 gpm is measured, 31.32 gpm is estimated. In addition, discharge of springs in northern part of district, seepage on low ground, small overflows at numerous points, and water passing off in form of steam from holes and crevices not measured but believed very nearly, if not quite, equal to total given above. In further addition, estimated surface flow in district south of area mapped (Low Terrace of present report), 50 gpm.

I have been acquainted with the Steamboat Springs district for about 24 years and can say of my own knowledge that the volume of water flowing at the different points is quite variable, probably due to caving in the fissures near the surface and to sealing up of the smaller surface openings by mineral deposition, or possibly to the opening up of underground passages which permit the flow of water to lower outlets, or into the gravel beds in the lower valley floor.

While engaged on stream gaging work for the U.S. Geological Survey some 15 years ago, and later when in charge of the work of the U.S. Reclamation Service in Nevada, I have seen flowing from the springs on the terrace shown on the topographic map (fig. 8), a volume of water which I have at various times estimated at from two to three cubic feet per second (900 to 1,340 gpm). In February last, in one of the open fissures, near the north end of the sinter terrace, which is now almost dry, a stream of water was flowing with a surface velocity of over one foot per second, as measured by floats, with a sectional area of more than two square feet. All of this water disappeared in the closed part of the fissure further north, and no part of it overflowed at this point.

[This observation is interesting but of doubtful value; tracer experiments have shown that convection occurs in some open fissures, related to sites of vigorous subsurface boiling with no net discharge, thus duplicating conditions recorded here by Taylor.]

Some years ago at Steamboat Springs station [Steamboat Resort of pl. 3], a shaft was sunk to a depth of 70 feet by Col. J. W. Hopkins and associates. [This is probably the same shaft, after further deepening, that was observed by Lindgren in 1901 (table 1); it is now used for hot vapor baths in the Steamboat Resort; its depth in 1946 was 46 ft below the floor of the building]. To unwater this shaft, a pump with rated capacity of 300 gallons per minute, operated by a 12 horsepower gasoline motor, was used. The capacity of the pump was not sufficient to take care of the water so as to permit sinking below 70 or 72 feet. The work of sinking this shaft continued over several months, but during that time the flow of water on the terrace to the north, covered by the topographic map, was not noticeably affected.

In view of the foregoing, I have no hesitancy in expressing the opinion that at least two, and possibly even more than three, cubic feet of water per second, equivalent to 1,300,000 to approximately 2,000,000 gallons per day, can be developed in the district.
* * * By far the greater portion of the surface flow of water appears on the terrace (Main Terrace) between the 4,650 and 4,675 contours, and is evidently flowing in a northerly direction in the rock crevices leading from the higher ground to the southward.

[The many discharging springs and points of visible water in the north half of figure 8 contrast strongly with the situation during 1945 and later years, when water was visible only at times in vent 21n and springs 53, 53c, and 53w (tables 4 and 5).]

TABLE 4.—*Recorded springs and geysers, Steamboat Springs, Nev., listed by area and altitude*

[Locations shown on pl. 3 unless noted]

Spring	Altitude of vent (feet)	Maximum recorded temperature (° C)	Location and remarks
			Main Terrace
41	4,676.8	Boiling	Crest of terrace. Geyser, active April to June, 1949, from new vent. Erupted maximum 4 ft above ground, negligible discharge, most returned to vent.
40	4,676.6	Boiling	Crest of terrace. Geyser, active at intervals 1947, 1950, 1952. Maximum height about 8 ft above ground; slight external discharge.
24sw	4,675.5	Boiling	Near Rodeo well. Geyser, small, observed erupting Feb. 1947 and June 1953; little, if any, other activity.
24w	4,673	Boiling	Northeast of Rodeo well. Geyser, active at intervals 1947-48, 1951. Maximum height 10-12 ft in February, 1951.
36ne	4,671	Boiling	West of drill hole GS-5. Geyser, Small, active 1951-53.
38	4,667.9	Boiling	North-central part of terrace. Geyser, occasional eruption 1945-48, 1951-52, to maximum height of 3 ft.
22 [1]	4,666.8	93½	North of Nevada Thermal Power Co. well 2. Spring, discharged at intervals 1945-49, 1952.
24 [1]	4,666.3	96½	Northeast of Rodeo well. Spring, discharged at intervals 1945, 1951-52; related to 23, 23n. Noted for high content of heavy metals in sediment.
24	4,665.8	95½	East of 24. Spring, discharged as much as 5 gpm early in 1952; discharge included with 24.
19nw	4,664	55	Northwestern part of terrace. Spring, observed discharging ¼ gpm, June 1953.
23 [1]	4,663.4	95½	Northeast of Nevada Thermal Power Co. well 2. Pulsating spring, discharged at intervals 1945-46, 1948-49; related to 24, 23n.
23n [1]	4,662.7	Boiling	North of 23. Small geyser, studied in detail during intervals of activity 1945-50; related to 23, 18.
5 [1]	4,662.3	95½	Near south end of terrace. Spring, intervals of discharge 1946-52; related to 2, 2nw.
19 [1]	4,661.2	88	North-central part of terrace. Spring, intervals of discharge 1945-46, 1951-52; related to 19n.
20 [1]	4,660.0	94	North-central part of terrace. Spring, continuous discharge except for winter 1947-48; related to 20n.
2nw	4,660	90	Southern end of terrace. Spring, active winter 1947-48; related to 2, 5.
19n	4,659.9	81	North-central part of terrace. Spring, continuous discharge 1946-51; related to 19.
			Main Terrace—Continued
15w	4,659.7	93	East-central edge of terrace. Pulsating spring, discharged winter 1950-51 as much as 2 gpm.
2 [1]	4,659.0	90	Southern end of terrace. Spring, continuous discharge; related to 2nw, 5.
15sw	4,658.7	Boiling	East-central edge of terrace. Geyser, small, active summer and fall, 1950; related to 15s.
17 [1]	4,658.5	95½	Northeast of Nevada Thermal Power well 2. Spring, pulsating at times, discharged at intervals 1945-46, 1948-51; related to 27.
20n [1]	4,658.2	69½	Northern part of terrace. Spring, continuous discharge 1948 to early 1950; related to 20.
18 [1]	4,657.7	96	Northeast of Nevada Thermal Power Co. well 2. Spring, continuous discharge except for brief intervals 1948, 1949, 1952; related to 23, 23n.
14w	4,657.3	Boiling	East-central edge of terrace. Geyser, small, active, spring and fall, 1950.
15s	4,657	89½	East-central edge of terrace. Pulsating spring, active November 1950 to September 1951; related to 15sw.
12sw	4,656.1	77	East-central edge of terrace. Spring, discharged as much as ½ gpm when active 1950-52.
34n	4,656	Boiling	Northern part of terrace. Geyser, small, splashing above ground, summer of 1950; a subsurface geyser at some other times; related to spring 34.
34 [1]	4,655.7	90	South of 34n. Spring, pulsating discharge 1946, related to intermittent vigorous boiling in same fissure 6 ft north; related to 34n geyser.
21s [1]	4,655.5	73½	Northern part of terrace. Spring, active 1950-51; related to 21.
16 [1]	4,654.9	95	East-central edge of terrace. Pulsating spring or continuous spouter, active 1945; related to 16se and others to the south.
16se [1]	4,654.9	95½	East-central edge of terrace. Pulsating spring, active 1950-52; related to 16.
13nw	4,654.5	90	East-central edge of terrace. Spring, active June 1945 and 1950-52, discharge as much as ½ gpm; related to 13.
14	4,654.4	93½	East-central edge of terrace. Spring with seeping discharge June 1945 and October 1949 through 1952. Near normal composition when discharging but slightly to strongly acid when not; pH and chloride content decreased with receding water levels in 13, 16.
3 [1]	4,654.4	78	Southern part of terrace. Spring, continuous discharge.
42	4,654.2	95	East-central edge of terrace. Spring, intervals of seeping discharge 1950-52; related to 42w, a nondischarging vent.
11	4,653.9	85½	East-central edge of terrace. Spring, intervals of discharge 1945, 1951-52, as much as ½ gpm.
27 [1]	4,653.4	95	North-central part of terrace. Spring, intervals of discharge 1945-47, 1950; related to 17.
13 [1]	4,653.3	Boiling	East-central edge of terrace. Geyser, intervals of activity 1945 and 1950-52; maximum height of eruption, 12 ft; related to 12, 14.
21 [1]	4,653.1	91	Northern part of terrace. Spring, active 1945-50; related to 21s.
10	4,653.0	79	East-central edge of terrace. "Low temperature" geyser, gassy. Slight discharge 1946-47, 1949-52; maximum rate about 5 gpm during eruption December 1951.
12 [1]	4,652.6	Boiling	East-central edge of terrace. Geyser, active December 1949 to January 1951, erupting to maximum height 25 ft; related to 13.
52	4,652	51	Spring, southeast of spring 2 and south of area of pl. 3. Intervals of seeping discharge 1945-48 as much as ½ gpm.
28 [1]	4,651.5	95	North-central part of terrace. Spring, intervals of discharge 1945-46, 1948-49.
8 [1]	4,647.8	93	East-central edge of terrace. Spring with continuous small discharge 1945 to September 1952 when discharge ceased. Deposited stibnite needles at surface.
4	4,639.1	66	Southeastern part of terrace. Gassy spring with intervals of discharge as much as 1 gpm, generally during winters.
43	4,635.6	80	Near east base of terrace east of drill hole GS-4. Spring with intervals of seeping discharge, 1945-47.
6	4,634.0	90	Near east base of terrace. Spring with intervals of discharge as much as ¼ gpm, generally in winter.
51	4,583	75	East-central base of terrace. Not a natural spring; discharge from old adit, about 4 gpm, 1945-52.
53	4,567.9	90	North of terrace (see pl. 3) and 300 ft southeast of East Reno well. Pulsating discharge 1945-52, only when East Reno well not erupting; maximum discharge about 2 gpm.
53w	4,565	Boiling	200 ft west-southwest of 53. Geyser, small, active from 2 vents January to April 1948.
53e	4,559	88	250 ft east-northeast of 53. Spring, brief intervals of discharge when East Reno well not erupting; maximum discharge about 2 gpm.

[1] Springs measured when discharging and included in computations for pl. 4.

TABLE 4.—*Recorded springs and geysers, Steamboat Springs, Nev., listed by area and altitude*—Continued

Spring	Altitude of vent (feet)	Maximum recorded temperature (° C)	Location and remarks
			Low Terrace [Located on pl. 3 unless noted]
1	4,628.7	Boiling	Northwestern part of terrace on old road to Main Terrace; perhaps more related to Main Terrace. Geyser and pulsating spring, discharged at short intervals 1945–46, 1948, at rates less than 1 gpm.
30s	4,624	Boiling	East-central part of terrace. Geyser, small, splashing above surface at intervals 1945, 1947–48; related to No. 32 Geyser well.
32	4,623.3	Boiling	East-central part of terrace. Abandoned well erupting as major geyser as much as 70 ft; active 1945–52 commonly near 1 week intervals; eruptions influence all nearby vent.
30n	4,623.1	Boiling	Northeast of 32. Geyser and pulsating spring, active 1945; discharged as much as 1 gpm; affected by 32.
31s	4,623	Boiling	Northeast of 32. Geyser, small, rare eruptions 1950 related to eruptions of 32.
32se	4,623	Boiling	Southeast of 32. Geyser, small, active at intervals 1947, 1950–52; eruptions related to water level of 32.
32e	4,623	Boiling	East of 32. Geyser, small, rare eruptions 1950 related to eruptions of 32.
31n	4,622.4	Boiling	Northeast of 32. Geyser, with only known eruption June 1946; discharged as much as 1 gpm in June 1950; water level very responsive to 32.
25ss	4,613.3	95½	Northern part of terrace. Spring, active 1949, discharged as much as ½ gpm; related to 25.
25[1]	4,612.7	96	Northern part of terrace. Pulsating spring, discharged 1945–50, 1953, except for brief intervals; related to nearby springs, especially 26.
25s[1]	4,612.7	96	South of 25. Pulsating spring, active 1950–52; related to 25.
26[1]	4,612.7	95	North of and related to 25. Geyser and pulsating spring, small, active 1945–46, 1948.
26nw	4,612.7	86½	Northwest of 25. Spring, with intervals of discharge 1947–50 as much as 1 gpm.
44	4,606.6	51	Northern part of terrace. Spring, continuous discharge 1945–52 as much as 1 gpm.
54	4,600.8	45	Northeastern part of terrace. Spring, continuous discharge 1945–52 as much as 1 gpm.
33	4,594.7	60	Southeastern part of terrace. Spring, almost continuous discharge 1945–52. Measured but discharge strongly affected by Steamboat wells.
50[1]	4,594.4	60	Northeast edge of pl. 3. Spring, continuous discharge 1945–52.
50n	4,594.2	50±	Northeast edge of pl. 3. Seepage, no main vent, continuous discharge 1945–52 as much as 1 gpm.
44e	4,590	50±	Northeast edge of pl. 3. Seepage area, continuous discharge 1945–52 as much as 1 gpm.
44ne	4,590	50±	Do.
47	4,584	75	On creek east of terrace (see pl. 1). Continuous discharge, total from several vents about 8 gpm but submerged by creek during high stages.
46	4,575.9	77	On creek northeast of terrace. Spring, continuous discharge as much as 2 gpm 1945–52 but submerged by creek during high stages.
45	4,575	71	On creek northeast of terrace. Springs with continuous discharge as much as 3 gpm but visible only during lowest stages of creek.

[1] Springs measured when discharging and included in computations for pl. 4.

PRESENT ACTIVITY OF NATURAL VENTS

The term "activity," as used in this report, includes (1) discharge of water at the surface, from hot springs and geysers; (2) discharge of other gases at the surface, with or without liquid water; and (3) all subsurface circulation of fluids, some of which is thermal water that flows below the surface and escapes unseen, directly into Steamboat Creek.

The locations of springs are determined by many factors. Normal springs of continuous discharge occur where the water table intersects the topographic surface. If the water table is not too far below the surface, thermal water can be discharged in perpetual spouting springs and geysers. Many vents discharge only steam (or water vapor) and other gases from a water table that is below the ground surface.

The estimated position of the water table is shown on all sections of plate 2; most of the basic data on water levels are shown in tables 4 and 5.

DISTRIBUTION OF SPRINGS AND FUMAROLES

All springs known to have discharged from June 1945 through the spring of 1953 are listed in table 4, according to principal area and altitude. Data for the systematically measured springs, with certain exceptions mentioned below, are shown on plate 4.

Most springs of the Main Terrace are near the crest of the terrace on the eastern flank. The altitude of nearly all these springs of plate 4 is within the range of 4,634 to 4,676 feet. The springs of highest discharge lie between 4,651 and 4,667 feet, in close agreement with Taylor's observations. Of the 46 springs of the Main Terrace, 13 have erupted as geysers (White, 1967b), and 6 have been pulsating springs. Only three springs (Nos. 2, 3, and 8) discharged without interruption from June 1945 to August 1952 (pl. 4), and one of these (No. 8) ceased flowing in the fall of 1952. All springs with continuous or nearly continuous discharge are relatively low in altitude.

The springs that are highest in altitude are also high in temperature. Although some at lower altitudes are near boiling, most are considerably below boiling. The factors that influence temperature at point of discharge are discussed on pages C87–C89.

The Low Terrace (pl. 3) is a subsystem distinct from but similar to the Main Terrace. Most springs discharge at altitudes between 4,612 and 4,624 feet (table 4). Of 20 springs on the Low Terrace, 9 have erupted as geysers and 2 have been pulsating springs without geyser action; 6 discharged continuously during the observation period, but, of these, only spring 50 was measured systematically and is included in the computations for plate 4. The other five springs (44, 54, 50n, 44e, and 44ne) are all small and have discharges of about 1 gpm or less.

The high-altitude springs of the Low Terrace are the hottest, and the low-altitude springs on the east and north border of the terrace are much below boiling.

Table 4 also includes three springs (45, 46, and 47, pl. 1) on Steamboat Creek east and north of the Low Terrace. These springs are probably indicative of many unseen small springs that discharge directly into Steamboat Creek. Measurements of total discharge and chloride content of Steamboat Creek have revealed that much chloride-bearing water is flowing unseen directly into the creek. About 100 gpm (gallons per minute) of this unseen thermal water, assumed in the calculations

TABLE 5.—*Altitudes of some wells, natural vents, and water levels in the Steamboat thermal area*

[Measurements in feet]

Well or spring	Location	Ground altitude	Range in recorded depths to water level	Best value for depth to water table	Approximate altitude of water table
GS–1	Low Terrace	4,622.9	[1] −6.0–+52	−5	4,618
2	High Terrace	4,720.8	−37.1–−51.7	−41	4,680
3	Main Terrace	4,676.3	−6.3–−14.4	−8	4,668
4	do	4,664.7	[1] −3±–+65	−3±	4,662
5	do	4,661	−6.4–+60	−5	4,656
6	Sinter Hill	4,838	−143–−154	−148	4,690
7	Silica pit	5,027	−106–−121	−112	4,915
8	Main Terrace	4,607.2	−15.1–−30.0	−30	4,577
Auger hole:					
4	East of Main Terrace	4,572	−6.5	−6.5	4,565
5	do	4,572	−7.8	−7.8	4,564
6	do	4,676	−9.4	−9.4	4,667
12	West of Main Terrace	4,680.6	−15.4	−15.4	4,665
13	do	4,677.1	−10.2	−10.2	4,667
South Steamboat well	Low Terrace	4,611.1	−12.4–+2.1	−7	4,604
Steamboat well 4	do	4,612.8	−6.7–−24	−13.5	4,599
No. 32 Geyser well	do	4,623.4	[2] −0–−3	.5	4,623
Murray	Northeast of Main Terrace	4,533	−12.9	----------	4,520
Harold Herz 1	North of Main Terrace	4,606.4	−53.6–−60.2	−56	4,550
Harold Herz 2	do	4,598.0	−55.1–−55.6	−55	4,543
Mount Rose	do	4,595.1	−52.8–−56.5	−56	4,539
East Reno	do	4,588.2	−6.1–−9.3	−6	4,582
West Reno	do	4,607.5	−26.2	−26	4,581
Senges	Northwest of Main Terrace	4,649.6	[3] −11–−38.4	−11	4,639
Murray well	Pine Basin	4,787.0	[3] −70?–−112	−70?	4,717?
Rodeo	Main Terrace	4,676.7	−6.7–−15.6	−8.6	4,668
Nevada Thermal Power Co.:					
4	West of Pine Basin	[4] 4,870	−217	[5] −95	~4,775
5	do	[4] 4,875	−234–−282	[5] −120	~4,775
6	Pine Basin	[4] 4,790	−95–−115	[5] −86	~4,705
Vent 35	Main Terrace	4,676.8	−7.5–−10.3	−8.9	4,667.9
Vent 36	do	4,675.5	−6.7–−8.3	−7.5	4,668.0
Spring:					
22	do	4,666.8	0 when flowing	near 0	4,666
21sw	do	4,667.3	−6.0	−6	4,661
21	do	4,653.1	−1–0	near 0	4,653
16	do	4,654.9	−2–0	−1	4,654
8	do	4,647.8	0	0	4,648
21n	do	4,648.8	−10.5–−12+	−12	4,637
6	do	4,634.0	−1–0	−1	4,633
25	Low Terrace	4,612.7	0	0	4,613
33	do	4,594.7	0	0	4,595
50	do	4,594.4	0	0	4,594
53	do	4,567.9	−2–0	−1	4,567

[1] Positive wellhead vapor pressure highest during or immediately after eruption, expressed as feet of water.
[2] Immediately after an eruption, water level is much lower than indicated here.
[3] From drillers' old reports.
[4] Estimated.
[5] See text.

to contain 820 ppm Cl, is discharged opposite the Low Terrace, and about 150 gpm opposite the Main Terrace.

Hot springs do not discharge in the western part of the thermal area, even where favorable structures exist, because the water table is too far below the surface (fig. 5). Sinter Hill and the High Terrace were very active sites of discharge in the past when the water table was much higher (White and others, 1964), but present activity visible at the surface consists only of a few feeble fumaroles and other local "hot spots" related to rising gases (silica pit, clay quarry, and other places).

FACTORS CONTROLLING DISCHARGE

Where the water table intersects the ground surface, geologic structures and permeabilities determine to major extents the rates and temperatures of discharge. Where the general water table is below the surface, discharge can occur from a vent that is either a geyser or a perpetual spouter.

In parts of the thermal area, the water table is so close to the ground surface that many factors determine whether discharge occurs, and if it does, the rate of discharge. Stated in another way, the position of the water table at any single location and time is a resultant of many competing factors; a change in any factor changes the position of the water table; if the water table is at or above the ground surface, changes in discharge occur.

The factors to be considered in this section include (1) precipitation, (2) barometric pressure, (3) earth tides, (4) earthquakes, (5) discharge from geothermal wells, and (6) other random changes. Major factors that control long-term changes involving times of many years are implied but not considered in detail. These include major climatic change; erosion and sedimentation of Steamboat Creek, which determines base-level of the water table of the surrounding area; changes in the magmatic hearth that supplies the heat of the system; structural events, vein filling, and chemical solution that create new channels or change permeabilities in old channels.

Of the six factors that cause short-term fluctuations, the most important is precipitation, on all scales from single storms through seasonal, annual, and longer term changes. Many readers may be surprised to learn, however, that changes in barometric pressure are more important than precipitation in determining most day-to-day changes in discharge and water level. The water table is a delicately balanced system that is highly responsive, especially near the crests of the Low and Main Terraces, to changes in any controlling factor. Thus, when the pressure of the atmosphere on the water surface increases, the water level declines, and vice versa. The water table, then, acts as an inverted barometer. The liquid in one type of familiar barometer is mercury, so that we normally express barometric pressure in terms of inches or millimeters of mercury (1 atm of pressure =29.92 in., or 760 mm of Hg). Air pressure can also be stated in terms of water (1 atm=33.899 ft water at 4°C).

The effects of barometric pressure are considered first because the day-to-day changes are prominent, effects are almost immediate, and effects are more nearly predictable quantitatively than for any other factor. Records of water levels and spring discharges can then be corrected to an assumed constant barometric pressure and examined for more obscure or less predictable influences. These in particular include earth-tidal influences and lag effects of precipitation.

INFLUENCE OF BAROMETRIC PRESSURE

NONFLOWING VENTS

Changes in barometric pressure have very pronounced effects on water levels in nonflowing vents in the high southwestern part of the Main Terrace (pl. 3), where no discharge of water occurs other than the eruption of an occasional small geyser. Northward along the main fissure system, altitudes of the terrace crest decrease until the water table intersects the ground surface and springs can discharge. As might be expected, a flowing spring can show a major change in its rate of discharge in response to a barometric change, but actual change in altitude of the water surface of the spring is slight. Thus, a discharging spring greatly lessens the changes

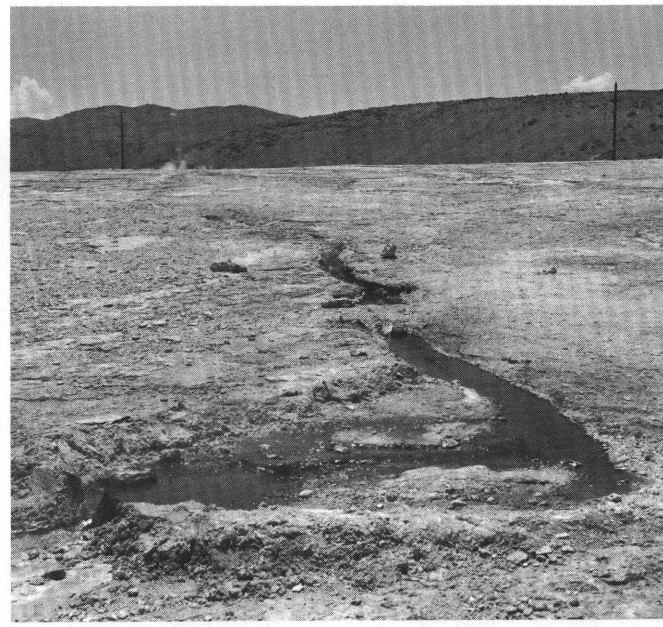


FIGURE 9.—Spring 24 near crest of Main Terrace, discharging water dark with suspended silica and metal sulfides. This spring is also very responsive to changes in barometric pressure.

that might otherwise occur in the water levels of nearby nonflowing vents. For example, spring 24 (pl. 3 and fig. 9) rose only 0.1 foot in altitude from September 14 to 21, 1945, while its discharge was increasing more than 600 percent. At the same time, the water level in vent 35, which is 1,100 feet to the south, rose 1.4 feet, and in vent 36, which is 400 feet to the south of spring 24, the water level rose 1.1 feet (table 6; pl. 4, curves *E*, *F*). In contrast, the rise in vent 37, about 100 feet southeast of spring 24, was only 0.15 foot, or only slightly more than in the spring.

In the series of measurements shown in table 6, barometric pressures in inches of mercury are also shown as equivalent feet of cold water. If a hydrologic system behaves as a perfect but inverted water barometer, a decrease in air pressure equivalent to 1 foot of water should result in a rise of 1 foot in water level. The barometric efficiency (expressed in percent) is the change in altitude of a water level divided by the change in barometric pressure as expressed in linear units of water. Inches of mercury may be converted to inches of water (assumed temperature 4°C and relative density 1.000) by multiplying by 13.546. A correction for density differences related to thermal expansion in the system may also be applied, if known. The water involved in such changes at Steamboat Springs has a relative density close to 0.96 at 100°C and 0.90 at 170°C, neglecting minor salinity effects. Because of many uncertainties in the density differences of all water actually involved, a

TABLE 6.—*Barometric influence on nonflowing vents of the Main Terrace, September 1945, in order of decreasing altitude of water level*

Measurement	Sept. 14	Sept. 21	Sept. 28	Average Sept. 14 and 28	Difference Sept. 21 from average of Sept. 14 and 28
	Barometric pressure				
Hg inches	25.53	25.08	25.66	25.60	−0.52
H_2O feet [1]	28.93	28.41	29.07	29.00	−.59

Vent	Altitude (feet) reference point	Depth, in feet, below reference points				Average altitude of water level Sept. 14 and 28 (feet)	Difference Sept. 21 from average of Sept. 14 and 28 (feet)
		Sept. 14	Sept. 21	Sept. 28	Average Sept. 14 and 28		
36	4,675.1	−7.47	−6.36	−6.70	−7.09	4,668.0	+0.73
35	4,676.2	−8.88	−7.49	−7.92	−8.40	4,667.8	+.91
37	4,667.6	−1.20	−1.05	−1.34	−1.27	4,666.3	+.22
5	4,664.0	−3.08	−2.31	−2.27	−2.68	4,661.3	+.37
15	4,655.2	−.11	−.09	−.06	−.09	4,655.1	.00
34	4,657.6	−3.01	−2.96	−3.04	−3.02	4,654.6	+.06
14	4,654.4	.00	−.02	+.01	.00	4,654.4	−.02
42w	4,655.4	−1.92	−1.86	−1.95	−1.93	4,653.5	+.07
9	4,653.0	−.87	−.75	−.78	−.82	4,652.2	+.07
12	4,652.7	−1.79	−1.76	−1.97	−1.88	4,650.8	+.12
43	4,636.0	−.75	−.49	−.42	−.58	4,635.4	+.09
7	4,634.4	−1.00	−.94	−.43	−.72	4,633.7	−.22

[1] Water assumed to have sp gr of 1.00. If average temperature at depth in the system is 170°C and average density is 0.90, these figures should be increased by 11 percent.

water density of 1.000 is assumed in calculating barometric efficiencies at Steamboat Springs. Actual efficiencies, because of thermal expansion, are probably 5–10 percent lower than these calculated values.

From the data on barometric pressure in table 6, vents behaving as inverted water barometers with 100 percent efficiency should show water-level changes of +0.59 foot on September 21 as compared to the average levels of September 14 and 28. However, we see that vents 35 and 36 have responded with efficiencies considerably greater than 100 percent, even if density of water in the system is corrected 5–10 percent for thermal expansion. Calculated responses that exceed 110 percent clearly require some other explanation, probably involving boiling and changes in proportion of vapor to liquid phases in the upper part of the spring system.

A water-stage recorder was installed and maintained at vent 35 from May 1946 to early fall of 1952, except for brief periods of breakdown caused by the highly corrosive environment. Curve *A* of figure 10 was traced and reduced from the original records from vent 35 from June 17 to July 16, 1946. A recording microbarograph was also maintained at the Steamboat Resort, half a mile south of vent 35 and 70 feet lower in altitude.

The barometric efficiency of a vent should be calculated over a period of time by comparing barometric changes with corresponding changes in water level. If the barometric change were the sole influence, the change in water level would always be exactly proportional to the barometric change but opposite in sign. Many other influences do exist that tend either to augment or offset the barometric influence. Because of interfering effects, the calculated barometric efficiency differs somewhat, depending on the particular method and time span used in its calculation. The efficiency of the water level of vent 35 for the interval of time shown in figure 10 was calculated by two different methods. In one, only major barometric changes of at least 0.10 inch of mercury were considered. The total of such major barometric changes for the month was 2.67 inches of mercury (equivalent to 3.01 ft of water of density 1.000); the corresponding changes in water level (all opposite in sign to each barometric change) totaled 2.18 feet of water; the calculated barometric efficiency, 2.18/3.01, was 72 percent.

In a second method, the existing barometric pressure and water level were noted at a specified time each day. Noontime changes for the period of June 17 to July 16, 1946, totaled 1.48 inches of mercury (or 1.67 ft of water of density 1.000), and the total net change in water level was 1.42 feet of water; the total barometric changes were actually accompanied by 1.47 feet of change in water level opposite in sign to each barometric change, and 0.05 foot of change of the same sign, giving a net change of 1.42 feet of water that can be attributed to barometric influence. The calculated barometric efficiency for the month was 85 percent. This method indicates an efficiency slightly higher than the first and is preferred for short-time responses. Similar calculations were made by still other methods with slightly different efficiencies indicated for each.

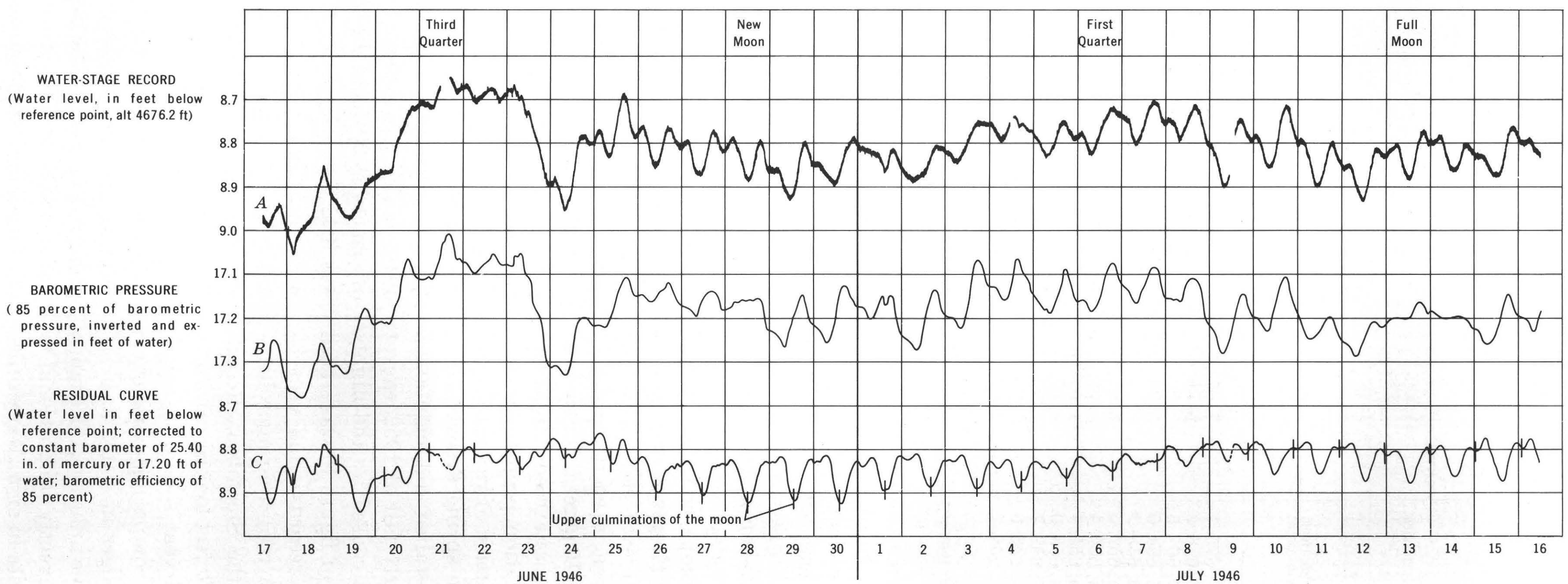


FIGURE 10.—Fluctuations of water level, vent 35, Main Terrace, showing influence of barometric pressure and earth tides.

The barometric efficiency of vent 35 was calculated for 7 individual months from June to December 1946, by considering daily noontime changes. The indicated efficiency ranged from 40 percent in November to 118 percent in September; these differences were caused by other uncompensated changes. The average for the 7 months was 74 percent.

In figure 10 curve *B* is the barometric record, inverted and computed to feet of water (assumed density 1.000, with efficiency of 85 percent). The general similarities between curves *A* and *B* are obvious, but significant differences are also evident. Curve *A* was then converted to an assumed constant barometric pressure by substracting the fluctuations of *B*; curve *C* shows the residual fluctuations.

This correction for barometric influence is similar to that used by Robinson (1939). Robinson found that residual curves he obtained from a well in New Mexico and another in Iowa indicated earth tides. The same conclusion is evident here, as described under "Influence of Earth Tides" (p. C42–C44).

Records from vent 31n on the Low Terrace were analyzed in the same manner and showed a similar relationship between barometric pressure and water stage (fig. 11, curves *A*, *B*). Vent 31n ordinarily does not discharge, but it has erupted at least once as a small geyser (table 4). The vent is on a fissure adjacent to vent 32, an abandoned well that erupted occasionally as a major geyser and hereafter is called No. 32 Geyser well. An average barometric efficiency of about 60 percent was found by the daily-change method for the full period of November 25 to December 25, 1946; but a 75-percent correction is evidently somewhat more accurate for the short-term fluctuations of November 15 through 17, judging from the comparisons shown in subsidiary curves adjacent to *b* and *c*. The indicated effect of precipitation and other factors are considered elsewhere. Several fluctuations in water level occurred early in the morning of December 4, 1946, that were only partly compensated for by the barometric correction calculated at 60 percent efficiency over the period. The residual curve *C* shows fluctuations about as large as the applied corrections, suggesting that efficiencies for very short-term barometric changes are about 120 percent (with assumed water density of 1.000); even if an average temperature of 170°C is assumed for the water involved in the Low Terrace fluctuations, with an average density of 0.90, a short-term barometric efficiency of nearly 110 percent is indicated.

A surprising feature of the high barometric response of some vents is the fact that calculated efficiencies range from 60 to 85 percent for as long as 1 week of sustained high or low pressure, despite the fact that all vents are connected rather closely to flowing springs, which should dampen the response.

FLOWING SPRINGS

The curves for measured discharge and barometric pressure on plate 4 (curves *C*, *D*) generally show an inverse relation. In about 70 percent of the observation sets, discharge decreased with increased air pressure or vice versa. The 30 percent that were exceptions are with little doubt related to other masking influences.

A striking example of the effect of air pressure on discharge is found in the data of September 21, 1945, as compared to earlier and later observations (table 7; pl. 4). The record is particularly clear because these barometric changes were large and especially because precipitation was absent as a complicating factor.

The measured springs of table 7 are listed in order of decreasing altitude. Most springs were highest in discharge on September 21, when air pressure was lowest. The discharge of several springs was changing during this time interval for reasons other than air pressure.

Springs 22, 24, and 18 showed the greatest change, both in relative and absolute amounts, and account for most of the differences in total discharge between the sets of measurements. Springs 22 and 24 are the highest in altitude, but altitude is not the sole determining factor because springs 19, 2, and 17, with relatively small changes in discharge, are all higher in altitude than spring 18. Additional important factors are the near-surface dimensions of the vent and the subsurface channel characteristics of each spring. Spring 24, which shows the greatest fluctuations in discharge, is an open fissure 30 feet long and about 3–6 inches wide (fig. 9). The fissure is bridged by sinter near its middle part and narrows at both ends to a fracture sealed with sinter. Rate of discharge is determined by water level in the fissure; when the level is above the sill level at the north end, the spring discharges. Slight changes in water level in this broad fissure produce major changes in discharge.

Springs 22 and 18 have vents with minimum visible cross-sectional areas of about 6 square inches. In contrast, the vents of springs 19, 2, and 17 are small and range from 1 to 3 square inches in cross-sectional area at depths of greatest visible restriction. The channel characteristics of each spring are evidently very important in determining barometric response. Spring 18 is 220 feet north of spring 24 and is on the extension of the same fissure (pl. 3). The subsurface connections between these two springs must have tight restrictions to explain the existing difference of 8½ feet in altitude.

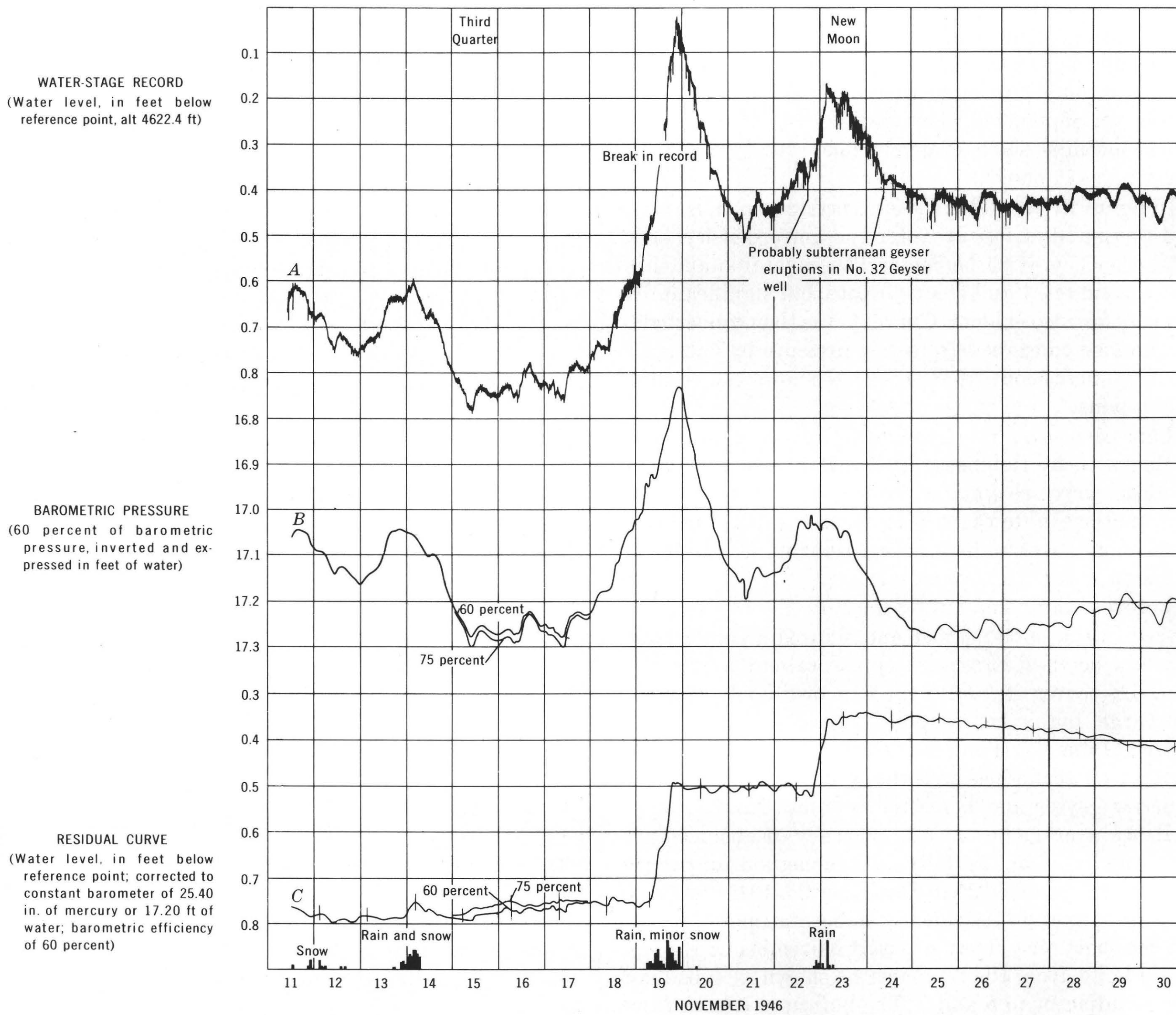
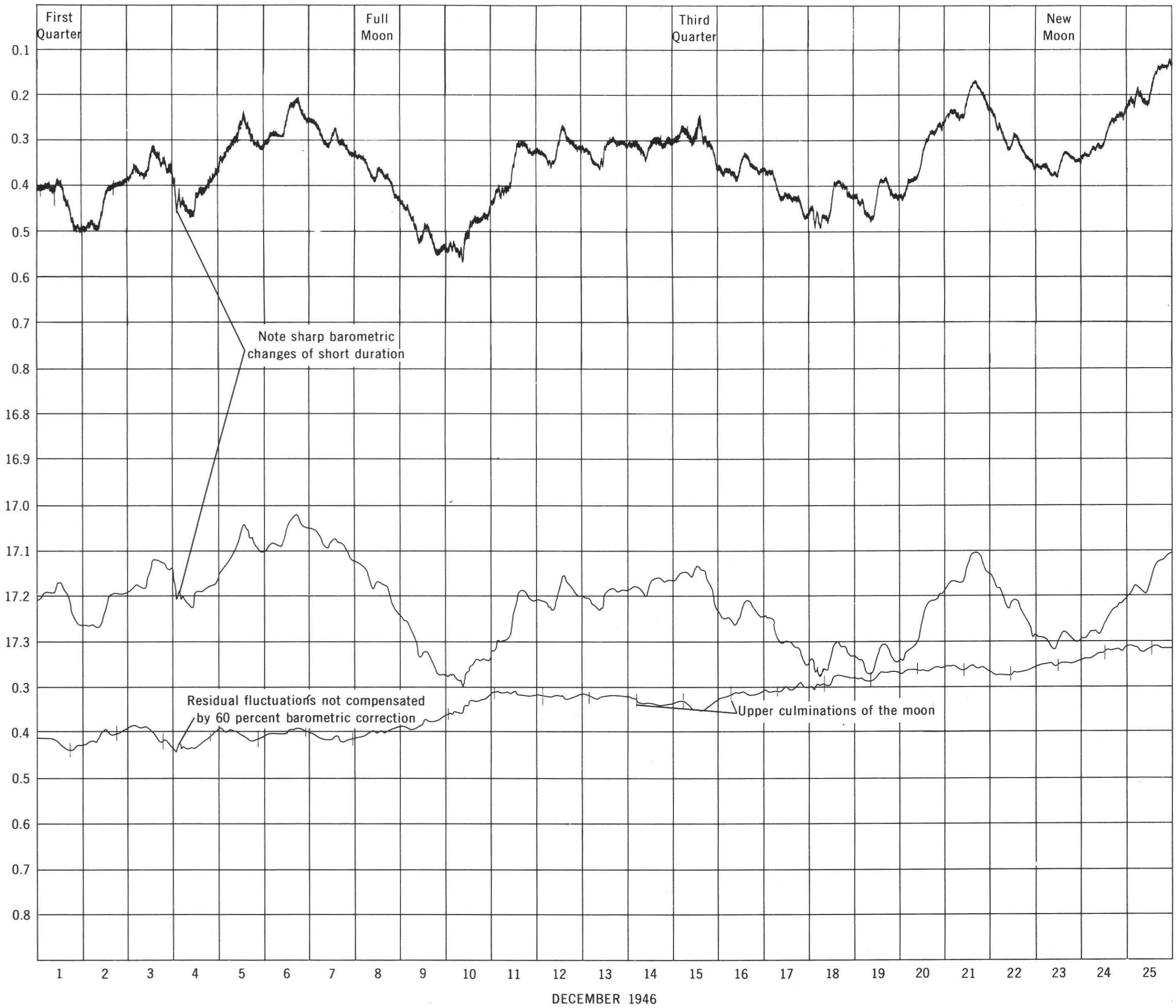


FIGURE 11.—Fluctuations of water level, vent 31n, Low Terrace, showing influences of barometric pressure and It is shown by the

Similar reasoning indicates that spring 16 and in fact all low-altitude springs of table 7 must also have very restricted interconnections with the open fissures near the crest of the Main Terrace. Spring 16 has a relatively large vent of about 1 square foot. The vent enlarges downward into a chamber of unknown dimensions just below the surface, but tight restrictions must exist at greater depth; its fissure is structurally related to others of the terrace, and all presumably interconnect at depth; if all fissures were wide and unrestricted, the system would discharge entirely from vents that are lowest in altitude.

Measurements of discharge were made daily from November 4 through December 7, 1946, and two to three times a week for the remainder of December (fig. 12). The principal purpose of this series was to obtain detailed data on influence of heavy precipitation, discussed in another section, but barometric responses are considered here. Changes in total discharge (fig. 12, curve C) that are related to barometric pressure (curve D) are very conspicuous, as for November 9. The behavior of four individual springs during this same interval is shown in figure 13 to illustrate the marked individuality of different springs.



precipitation. Precipitation was slight between November 25 and December 25 and is not shown on the graph. storm on plate 4.

A more detailed analysis is permitted from the data of figure 14, which shows the total measured discharge for each date plotted against barometric pressure. Changes that follow negative slopes of about 15°–30° are related primarily to barometric pressure, a decrease in pressure ordinarily causing an increase in discharge and vice versa. The fact that all points do not plot along a single line of negative slope is clear evidence for influences other than barometric pressure. Intervals of major precipitation are indicated on certain tie lines, and each of these caused an increase in discharge, with the exception of November 19 to 20. Also evident in this figure is a general trend toward increased discharge unrelated to barometric change from November 4 to 23, followed by a steady decrease through late November and most of December. These changes, at least in part cumulative are related to series of storms and subnormal precipitation.

The high-altitude springs with unrestricted vents, as seen from the data of table 7, are the most responsive to barometric changes; some of these springs discharge only when water levels in the Main Terrace are relatively high and total discharge is therefore high. As the general water table is raised for any nonbarometric

TABLE 7.—*Barometric influence on flowing springs, September 1945, listed in order of decreasing altitude*

Spring	Altitude (feet)	Sept. 14	Sept. 21	Sept. 28	Average Sept. 14 and 28	Percent difference Sept. 21 and average of Sept. 14 and 28
		Barometric pressure (inches of Hg)				
		25.52	25.08	25.66	25.59	
		Discharge (gpm)				
22	4,666.8	4.46	11.95	6.07	5.27	+127
24	4,666.3	2.04	14.51	.00	1.02	+1,320
19	4,661.2	2.20	2.34	2.02	2.11	+11
2	4,659.0	1.20	1.87	2.07	1.64	+14
17	4,658.5	8.17	8.45	7.38	7.78	+8.6
18	4,657.7	11.34	17.07	11.14	11.24	+52
16	4,654.9	1.99	[1] 1.08	1.65	1.82	−41
3	4,654.4	1.31	1.38	1.47	1.39	−.7
27	4,653.4	3.06	2.80	2.21	2.63	+6.5
21	4,653.1	6.51	6.51	6.75	6.63	−1.8
28	4,651.5	.64	.40	.00	.32	+25
8	4,647.8	.22	.27	.23	.23	+17
26	[2] 4,612.7	1.80	2.20	1.60	1.70	+29
25	[2] 4,612.7	.86	.64	.64	.75	−15
	Total [3]	45.80	71.44	43.23	44.52	+60

[1] Continuous spouter at the time, with much discharge lost as spray in very strong wind.
[2] Low Terrace.
[3] Excludes spring 50, not measured at this time.

reason such as precipitation, these responsive springs become active, and the reverse occurs when the general water table falls enough for these springs to cease flowing. This explains why the trends on the low-discharge side of figure 14 have negative slopes close to 15°, whereas trends on the high-discharge side (strongly affected by short-term responses to precipitation as well as barometric change) have negative slopes near 30°.

Curve *E* of figure 12 shows the assumed changes in discharge that might be expected if no barometric change occurred. To construct this curve, each point in figure 14 was projected, using reasonable slopes, to a constant barometric pressure of 25.40 inches of mercury. A few of the assumed slopes of projection are indicated in figure 14. The differences between curves *C* and *E* of figure 12 are a measure of the barometric response, and the residual fluctuations in curve *E* are largely a measure of direct influences of precipitation isolated from the decreases in barometric pressure that normally accompany precipitation.

BAROMETRIC INFLUENCE IN OTHER AREAS

Barometric influence has been detected in water-stage records from many artesian wells. Piper (1932) described a municipal well at The Dalles, Oreg., where the efficiency seems to be close to 100 percent. Robinson (1939) found an artesian well in New Mexico and another in Iowa that showed very strong barometric fluctuations. The well in New Mexico had an efficiency of 70 percent as a water barometer, assuming a density of 1.2 for the saline water of the system. Another well penetrating the same brine horizon, but in an area of saline springs, showed a dampened barometric response. The barometric efficiency of an artesian well in Iowa City was calculated by Robinson to be about 75 percent, assuming a water density of 1.0.

Although barometric effects have been detected most commonly in nonflowing artesian wells, minor influences have been found in a few water-table wells. Efficiencies of the order of 3 percent were found by Lugn and Wenzel (1938) where the soil above the water table was saturated with water or was frozen and served as a confining layer that restricted transmission of barometric changes to the water table. The barometric changes exerted on the water interface in a well evidently were not counterbalanced immediately by changes effective elsewhere through a permeable soil zone to the water table.

Meinzer (1939, p. 212–213) and Tolman (1937, p. 332) briefly mention barometric fluctuations, but they attempt no general theoretical treatment of the subject.

Nomitu and Seno (1939, p. 417–423) described barometric effects on the discharge of Beppu hot springs in Kyushu, Japan. The influence was strongest on individual springs that were farthest from the ocean and highest in altitude. Springs near shore were strongly influenced by loading and unloading of the aquifer by ocean tidal changes, and a barometric influence was not clearly distinguishable.

Peck (1960) considered water-table fluctuations related to barometric effects upon air (or other gases) entrapped below the water table.

Yuhara (1961, p. 297–302) described and analyzed the differing influences of barometric pressure, precipitation, and ocean tidal loading on the Atami spring system near the seashore on Izu Peninsula southwest of Tokyo, Japan. The tidal efficiency of the spring system relative to changing pressures from ocean tides on the sea floor is calculated to be 49.5 percent, with a phase lag of about 1 hour. The barometric response is negative, as at Steamboat Springs, and is about 50 percent.

EXPLANATION OF DIFFERING RESPONSES OF MAIN AND LOW TERRACES

The Main and Low Terraces at Steamboat Springs are two distinct subsystems, at least in their barometric response relative to altitude. The springs and vents at highest altitude on each terrace are most strongly affected, and those at lower altitudes are much less responsive. These relationships can be explained by assuming that the highest water levels of each terrace are close to the maximum potential of each subsystem. The springs that emerge at low altitudes through restrictive fissures

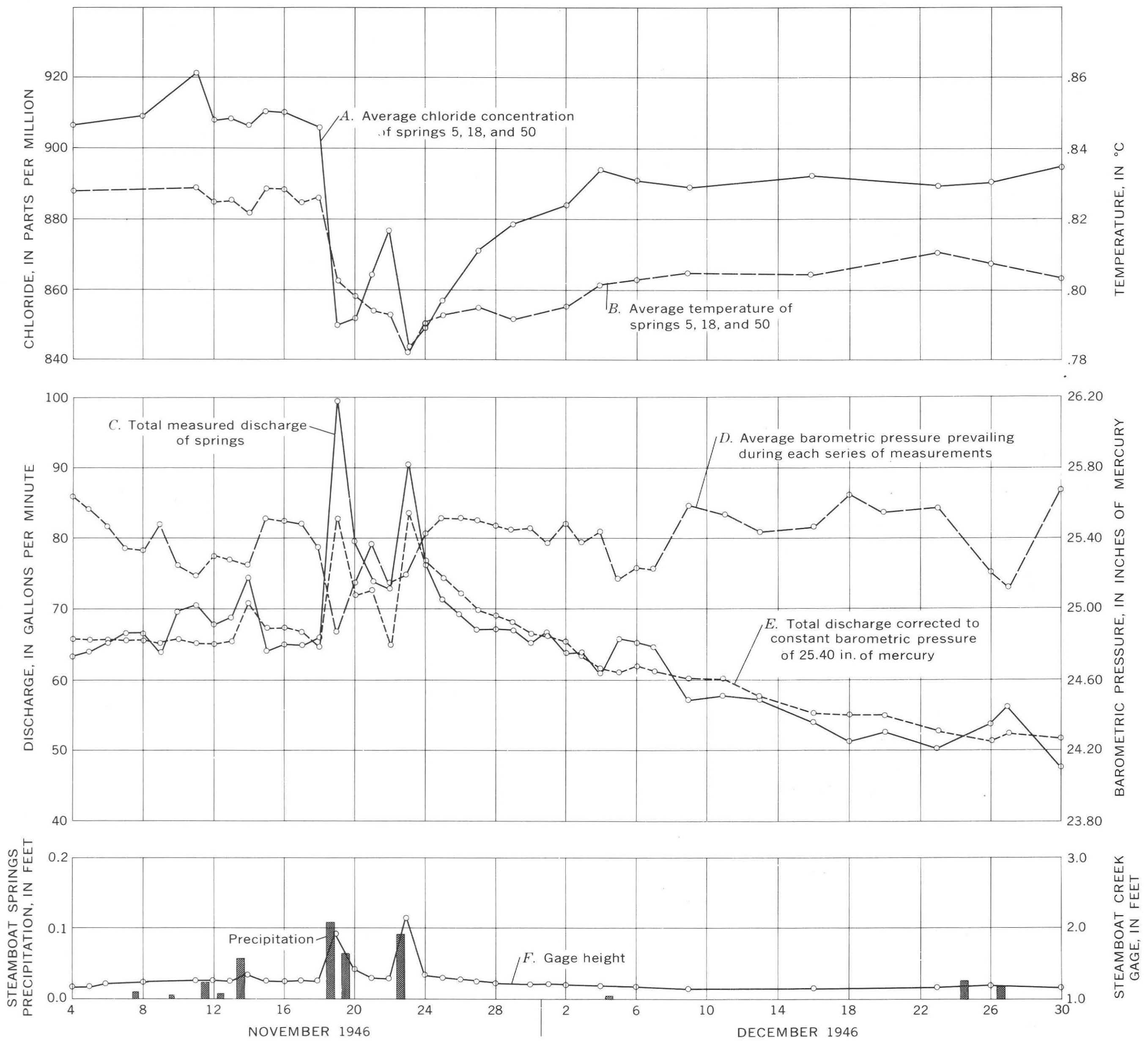


FIGURE 12.—Total measured discharge from hot springs with related factors, November and December 1946.

escape under hydrostatic pressures determined by the highest water levels of each subsystem. The hydrostatic pressure on water escaping from spring 24, for example, may usually be equivalent to about 2 feet of water, which is the difference in altitude between the spring outlet (4,666.3 ft) and the water levels in vents 35 and 36 (average close to 4,668.0 ft, from curves *E* and *F*, pl. 4). The indicated equivalent pressure on spring 8 (4,647.8 ft) is equivalent to 20 feet of water. A change in barometric pressure of 1 inch of mercury (or 13 in. of water) is about 50 percent of the hydrostatic pressure effective on spring 24, but it is only a little more than 5 percent of the pressure effective on spring 8. Spring 24 should be much more responsive than spring 8 to changes in barometric pressure, according to the above analysis.

We have seen (p. C23) that vent 31n at an altitude of 4,622.4 feet on the Low Terrace responds barometrically with an efficiency of 60–70 percent. Other nonflowing vents near the crest of the Low Terrace also showed high response to barometric change from weekly measurements, but suitable nonflowing vents low on the flank of this terrace are lacking for comparison. The data of table 6 indicate, however, that all vents on the Main Terrace below 4,660 feet in altitude had barometric efficiencies of less than 20 percent. The Low Terrace, therefore, is clearly a separate subsystem

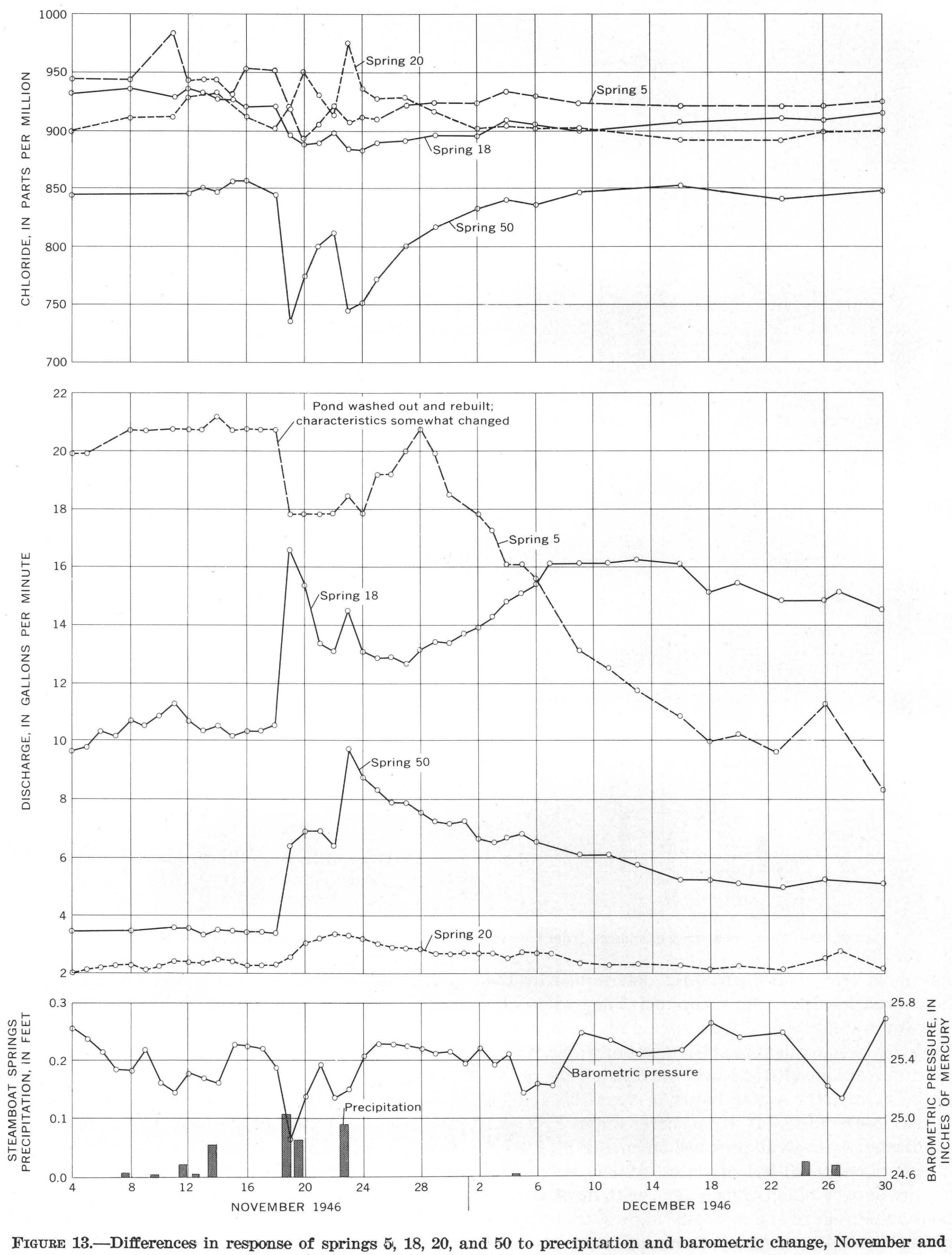


Figure 13.—Differences in response of springs 5, 18, 20, and 50 to precipitation and barometric change, November and December 1946

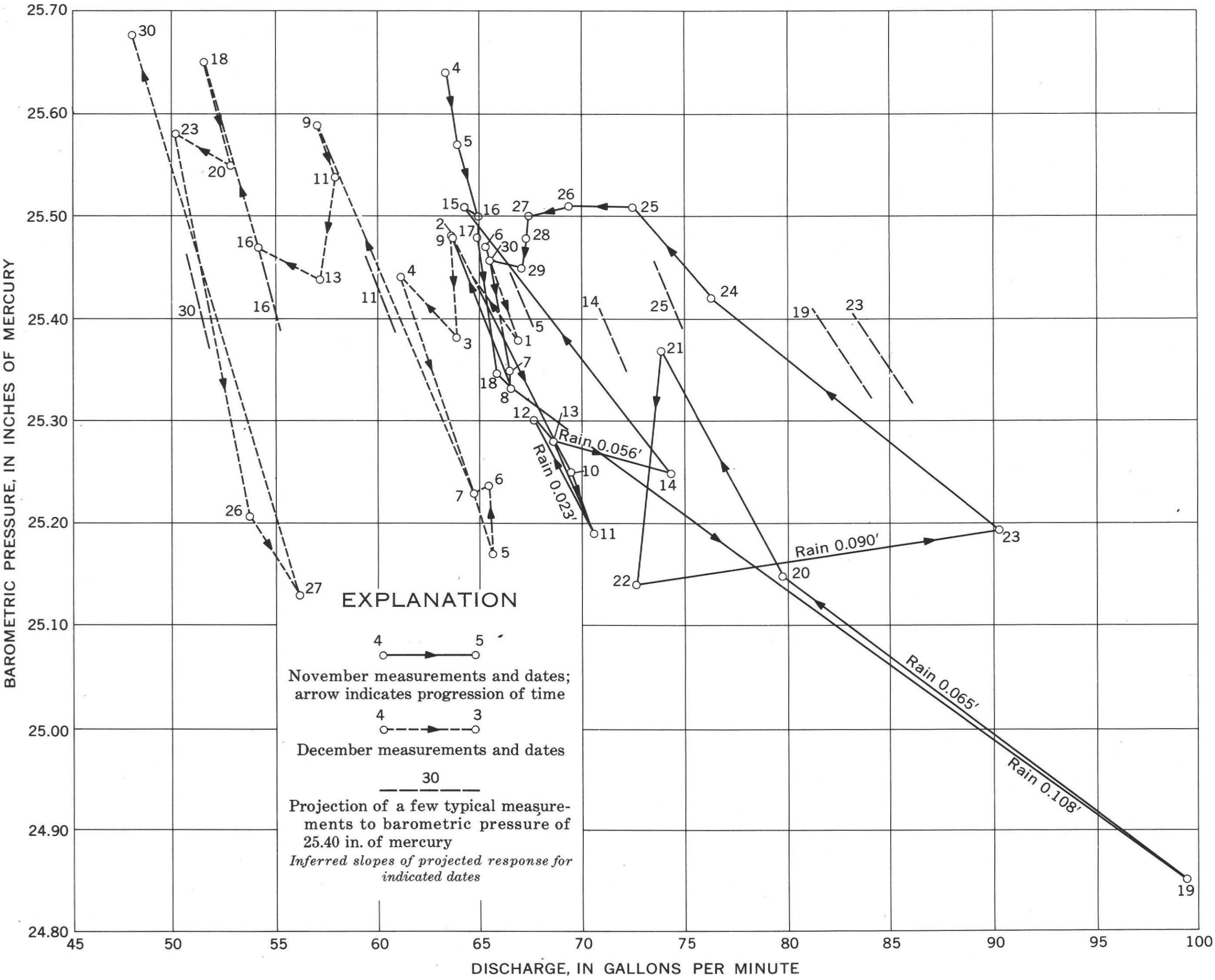


Figure 14.—Total measured discharge compared with barometric pressure for successive series of measurements, November and December 1946.

in regard to its barometric response. Structural, chemical, and isotopic evidence indicates, however, that both terrace systems are structurally related and must have interconnections at depth. This apparent anomaly is resolved if the interconnections are relatively deep and if barometric response is determined largely by boiling relationship at levels above rather than below the most restricted interconnections. According to this reasoning, the amount of water diverted at depth into the Low Terrace subsystem is limited by the dimensions of the tightest restrictions along the channels. The total discharge from the Low Terrace is considerably less than the total from the Main Terrace system, in spite of the disadvantage caused by the higher altitude of the latter. If the two terraces are indeed connected, but channels of the Low Terrace system suddenly become highly permeable, all thermal water of the total system would then discharge from the Low Terrace.

The barometric response is probably determined largely within the upper few hundred feet of each subsystem. Within this relatively shallow zone, as will be seen from drill-hole data, temperatures are at, or close to, theoretical boiling points for the existing pressures. A decrease in barometric pressure lowers the boiling temperature at each point at depth; a higher proportion of water then vaporizes to steam and the water column thereby expands. When water levels rise near the crests of the terraces, the higher altitude springs respond with increased discharge. When barometric pressure increases, less boiling occurs and water levels and discharges decline.

We might expect that when a barometric high is sustained for sufficient time, the discharge would first

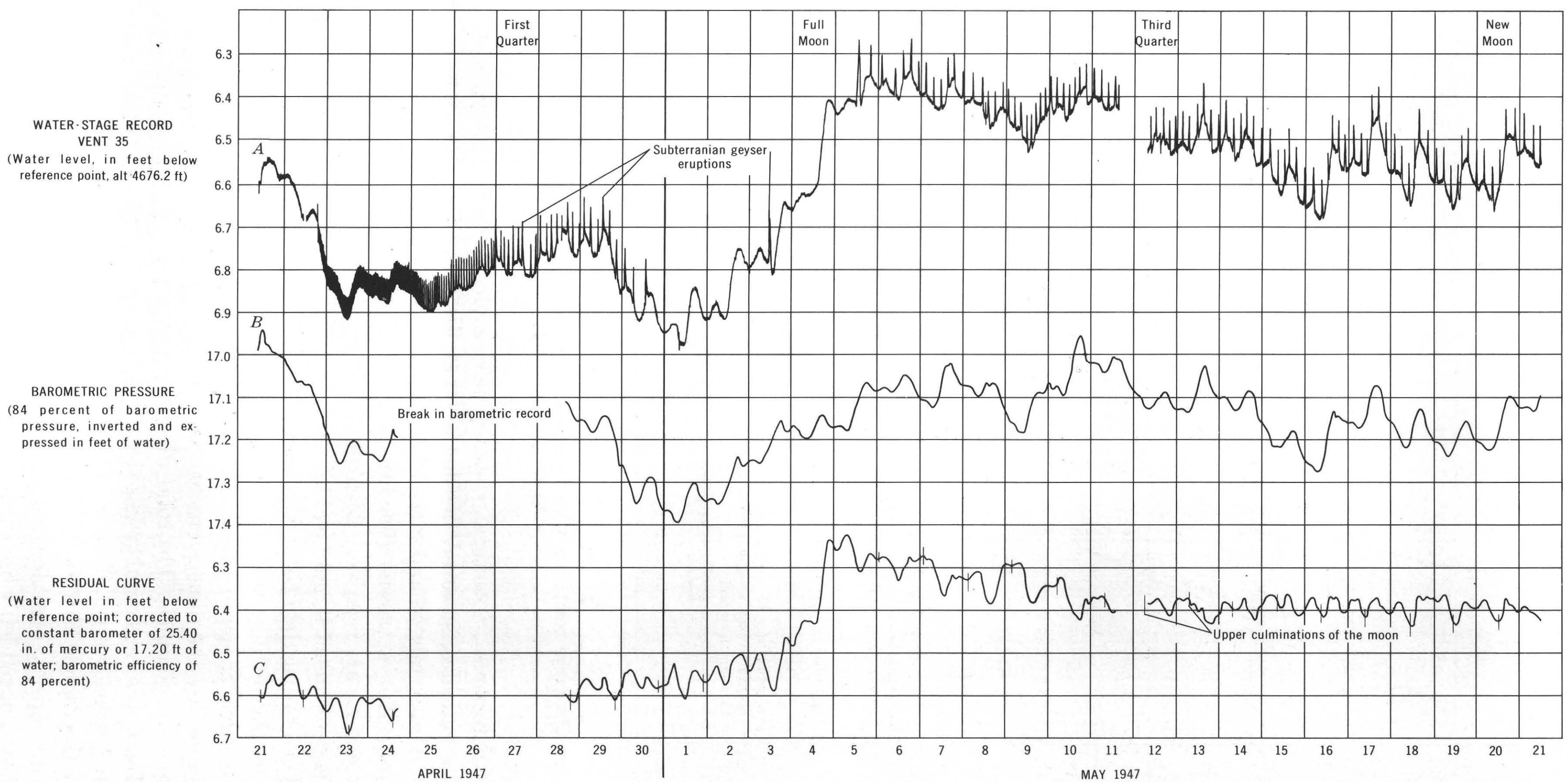


Figure 15.—Fluctuations of water level, vent 35, Main Terrace, showing subterranean geyser eruptions and an unexplained random change on May 4, 1947.

decrease and then gradually increase to its former average as the total system adjusted to the new pressure environment. Higher-than-average barometric pressure has occasionally been sustained for as long as 6 days. One example from the Low Terrace is shown in figure 11 for the interval of December 15–20, 1946. The pressure-corrected curve *C* first decreased slightly on the 15th in response to increasing pressure, and then, during the following 5 days, slowly rose nearly 0.1 foot. Even sharper rises in residual curve *C* occurred from December 8 to 10 and 22 to 24 during shorter periods of high barometric pressure. Barometric pressures are seldom sustained much below average for more than 2 or 3 days, and most responses to low pressure are complicated by precipitation.

The evidence is less clear that water levels on the Main Terrace adjust to sustained high or low barometric pressure. Vent 35 shows fluctuations in residual curves *C* of figures 10 and 15, but neither these nor other unpublished records clearly demonstrate adjustment to sustained periods of high or low pressure.

Likewise, the flowing springs of the Main Terrace do not clearly adjust to sustained barometric highs or lows. The longest sustained period of high pressure of figures 12 and 14 was from December 15 to 21 (the same time interval considered in figure 11, which also shows the detailed barometric record). There is a suggestion in figure 14 that the long-term decline in discharge that started about November 25 was temporarily interrupted from December 16 to 20, perhaps in adjusting to the sustained barometric high.

The evidence for slight adjustment of the Low Terrace to sustained barometric highs and lows and the lack of clear evidence for the Main Terrace can be explained if the reservoir of involved fluids is much larger for the Main Terrace. The springs and vents of high barometric response on the Low Terrace (pl. 3) include the group from 32se to 30n and 29; the area of high response is perhaps only 200 by 250 feet, and open fractures are rare. In contrast, all the upper part of the Main Terrace has moderate to high response; its area is roughly 150 feet wide by 1,500 feet long, and open fissures with high storage capacity are relatively abundant. The difference in responsive reservoir capacity of the two terraces probably explains the slight differences in their long-term barometric response.

EFFECTS OF PRECIPITATION

CLIMATIC SETTING OF THE AREA

The climatic setting is a very important aspect of any hydrologic study. Only meager data are available for the thermal area, but a weather station has been maintained for many years at Reno, 11 miles north of the springs. From 1870 to 1965 the Reno station was located at six or seven different sites ranging in altitude from 4,800 to 4,400 feet, the ground level at the Reno airport, where the station has been located since 1942. The average altitude of the thermal area is about 4,750 feet, or 350 feet higher than the airport. This difference results in slightly lower temperatures at Steamboat Springs. Tables 8 and 9 contain the most pertinent temperature data for the Reno station, and tables 10 and 11, the precipitation data.

TABLE 8.—*Average temperature by year, Reno weather station*

[Average of monthly mean]

Year	°F	Year	°F
1940	53.3	1947	50.0
1941	51.6	1948	47.8
1942	50.1	1949	47.7
1943	50.1	1950	50.4
1944	48.5	1951	49.3
1945	49.4	1952	48.4
1946	49.3	1953	49.2

Average, 1899–1953, 50.1°F (10.1°C)

TABLE 9.—*Monthly mean temperature, 1899–1953, Reno weather station*

Month	°F	Month	°F	Month	°F
January	31.8	May	54.7	September	61.0
February	36.3	June	62.5	October	51.2
March	41.3	July	70.5	November	41.3
April	47.9	August	69.0	December	33.7

TABLE 10.—*Yearly precipitation, Reno weather station*

Year [1]	Precipitation (inches)	Year [1]	Precipitation (inches)
1870–71	[2] 2.42	1891–92	6.00
1871–72	4.72	1892–93	11.30
1872–73	3.73	1893–94	5.64
1873–74	5.70	1894–95	6.84
1874–75	4.09	1895–96	8.03
1875–76	6.42	1896–97	9.95
1876–77	5.30	1897–98	6.55
1877–78	5.42	1898–99	5.67
1878–79	4.04	1899–1900	9.05
1879–80	6.40	1900–1901	11.35
1880–81	5.74	1901–02	6.71
1881–82	6.33	1902–03	7.49
1882–83	3.15	1903–04	8.54
1883–84	7.20	1904–05	7.40
1884–85	3.17	1905–06	7.76
1885–86	6.35	1906–07	11.70
1886–87	5.00	1907–08	6.33
1887–88	4.93	1908–09	9.50
1888–89	7.23	1909–10	6.16
1889–90	15.36	1910–11	12.73
1890–91	11.49	1911–12	4.76

See footnotes at end of table.

TABLE 10.—*Yearly precipitation, Reno weather station*—Con.

Year [1]	Precipitation (inches)	Year [1]	Precipitation (inches)
1912–13	6.22	1934–35	6.46
1913–14	14.93	1935–36	6.50
1914–15	5.44	1936–37	6.35
1915–16	9.37	1937–38	11.90
1916–17	6.86	1938–39	4.16
1917–18	6.03	1939–40	10.89
1918–19	9.67	1940–41	6.90
1919–20	7.93	1941–42	8.32
1920–21	6.66	1942–43	8.65
1921–22	10.18	1943–44	6.10
1922–23	9.68	1944–45	7.44
1923–24	3.79	1945–46	6.54
1924–25	6.82	1946–47	4.25
1925–26	6.11	1947–48	3.13
1926–27	8.28	1948–49	5.86
1927–28	6.06	1949–50	6.92
1928–29	5.18	1950–51	8.85
1929–30	5.90	1951–52	11.19
1930–31	6.47	1952–53	6.40
1931–32	8.90		
1932–33	4.13	Average, 1870–1953	7.20
1933–34	8.56		

[1] Data before 1944–45 are for the fiscal year, July through the following June; for 1944–45 and later years, precipitation is by hydrologic year (Oct. 1 through following Sept. 30).
[2] December 1870 to June 1871, only.

TABLE 11.—*Average monthly precipitation, 1899–1953, Reno weather station*

	Inches		Inches		Inches
January	1.39	May	0.52	September	0.23
February	1.07	June	.32	October	.38
March	.74	July	.22	November	.60
April	.48	August	.20	December	.95

The climate of western Nevada is generalized below (U.S. Dept. Agriculture, 1941).

Nevada lies just east and to the leeward of the Sierra Nevada Range, an effective barrier to precipitation from the generally eastward-drifting air, which loses most of its moisture in ascending the western slopes of the mountains. One of the greatest contrasts in precipitation within a short distance found in the United States occurs between the Western, or California slopes of the Sierra and their eastern slopes and the contiguous lowlands of western Nevada. In ascending the west slope, the air loses most of its moisture through condensation and precipitation, and descending the eastern slope it is warmed by compression, so that very little precipitation occurs. This rain barrier is felt not only in the extreme western part but generally throughout the State, with the result that the lowlands of Nevada are largely desert or semidesert.

With its varied and rugged topography—its mountain ranges, narrow valleys, and low, sage-covered deserts, ranging in elevation from about 1,500 to more than 10,000 feet—Nevada presents wide local variations of temperature and rainfall. The most striking climatic features are bright sunshine, small annual rainfall in the valleys and deserts, heavy snowfall in the higher mountains, dryness and purity of air, and phenomenally large daily ranges of temperature.

Because of the high altitude and the extreme dryness and clearness of the air, there is rapid nocturnal radiation of heat, resulting in wide daily ranges in temperature. Even after the hottest days, the nights are cool; at Reno the average range between the highest and the lowest daily temperatures is 22° F. in January, increasing month by month to 35° in July. * * * In extreme instances the daily range may become as great as 50° to 60°.

Nevada has, on an average, less precipitation than any other State and most of that occurs during the winter season, with the summer falls very light. * * * Variations in precipitation are due mainly to differences in elevation and exposure to moisture-bearing winds.

Humidity is normally low, and the dryness of the air makes both the heat of summer and the cold of winter less disagreeable.

Nevada has a generous supply of sunshine, the average percentage of the possible amount at Reno * * * being 74 * * *. The low humidity and ample sunshine produce rapid evaporation.

Winds are generally light. Storms with high winds rarely occur and more rarely still cause appreciable damage. The prevailing wind direction is west, though at a few stations, because of local topography, it is south or southwest.

The precipitation year 1951–52 (table 10) was one of the highest on record and the year 1947–48, the lowest. Therefore, the interval of detailed study, from 1945 to 1952, fortunately included the maximum contrast in precipitation that could be expected.

When detailed measurements were started in 1945, two rain gages were installed within the thermal area and two others were located within 5 miles in the hope that some influence by localized storms could be detected, thereby indicating the recharge area for the meteoric water supply of the system. The gages, however, proved to be of greater interest to sheep, cows, and boys with guns than to recharge problems. One gage just west of the Main Terrace was kept in reasonably good condition from 1945 through 1947 and for less regular intervals thereafter.

Table 12 compares precipitation from individual storms for Reno and the Main Terrace. Except for several local summer thunderstorms, precipitation at Steamboat Springs is considerably the heavier. Individual storms differ greatly, and the percentage difference is commonly greater for the light storms. During near-average years, precipitation is probably about 50 percent higher at Steamboat Springs than at Reno, owing primarily to the difference in altitude of 260 feet between the stations; both are located about the same distance (5 miles) from the front of the Carson Range.

In 1947, a year of minimum precipitation, precipitation at Steamboat Springs was double that of Reno, but in the two near-normal years of 1945 and 1946, it was only 45–50 percent greater. These data are far too scanty to be conclusive, but they suggest that the percentage difference decreases with increased precipitation.

TABLE 12.—*Precipitation, in inches, by individual storms at Steamboat Springs and Reno*

Date	Reno [1]	Steamboat Springs [2]
1945		
June 22–23	0.75	0.24
July 7–8	.07	.40
Oct. 6–7	.09	.19
14–15	.19	.52
29–31	1.86	2.58
Nov. 10–11	.08	.26
Dec. 4	.13	.25
11	.16	.18
15–16, 21–24	1.29	2.24
Total for measured storms	4.62	6.86
Total for period	5.19	[3] 7.5
1946		
Jan. 28–Feb. 3	0.28	0.34
Feb. 10	.12	.14
Mar. 18–20	.71	.84
July 22–25	.84	.28
Oct. 1–2	.34	.48
Oct. 7	.09	.06
Nov. 7, 11–14	.76	1.09
19–20	.99	2.08
22–23	.29	1.08
Dec. 22–25	.29	.34
27	.10	.23
Total for measured storms	4.81	6.96
Total for year	5.59	[3] 7.9
1947		
Jan. 27–28	0.15	0.06
Feb. 9	.01	.06
11–12	.49	.58
17	.01	.12
Mar. 1	.06	.36
23, 28	.06	.10
May 26–27	.21	.24
30	.02	.06
Aug. 10	.10	.29
Oct. 10–16	.17	.48
Nov. 4	.05	.24
Dec. 9	.01	.12
Total for measured storms	1.34	2.71
Total for year	1.55	[3] 3.1

[1] Altitude 4,407 ft, Reno weather station at airport.
[2] Altitude 4,665 ft, about 100 ft west of crest of Main Terrace.
[3] Estimated.

SEASONAL VARIATIONS

According to Becker (1888, p. 349), "Old residents informed me that the quantity of water flowing from the springs varies from year to year, being greater in years of heavy rainfall than in dry seasons and greater in spring than in autumn."

Plate 4 includes some evidence for seasonal variations in discharge. Considered by calendar year, the minimum monthly discharge in curve *D* occurred during the summer months for 5 of the 8 years and the maximum monthly discharge occurred during the winter months for 6 of the 8 years. However, curve *D* shows many individual fluctuations that have no evident relation to seasonal variation in precipitation, effects of individual storms, changes in barometric pressure, or any other obvious causes. One justification for continuing the detailed measurements at Steamboat Springs for 6 full years, 1946–51, and parts of 2 other years, 1945 and 1952, was the possibility that extended observations might clarify effects of all major influences, and the causes of minor effects would then become more evident. To some extent, but not entirely, this hope has been realized.

The data of plate 4, averaged by quarter years to minimize random variations, are shown in table 13. These are numerical averages of individual series of measurements and are shown graphically in figure 16. For about 60 percent of the individual quarterly changes, precipitation and average discharge changed in the same direction, but the remaining 40 percent were in opposite directions. No systematic lag in discharge is evident, and precipitation and discharge do not have a direct straight-line relationship.

TABLE 13.—*Average of measurements of flowing springs by quarter years, 1945–52, Steamboat Springs* [1]

Year	Quarter	Discharge	Cl (ppm)	Temperature (°C)	Precipitation [2] (inches)
1945	[3] 2	58.4	913	88.8	1.52
	3	60.0	915	89.3	.33
	4	46.6	906	87.2	3.89
1946	1	56.5	913	86.6	1.49
	2	50.7	925	86.8	.22
	3	46.8	928	86.3	.94
	4	62.3	912	85.2	2.94
1947	1	44.1	900	83.0	.79
	2	33.3	898	82.6	.40
	3	30.0	898	84.1	.12
	4	34.6	902	81.4	.24
1948	1	31.2	899	80.9	.38
	2	29.0	889	82.5	2.44
	3	39.7	898	86.7	.09
	4	49.9	905	85.6	.89
1949	1	60.2	900	82.1	1.82
	2	41.1	904	81.4	2.06
	3	30.4	903	81.7	1.09
	4	35.5	892	81.5	1.56
1950	1	49.2	889	82.1	2.80
	2	50.3	899	85.6	1.43
	3	41.9	905	86.1	1.13
	4	41.2	887	83.9	4.25
1951	1	55.3	872	84.1	1.65
	2	51.1	871	85.3	2.45
	3	41.7	872	87.8	.50
	4	44.7	879	88.6	3.52
1952	1	77.1	886	88.2	4.53
	2	76.8	900	86.9	1.74
	[4] 3	36.2	893	83.1	1.40

[1] Numerical average of weekly measurements; Cl and temperature weighted in proportion to discharge of individual springs.
[2] Precipitation at Hubbard Field Airport, Reno, Nev.
[3] June only.
[4] July–August.

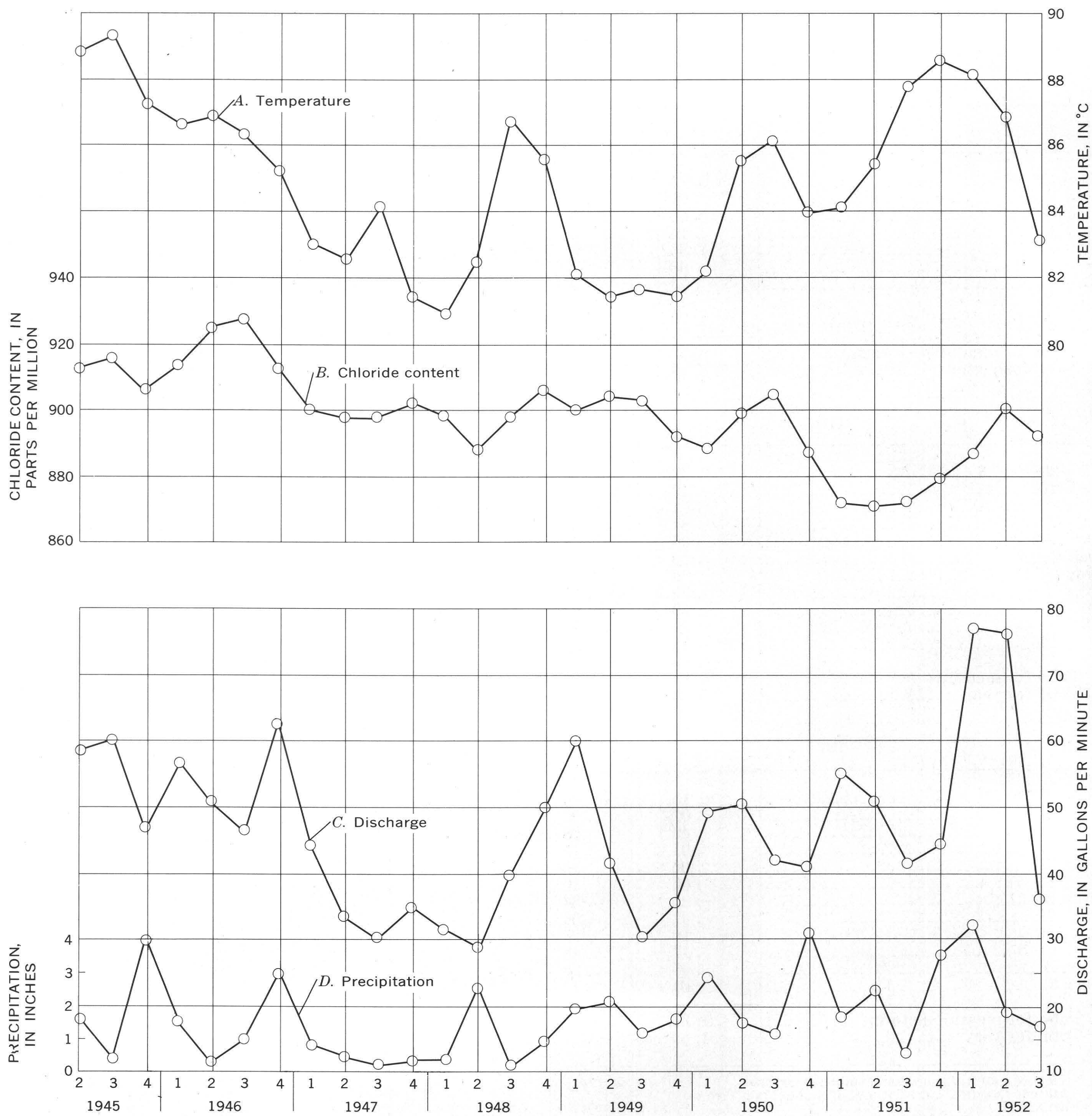


Figure 16.—Average of measurements of flowing springs by quarter years, 1945–52.

In table 14, seasonal averages are shown from the quarterly data. Confusing random variations are largely eliminated here, and average discharge is clearly lowest in the summers and highest in the winters. Becker's impression of seasonal variation is supported by these average measurements, but the greatest seasonal contrast occurs between winter and summer rather than spring and autumn.

The discharge, salinity, and temperature data of tables 13 and 14 and figure 16 are numerical averages of individual series of measurements that are weighted averages of total discharge from all measured springs.

The chloride content shows a slight but significant seasonal variation, tending to be highest (901 ppm) in the average summer quarter (table 14) and lowest (894 ppm) in the average winter quarter. The difference of

TABLE 14.—*Seasonal averages from quarterly data, 1945–52*

Quarter	Period	Discharge	Cl (ppm)	Temperature (°C)	Precipitation (inches)
Winter	January–March	53.4	894	83.8	1.92
Spring	April–June	48.8	900	84.9	1.53
Summer	July–September	39.7	901	85.6	.70
Fall	October–December	45.0	898	84.7	2.47

7 ppm is a little less than 1 percent of the total. In view of the fact that nearly 5,000 chloride analyses are involved in these averages, any seasonal difference greater than 1 or 2 ppm is probably significant. This seasonal variation is due largely to dilution by precipitation, which is at a minimum during the summer months. Residual concentration of chloride by evaporation is also greatest in summer, when temperatures of water and air are somewhat higher.

Seasonal changes in discharge and temperature are also given in tables 13 and 14 and in figure 16. The maximum average temperature, as expected, is generally attained in the summer quarters. Slight effects of air and near-surface ground temperatures also undoubtedly exist. The heaviest precipitation normally occurs in the late fall and winter months and is normally accompanied by a temporary increase in discharge and a decrease in temperature (pl. 4).

YEARLY VARIATIONS

In Nevada, as in most other western States, precipitation occurs largely in the winter months and is commonly stored as snow. Runoff and other hydrologic relationships related to precipitation are most clearly shown by a precipitation year that starts just before the autumnal storms; October 1 is the commonly accepted starting date. The data for Steamboat Springs are shown in table 15 and figure 17.

TABLE 15.—*Yearly average of weighted measurements of flowing springs, 1945–1952, by precipitation year (October through following September)*

Precipitation year	Discharge (gpm)	Cl (ppm)	Temperature (°C)	Precipitation, Reno, Nev.
1944–45 [1]	54.7	914	89.0	7.44
1945–46	50.2	918	86.7	6.54
1946–47	42.4	902	83.7	4.25
1947–48	33.7	897	82.9	3.13
1948–49	45.4	903	82.7	5.86
1949–50	44.2	896	83.8	6.92
1950–51	47.3	876	85.3	8.85
1951–52 [2]	58.7	889	86.7	11.19
Average	47.1	899	85.1	6.77

[1] June–September 1945, except for precipitation.
[2] Through August 1952.

In curve *C* of figure 17 discharge shows a nearly straight-line relationship to precipitation, except that discharge decreased gradually from 1946–47 to 1950–51 relative to precipitation, with a particularly notable decrease in 1949 and 1950.

In general, Becker's statement (1888, p. 349) is confirmed: "the discharge of the hot springs is greatest in years of heavy rainfall and is less in drier years."

For 5 of the 7 years of observations (fig. 17, curve *C*), each inch of change in precipitation changed the rate of discharge by about 4 gpm.

The chloride-discharge graph (fig. 17, curve *B*) shows that chloride tends to be high when discharge is high, and low when discharge is low. Two of the seven tie lines between adjacent years (1948–49 to 1950–51) are not in accord with this tendency, but the other five are reasonably consistent.

This tendency for a direct rather than an inverse relationship between discharge and salinity is one of the more unexpected results of the present study. If chloride is contributed largely by deep volcanic emanations and is extensively diluted within the system by meteoric water (White, 1957a), salinity should decrease when precipitation and discharge increase. Even if the dissolved salts are derived from salt deposits, other rocks, or previously deposited volcanic emanations, a decrease in rate of percolation and discharge should result in higher salinity, perhaps involving a time lag. None of these simple models fits the facts.

A major shift in relationships within the spring system evidently occurred from 1949 to 1951 when chloride content decreased by about 30 ppm, or more than 3 percent of the total in the average water that was being discharged. During this interval of time, the chloride content was decreasing while precipitation showed a net increase. This could be interpreted as increasing dilution of chloride of some deep origin, except for the fact that discharge of the springs was also decreasing relative to precipitation (curve *C*).

Alternatively, perhaps the rate of evolution of water and chloride of volcanic origin decreased from 1949 to 1951, the decrease in volcanic water being balanced by a slight increase from precipitation.

The temperature discharge graph (fig. 17, curve *A*) shows that the average annual temperature of the flowing springs is generally high when discharge is high and is low when discharge is low. This relationship is also supported by many measurements of individual springs. Temperature nearly always falls as the discharge of a spring declines and perhaps ceases to flow. Figure 18 shows a series of measurements from spring 22 in which temperature in general was clearly related

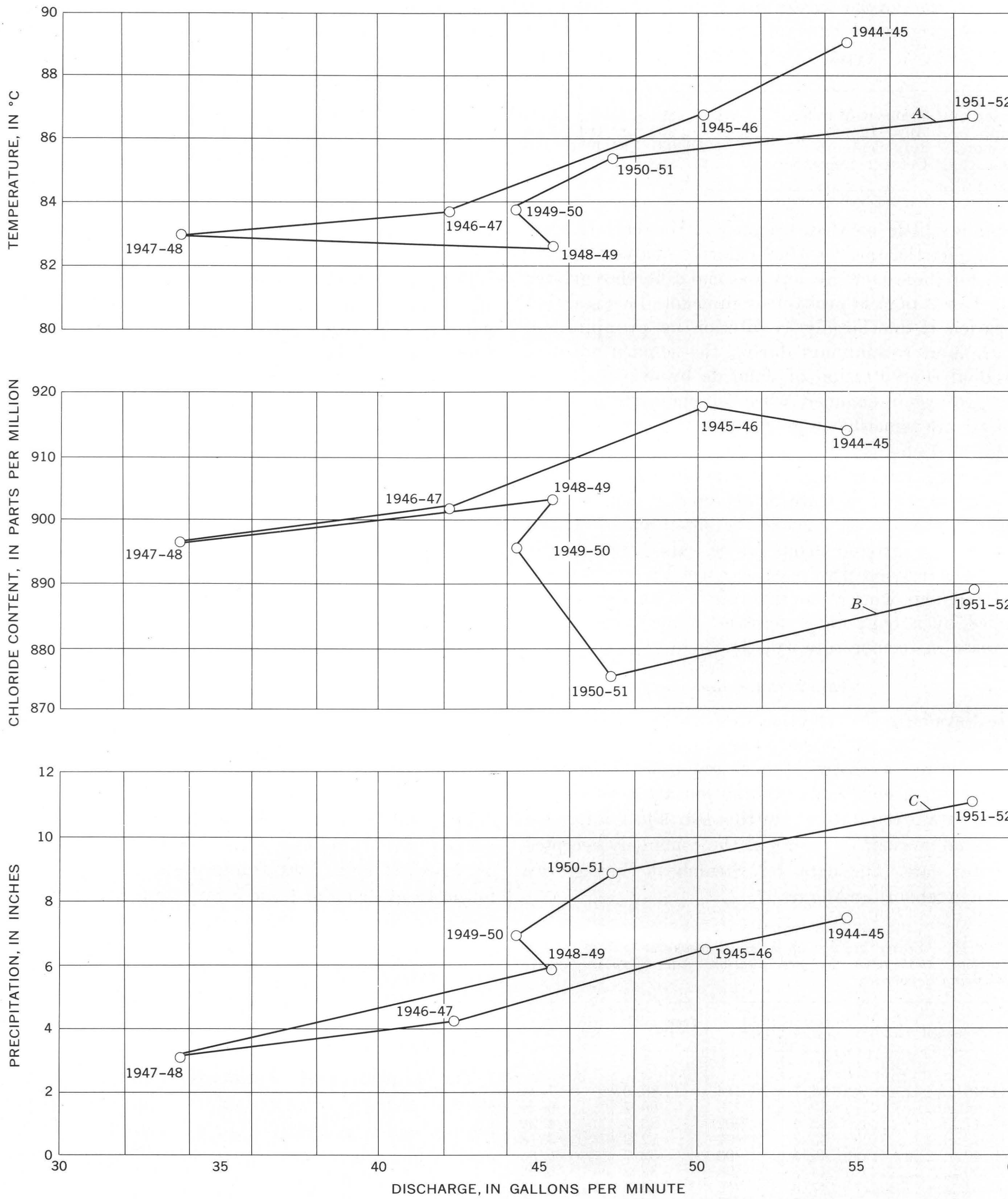


Figure 17.—Yearly average discharge in relation to temperature, chloride content, and precipitation, by precipitation years, October through following September.

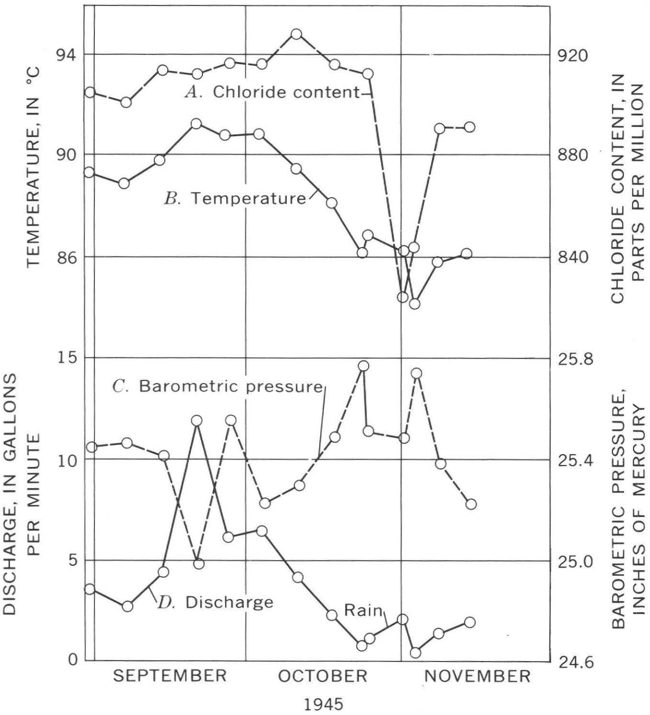


Figure 18.—Relationship of discharge, temperature, and barometric pressure, in spring 22, Main Terrace, and chloride content.

to discharge, which, in turn, was usually highly responsive to changes in barometric pressure.

Most changes in discharge are accompanied almost immediately by changes in temperature, but a definite lag in response was found in spring 2 during September and October 1947 (fig. 19). The discharge relationships of springs 2 and 5 were reversed abruptly about September 15, after strong local earthquakes that occurred late in the evening of September 7. The changes, if earthquake induced, must have occurred at considerable depth within the system, because no change in discharge had occurred by 9 a.m. of September 8. The maximum response in chloride content of spring 5 was delayed for at least 1 additional week and that of spring 2 for 2 weeks after the change in discharge had occurred. The temperature of spring 5, which was near boiling, remained almost unchanged, but that of spring 2 increased gradually and finally attained a maximum 3 weeks after the changes in discharge had occurred.

According to one possible explanation, the earthquake caused some change at depth that permitted the flushing out of a connection between springs 2 and 5. Other detailed changes in the springs suggest that gelatinous silica settles out in suitable, nearly stagnant, parts of the fissures, resulting in a decrease in permeability; and that if such an accumulation is flushed out, permeability and rate of discharge then increase.

EFFECTS OF INDIVIDUAL STORMS

Precipitation of as much as half an inch per storm generally has no detectable effects on the spring system (pl. 4 and fig. 12). Storms of 1 inch or more, on the other hand, generally have clearly observable effects if measurements are made within 1 or 2 days after such a storm (table 16). If observations are delayed until 2 days or more after a storm, effects are usually not detected. Thus, the discharge effects are only temporary, even for larger storms and differ from one storm to another for reasons not wholly understood. No increase in discharge can be credited definitely to the storm of November 9–10, 1949, with precipitation of 1.23 inches, even though measurements were made only 1½ days after precipitation ceased.

The detailed records of figure 12, especially curve *E*, provide good evidence that discharge responds almost immediately to precipitation of half an inch or more. When daily measurements are made and a barometric correction is applied, after effects of a major storm can generally be recognized for at least 2 or 3 days. The

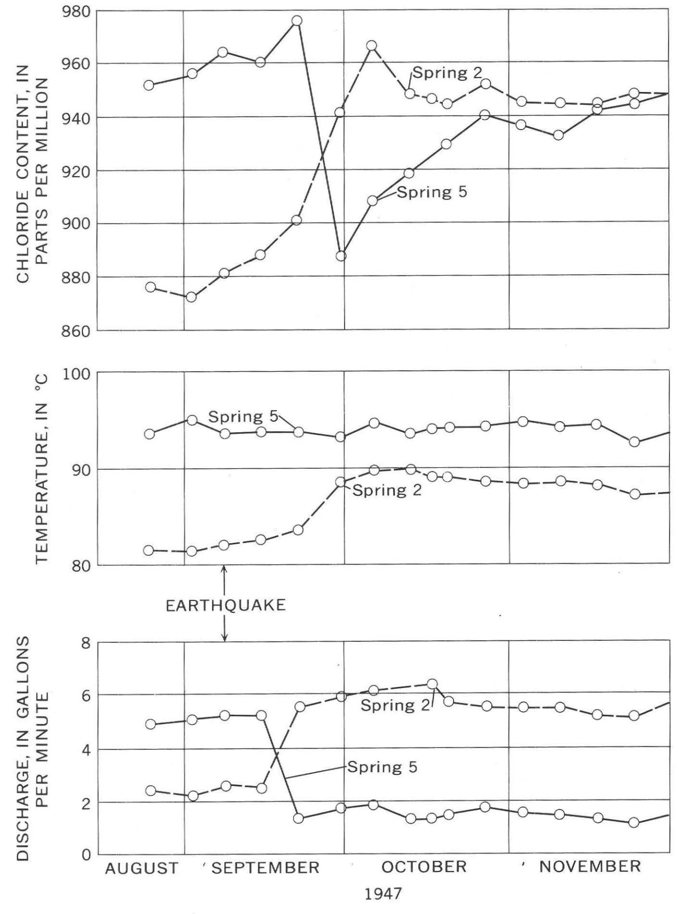


Figure 19.—Interrelated changes in springs 2 and 5, Main Terrace, perhaps resulting from local earthquakes.

TABLE 16.—*Effects of individual major storms on discharge of springs*

Storm	Precipitation Reno (inches)	Immediate surficial effects [1]	Time lapse to next series of observations (days)	Time lapse to next increase in discharge that might be a lag effect of the storm [2] (weeks)
Oct. 29–31, 1945	1.86	Present	1	6–9.
Dec. 21–24, 1945	1.23	do	1	7.
Nov. 13–14, 1946	.58	do	Few hours	(?).
Nov. 19–23, 1946	1.28	do	do	Not less than 9 months.
May 18–20, 1948	.95	(?)	5	3–4.
May 13–15, 1949	1.20	Slight	After only ½ in. of the total rain.	10?—none other earlier than 4½ months.
Nov. 9–10, 1949	1.23	(?)	1½	5–10?
June 15, 1950	.59	None?	1	2–4.
Oct. 26–27, 1950	1.11	(?)	7	11.
Nov. 18–20, 1950	1.12	(?)	11	8.
Jan. 18–19, 1951	.89	(?)	7	1–3.
Apr. 25–28, 1951	.91	(?)	7	1–3.
June 20–21, 1951	.67	(?)	8	8–12.
Oct. 24–25, 1951	.95	(?)	8	7–9.
Dec. 28–30, 1951	1.34	Present	During storm.	4–6.
Jan. 11–15, 1952	1.39	(?)	10	2–4.
Mar. 14–15, 1952	1.01		6	3–6.
Apr. 25, 1952	1.60	(?)	7	10–11.

[1] Includes increase in discharge and decrease in temperature and chloride content.
[2] From pl. 4 and fig. 12.

effect commonly is not directly proportional to the magnitude of precipitation. For example, the increase in discharge attributable to the storm of November 22–23, 1946 (fig. 12), was greater and persisted longer than the effects of the higher total precipitation of November 18 and 19, perhaps because sinter and sediment above the water table were already saturated, retaining as much moisture as possible.

Water-stage records from nonflowing vents yield more detailed data than periodic measurements of flowing springs. The record from vent 31n on the Low Terrace (fig. 11) indicates clearly on residual curve *C* that the effect of 0.58 inch of precipitation on November 13–14, 1946, was of very short duration and became unrecognizable within 24 hours after the end of the storm. For comparison, 2.08 inches of rain and snow on November 19–20 increased the water level of residual curve *C* by 0.25 foot, and only 1.08 inches of rain on November 22–23 produced an additional increase in water level of 0.15 foot. Each higher level tended to be maintained.

At vent 35 on the Main Terrace, in contrast, significantly different effects were produced by the same storms (fig. 20). A major decline in water level was taking place from November 11 to 19, probably from some local shift of activity. After precipitation ceased from the storm of November 13–14, the water level rose at least 0.2 foot in residual curve *B*, presumably attaining a maximum a day or more after precipitation ceased (part of record was lost). After further decline, precipitation on the 19th produced an almost immediate rise in water level with a second maximum about 1½ days after the storm had ceased. Response was again somewhat different for the storm of November 22–23, when the water level rose almost immediately, but very soon started to decline and continued down to the minimum 6 days later.

The type and degree of response to precipitation clearly cannot be predicted with confidence, especially if other random changes are also occurring. Some precipitation falls directly into fissures and springs, but the total direct fall is negligible. A little more, in rivulets and sheet wash, runs into fissures and springs but this also is slight, except during drenching downpours. Most of the immediate and slight-lag effects are probably from precipitation that falls on porous sinter and then seeps down into the spring system. The delayed responses in vent 35 that reached their maxima on November 15 and 21 (fig. 20, curve *B*) may be related to precipitation on alluvium just west of the Main Terrace. The magnitude of response and the lag effects are probably related to the amount of precipitation and the preexisting moisture content of ground above the water table. When this ground has nearly as much water as it can retain, response to new precipitation will be prompt and relatively high in magnitude. When the water content of the ground is low, new precipitation is retained largely above the water table and effects on the spring system are relatively slight.

The immediate and short-term lag effects so far discussed seem to be entirely near surface in origin, as might be expected. A water-stage record from the Rodeo well on the Main Terrace (pl. 3) is reproduced as figure 21. The pressure-corrected curve *B* shows no immediate response to the storms of November 18 and 20, 1950 (total precipitation, 1.12 in.), but a gradual increase of less than 0.1 foot, attaining a maximum on November 22, may be a lag effect of deeper origin. This well is cased to 149 feet in depth, so that direct near-surface influences are excluded; but it is surprising that changes in water levels in nearby fissures were not transmitted as pressure changes through aquifers below the casing. The actual effects of the two storms on water levels in the fissures are not known, because the water-stage records from natural vents on the Main and Low Terraces during this period were lost, but a significant change was expected from a storm of this magnitude. An aquifer relatively low in temperature and chloride was penetrated in the Rodeo well near 164 feet in depth, just above granodiorite bedrock. Precipitation in and near

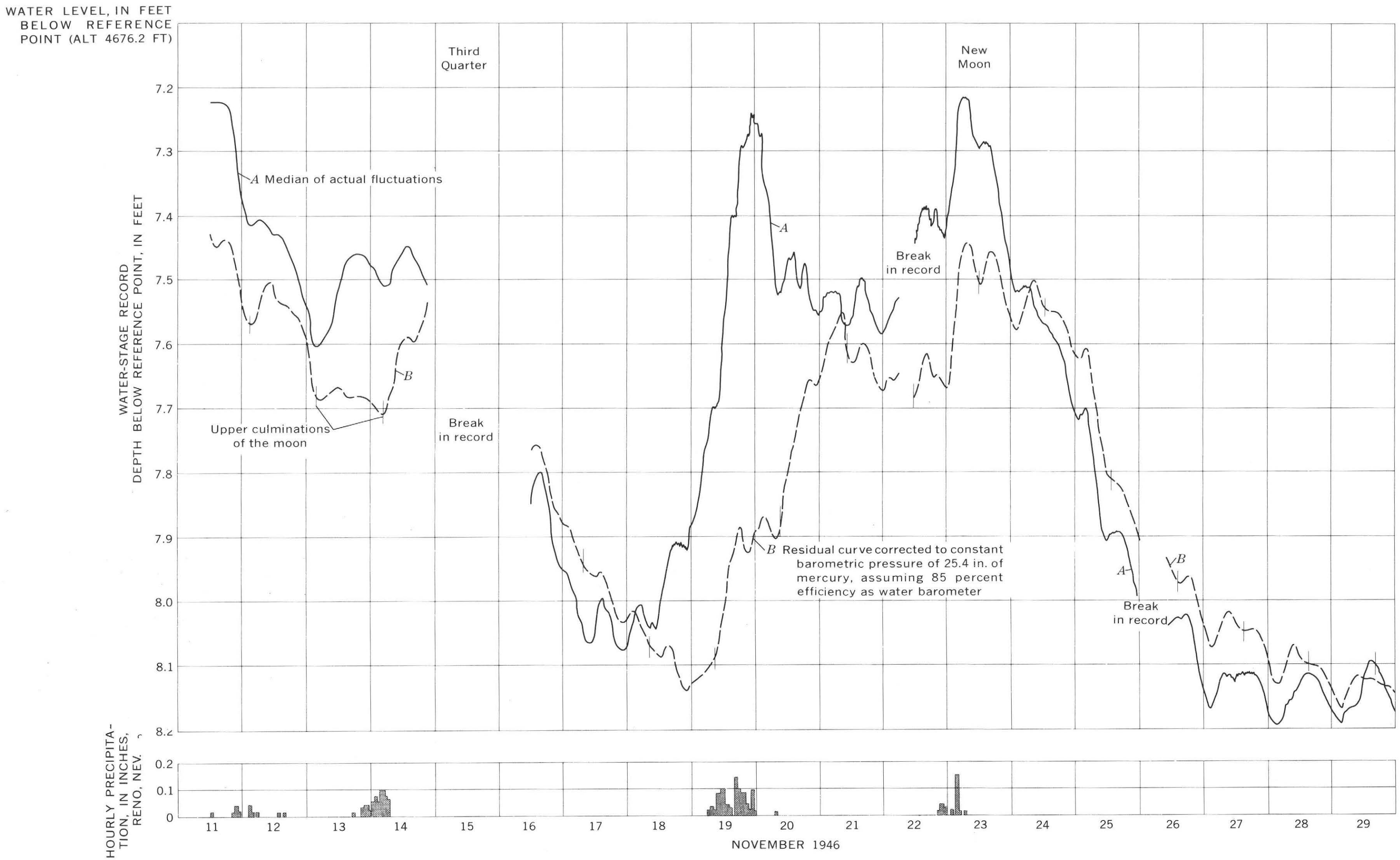


Figure 20.—Fluctuations of water level, vent 35, Main Terrace, showing influence of heavy precipitation.

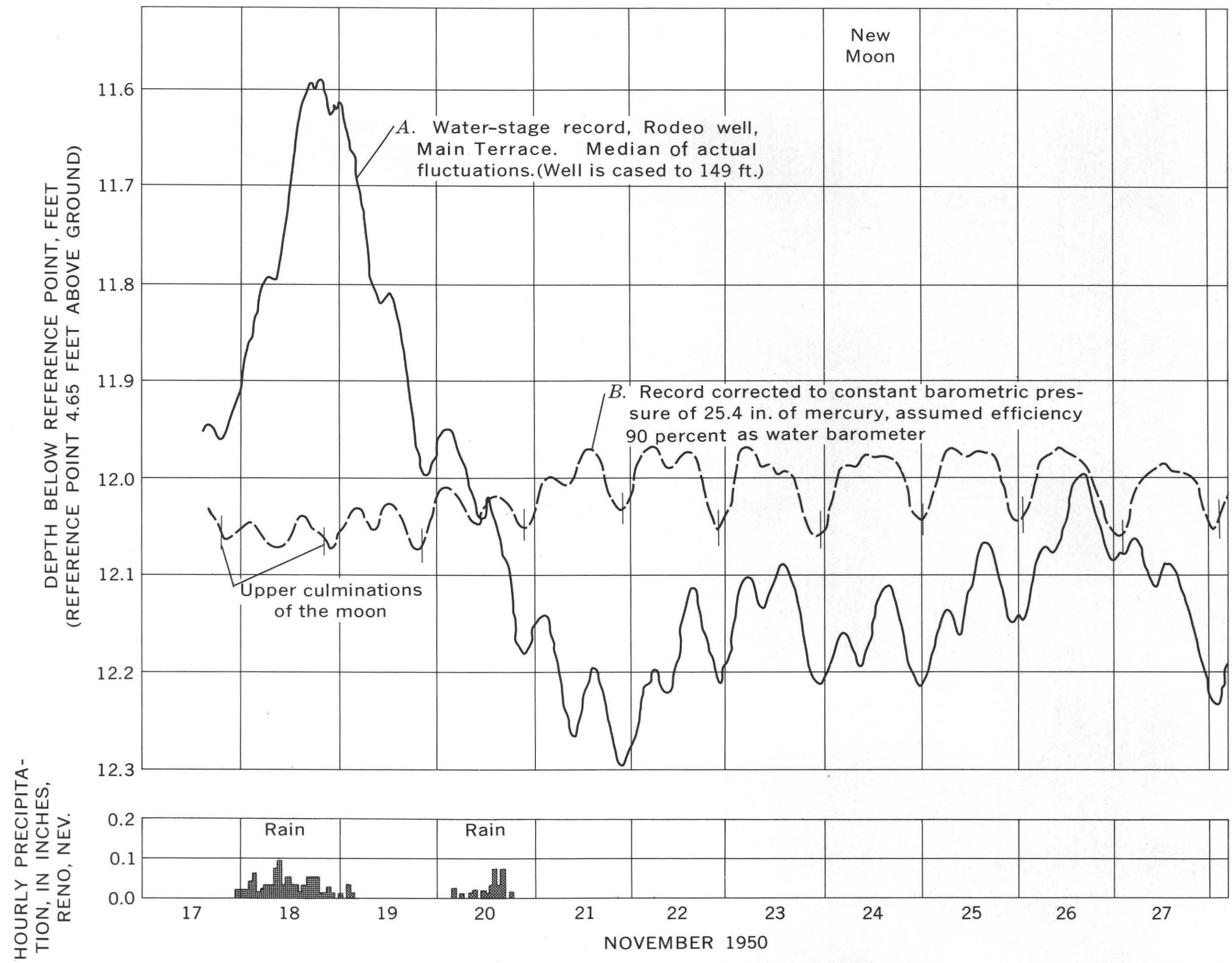


Figure 21.—Fluctuations of water level, Rodeo well, Main Terrace, showing barometric and earth-tidal influences and perhaps a delayed response to precipitation.

Pine Basin west of the Main Terrace may seep underground and into the system through this 164-foot aquifer, thereby accounting for the delayed response.

The very individualistic responses of different springs to precipitation and other factors are not evident from the averaged data (pl. 4 and fig. 12), but some differences are shown in figures 13 and 22, which show some details for the period of November–December 1946, and in figure 23. Springs 5 and 50 responded in discharge in proportion to the total discharge of all springs. Even the computed average rate of discharge of a small geyser (fig. 22, curve *B*) is similar to that of total discharge until its activity ceased on December 9. Between November 24 and December 9, the geyser decreased 5 gpm in discharge, and spring 18 on the northern extension of the same fissures increased 6 gpm. This is an example of an exchange of function that Marler (1951) has described in Yellowstone Park.

Spring 20 (fig. 13) differs greatly from most other springs in its very dampened increase in discharge and especially its increase in chloride content in response to precipitation. Rather than diluting chloride content, precipitation evidently taps some auxiliary supply of chloride. The only local source available for such immediate response is chloride that has been stored in the area near and just west of the spring as a result of previous near-surface evaporation.

Other springs in the northwestern active part of the Main Terrace, including 20n, 19, and 21 (fig. 23), are generally anomalous, with chloride contents increasing in response to precipitation. Springs to the east and south normally decrease in chloride in response to pre-

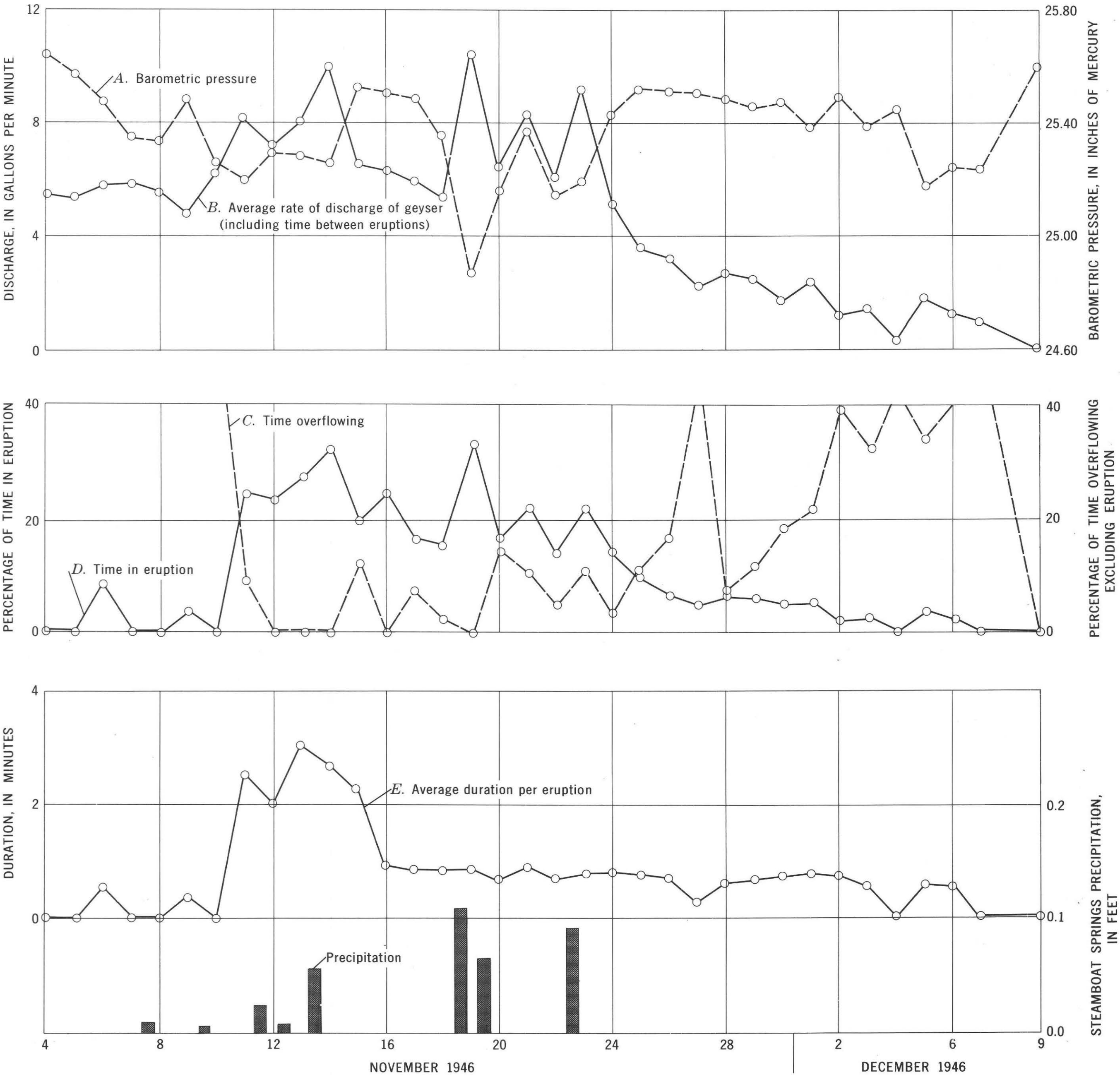


FIGURE 22.—Average discharge rate and other eruption characteristics of geyser 23n, Main Terrace, November and December 1946.

cipitation, as shown by springs 50, 18, and 5 (fig. 13) and spring 2 (fig. 23).

Evaporation of near-surface saline water in the area just west of the anomalous springs presumably concentrates salts at the surface. Rainfall on this immediate area dissolves the salts and sinks underground, flowing eastward below the surface to the fissures. Water levels are only a few inches to several feet below the surface west of the anomalous springs, as indicated by the fact that a seeping spring shown as 19nw on plate 3 became active in the spring of 1953, and saline crusts have been observed near here at other times.

Springs farther to the south are not influenced in the same way, perhaps because the surrounding sinter is relatively dense and impermeable, or because water levels are too far below the surface for extensive evaporation to occur. For example, depth of the water table in Rodeo well was normally about 8½ feet (table 5) and in drill hole, GS-3, about 8–10 feet. Probably little, if any, evaporation of thermal saline water occurs

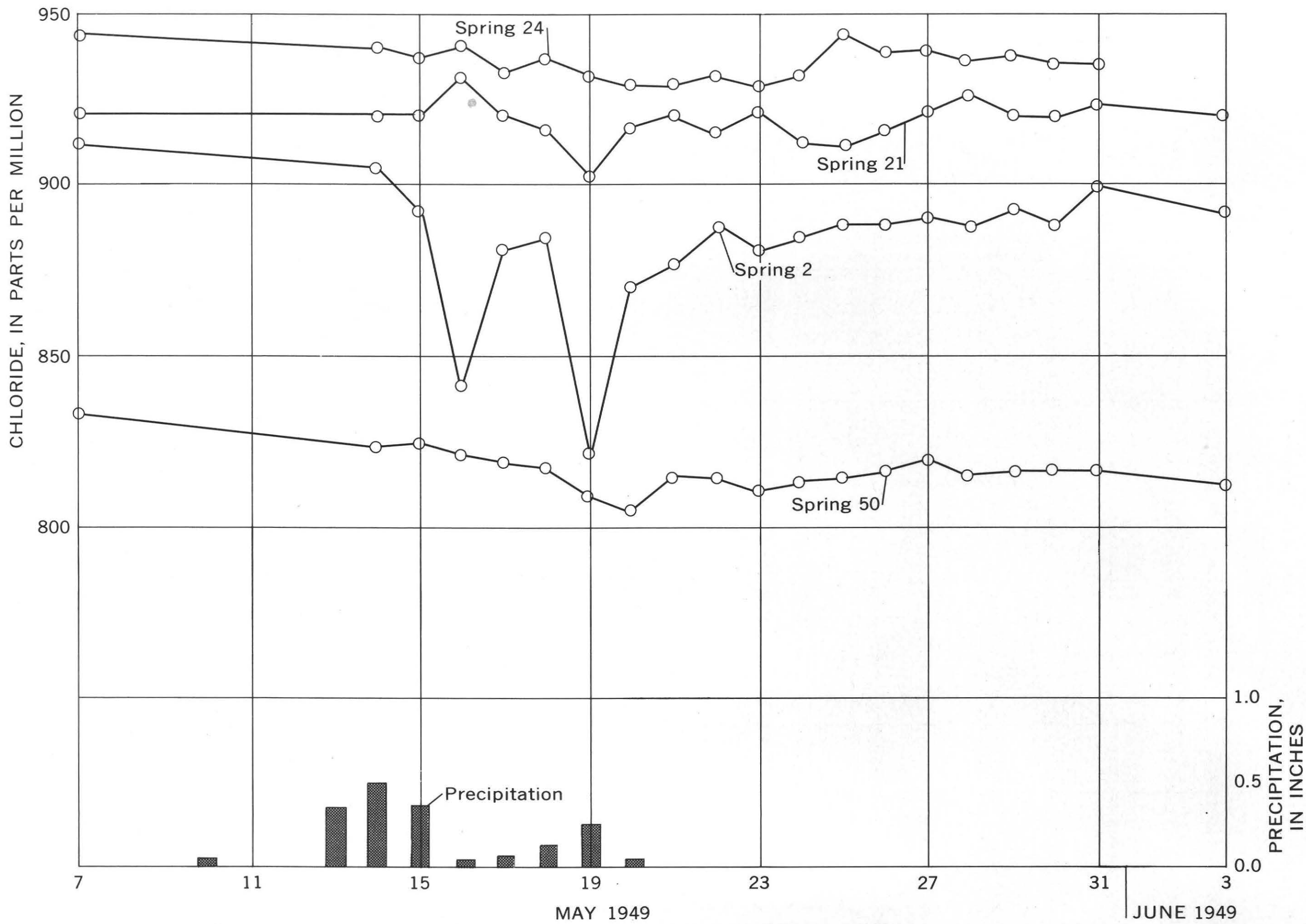


Figure 23.—Chloride concentrations of four springs and their differences in response to precipitation.

here, but the ground is porous enough to absorb nearly all precipitation that falls, mixing in the spring system either almost immediately or with a lag in time.

The springs that are diluted immediately by precipitation also show temperature decreases, but the springs that increase in chloride or lag in chloride response show little change in temperature.

INFLUENCE OF EARTH TIDES

An influence of earth tides was discovered at Steamboat Springs as a result of specific search, following a suggestion by T. W. Robinson (oral commun., 1946), of the Water Resources Division of the Geological Survey. Robinson pointed out that he had found earth-tidal influences in artesian wells that were responsive to barometric pressure (Robinson, 1939). When the water-stage records from vents high on the Main Terrace were corrected for barometric influence (figs. 10, 21), the residual curves clearly indicated fluctuations similar to those observed by Robinson. The semidiurnal nature of the major residual fluctuations, their gradual progression through the month, and the relation of shape of curve to phase of the moon proved that they were indeed earth tides. The maximum fluctuations were greatest, about 0.10 foot, near the time of new moon, were slightly less during full moon, but were only about 0.05 foot during the first and third quarters.

In contrast to the water-stage records from the Main Terrace, the pressure-corrected residual curves from vent 31n on the Low Terrace (fig. 11) contain no definite indication of earth tides (further discussed on p. C44). The minor fluctuations in these residual curves are at least in part related to differences between short-term long-term barometric response, whereas a constant barometric factor was applied in deriving each of the residual curves.

An earth-tidal influence on the flowing springs has not been demonstrated, but Main Terrace springs that respond most strongly to barometric pressure, such as springs 22 and 24 (table 4), almost certainly respond also to earth tides. In view of the fact that the maximum barometric response of nonflowing vents is equivalent to about 0.5 foot of water and earth-tidal influence, 0.1 foot of water, the response of a flowing spring to earth tides is likely to be at least 20 percent of its maximum response to barometric pressure. Diurnal barometric fluctuations (fig. 10, curves *A*, *B*) commonly have an amplitude of about 0.1 foot or a little more, and the average earth-tidal amplitude is about 0.07 foot. This is equivalent to 70 percent of the usual daily barometric fluctuation but only 15 percent of the maximum barometric fluctuation.

Until very recent years, earth tides had seldom been detected in fluctuations of water level. Klönne (1880) detected an influence of earth tides in fluctuations of water level in a flooded coal mine in Europe. Young (1913) described fluctuations in discharge from shallow wells in South Africa that correlate with earth tides. Four flowing wells from 65 to 200 feet deep ranged from 25½° to 33°C in temperature; their altitudes at point of discharge differed by as much as 16.7 feet. The highest altitude well was the only one that showed clear tidal fluctuations. Two others, 3.5 and 6.7 feet below the first, showed slight variations, and the fourth well, 16.7 feet below the first, was nearly constant in discharge. The highest well was also most responsive to barometric pressure (Young, 1913, pl. 4). These relationships to altitude are similar to those described at Steamboat Springs.

Robinson's recognition (1939) of earth-tidal influence on water levels in artesian wells has been mentioned. The maximum tidal fluctuation in brine wells in New Mexico (Robinson, 1939, p. 659–660) were a little less than the 0.10 foot found at Steamboat Springs. The pressure variations were about equal, considering differences in densities of the waters. The well in Iowa City, Iowa, showed fluctuations of as much as 0.18 foot, nearly double the maximum of the Main Terrace of Steamboat Springs.

Stewart (1961) described earth-tidal amplitudes as much as 0.25 foot in wells in crystalline rocks in Georgia, where barometric influence was found to be slight; Richardson (1956) published a brief description of tidal fluctuations in wells in Tennessee accompanied by strong effects of precipitation but with little or no barometric response. Melchior (1959) found both barometric and earth-tidal effects in wells in Belgium and in the Congo. Tidal fluctuations of 0.2–0.3 foot are common in wells in Missouri, and 0.1 foot in wells in Florida, according to J. D. Bredehoeft (written commun., 1966).

Theis (1939) ascribed tidal fluctuations in water level to elastic dilation and compression of water contained in the aquifer, correlating with bulging and recession of the earth's crust in response to lunar gravitational attraction. When the actual crustal tide is "high," the water level relative to a reference point at the surface of the fluctuating crust is "low."

Earth tides have also been studied by means of sensitive tiltmeters and gravimeters. Stetson (1944) estimated a displacement of the crust of the earth of about 2 feet in a vertical direction; the lag between maximum tidal force and maximum response is about 50 minutes.

Nishimura (1950) has made tiltmeter measurements at a number of localities in Japan, including Beppu hot springs, Kyushu. The tidal fluctuations at six points in the Beppu area showed little relation to each other. The strongest tilt had an angular rotation of about 0.03 second of arc, and the smallest was less than 0.005 second. The earth-tidal anomalies are peculiar to each individual point. Nishimura concluded that the differences are related to an active fault through the center of Beppu, the fault plane behaving as a "free boundary" (1950, p. 360). The tidal fluctuations at Beppu are caused by direct effects of moon and sun and by indirect effects from loading and unloading from ocean tides in the Bay of Beppu. The latter produce the largest effects. For this reason Nomito and Seno (1939, p. 416) found that tidal influences on flowing springs decrease with distance from the coast. Their study detected no direct earth-tidal effects on hot-spring discharge, although such effects may exist in certain sensitive springs.

The full significance of tidal fluctuation of water levels at Steamboat Springs is not clear. Perhaps all water levels related to confined aquifers respond to some degree to tidal stress, and those associated with unconfined water presumably do not. The existence of tidal effects at Steamboat Springs probably indicates the existence of extensive artesian confinement within the whole system. Although water circulating entirely within one fault plane or in a complex system of intersecting faults and fractures is very different in detail from the simple concept of an artesian aquifer, the principles of response to confinement still apply.

A complex system of faults and fractures, such as that of figure 4 (a possible model for Steamboat Springs), has a certain reservoir capacity in open spaces that can increase or decrease in volume in response to bulging and recession of the earth's crust by earth tides; an expansion of the reservoir would be accompanied by the lowering of water levels at the surface of the system and contraction would raise water levels. The parts of the system at the highest altitude and under the least confining pressure should be the most pressure sensitive and should therefore show the highest response; this supposition seems confirmed by the available data.

Of significance is the finding of earth-tidal influence on the Main Terrace and its absence from the Low Terrace despite barometric responses of nearly equal mag-

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nitude in the two areas. The relationships are probably explained by the nature of interconnections and restrictions of channels at depth within the whole system. The Low Terrace, as we have seen, has a low total discharge, despite the advantage it has in lower altitudes of outlet. If earth-tidal fluctuations in the Main Terrace were produced by volume change in the reservoir at depths shallower than interconnections with the Low Terrace subsystem, then we should see similar tidal responses in the Low Terrace. Such a response is not found. Therefore, the volume changes in the reservoir must take place at greater depths and response is transmitted principally to the Main Terrace. Looked at in another way, the maximum tidal fluctuations of 0.1 foot in the Main Terrace subsystem is equivalent to a pressure change of only 0.003 atmosphere. The hydrostatic pressure differential between the crests of the Main and Low Terraces is 44 feet of hot water, or about 1.2 atmosphere of pressure. Thus, the maximum tidal fluctuation on the Main Terrace changes the hydrostatic pressure at the point of diversion into the subsystem by less than four-thousandths of the existing differential. Such a small change probably accounts for the lack of tidal fluctuations in the Low Terrace.

INFLUENCE OF EARTHQUAKES

All local earthquakes from 1947 to 1952 of magnitude 3.0 or more that were recorded on University of Nevada seismographs have been listed by White, Thompson, and Sandberg (1964, table 5). Fifteen shocks that were either very strong or that evidently orginated near Steamboat Springs are shown in table 17. All water-stage and microbarograph records for the corresponding dates were examined for recognizable response to the shocks. Six earthquakes produced definite response, two may have produced slight response, and no effects were detected for seven. All six of the most influential earthquakes produced changes in the Main and Low Terraces, but two of these six did not influence the water level of South Steamboat well. Water in the lower part of this well, as mentioned previously, is considered to be entirely meteoric in origin, and is heated as it flows into the spring system. Three of these six shocks affected the local microbarograph records, presumably from air, surface, or ground waves; none resulted in detectable changes in discharge of producing wells of the Steamboat Resort.

The two separate shocks of September 7, 1947, probably originated much closer to Steamboat Springs than the computed distance of 32 miles shown in table 17. The University of Nevada had not yet installed its Sprengnether seismograph, and the epicenter was computed from distant California stations. According to Romney and Meeker (1948, p. 99), three small shocks were felt in Reno and Washoe Valley in addition to the two principal shocks. At the Steamboat Resort, however, John Czykowsky (oral commun., 1947) recognized seven distinct shocks. Only three in all were felt in Carson City, south of Steamboat Springs.

The shock of November 14, 1949, in contrast, had a reported epicenter only 2 miles from Steamboat Springs. Although its Richter magnitude was calculated at 3.5,

TABLE 17.—*Effects of strongest and nearest earthquakes on recording instruments at Steamboat Springs, Nev.*

Date	Hour (PST)	Distance, reported epicenter from Steamboat (miles)	Richter magnitude[1]	Vent 35, Main Terrace [2]	Spring 31n, Low Terrace [3]	South Steamboat well, south end of Low Terrace [4]	Rodeo well, Main Terrace [5]	Microbarograph, Steamboat Resort [6]
Sept. 7, 1947	21:52	32?	Moderate	Not operating	0.02-ft drop	Marked rise (record lost)		?<½-mm fluctuation.
Sept. 7, 1947	23:13	32? (See text.)	Moderate	do	Induced an eruption of No 32 Geyser well.	do		1-mm fluctuation.
Nov. 25, 1947	10:09	6	Moderate	0.02-ft drop	0.02-ft net rise	None		None.
Mar. 28, 1948	10:26	28	4.6	None or <0.02-ft change	None	Not operating freely		None.
Dec. 29, 1948	4:53	30	6.0	0.01-ft rise	0.22-ft fluctuation, 0.18-ft net rise.	0.02-ft drop, then 0.2-ft rise in 1 hr, 0.7-ft rise in 9 hr.		
July 18, 1949	7:31	22	4.5	0.02-ft drop?	0.03-ft total fluctuation	None		None.
Nov. 14, 1949	2:41	2?	3.5	None	None	None		None.
Dec. 7, 1949	10:44	26	3.8	None or <0.04 ft	None	None		None.
Oct. 25, 1950	19:18	17	4.2	None	None. No. 32 Geyser well had erupted 1½ hr earlier.	None	None	None.
Dec. 14, 1950	5:24	50	5.6	Record lost [7]	None. Geyser had erupted ½–1 hr earlier.	None or gradual rise 0.03 ft.	Record lost[7]	Record lost.[7]
Jan. 20, 1951	22:28	17	4.1	do [7]	Record lost [7]	None	do [7]	None.
Jan. 22, 1951	7:14	23	4.8	do [7]	do [7]	None	do [7]	?<½-mm fluctuation.
May 1, 1951	17:00	12	3.4	None	Just recovering from eruption of No. 32 Geyser well.	None	None	None.
June 29, 1951	18:54	4	3.3	Rise? cable broke	Not operating freely	None	None	None.
May 9, 1952	7:31	4	5.1	0.5-ft drop	0.1-ft net fluctuation, 0.06-ft rise.	Not operating freely		3-mm fluctuation.

[1] Earthquakes before Dec. 15, 1948, determined without data from Reno; reliability after that date greatly increased by a 3-component Sprengnether short-period seismograph installed at University of Nevada; for other data and references, see White, Thompson, and Sandberg (1964, table 5).
[2] On one of principal fissures on the Main Terrace, shown on pl. 3.
[3] Near and influenced by No. 32 Geyser well, shown on pl. 3.
[4] Contains warm meteoric water below depth of 160 ft; hot saline water in upper part is cased off.
[5] Does not intersect major fissures; any effect from earthquakes may be strongly dampened. Recorder destroyed by eruption of July 5, 1951 (see text).
[6] Any effects on microbarograph presumably related to air, surface, or ground waves.
[7] Most records from November 1950 through February 1951 were lost when equipment shack was destroyed by windstorm, night of Mar. 4–5, 1951.

this shock did not influence any Steamboat records. Its epicenter is very likely to have been much farther away from Steamboat Springs than indicated by its calculated distance.

In general, only the strongest or nearest earthquakes were detected in the instrumental records at the springs. Water levels in the principal fissures are most strongly influenced. Meteoric water migrating into the thermal system—as represented by the South Steamboat well—is less sensitive to influence by earthquakes, but a few shocks have produced measurable fluctuations.

There is no evidence that any fault of the Steamboat Springs system was seismically active from 1945 to 1952, and there is no geologic evidence to prove any post-Pleistocene displacement within the thermal area.

Some individual springs were affected at least temporarily by the earthquakes of September 7, 1947. The total measured discharge (pl. 4) increased from 28.2 gpm on September 1 to 34.9 gpm on the 8th. The discharge of most individual springs increased slightly, but spring 50 increased from 3.3 to 6.9 gpm; spring 25, which is only 300 feet southeast of No. 50, decreased slightly. This suggests that some discharge was diverted from spring 25 into spring 50, which is lower in altitude.

Less obvious delayed reactions can also occur. Figure 19 shows the discharge, temperature, and chloride relations of springs 2 and 5 in the fall of 1947. Very pronounced changes occurred in these two springs within 1–3 weeks after the strong local seismic shocks of September 7. These changes are probably not directly related to the earthquakes, but the channels of upflow were very probably affected in some way. One likely explanation involves an interconnection between the spring channels that had been filled by silica gel. This is a normal process of the spring system (Brannock and others, 1948; White and others, 1956). Periodically at times of unusually high discharge, much flocculated silica gel is flushed out in large quantities (fig. 9). As a result of seismic activity, water in the channels of either spring 2 or 5 may have been diverted in such a way that an interconnection between the springs was gradually cleared, eventually producing a change in discharge relationship. The changes in behavior shown in figure 19 are established facts; a likely explanation involves delayed reactions that were initiated in some way by the earthquakes.

The earthquake of May 9, 1952, produced notable changes in individual springs. The total measured discharge (pl. 4, curve *D*) decreased from 89.4 gpm on May 2 to 75.5 on May 16; spring 19 stopped flowing, spring 24 decreased from 26.9 to 13.1 gpm, and spring 50 increased from 2.8 to 8.0 gpm. Other springs changed only slightly. The weighted average chloride content (curve *A*) and weighted temperature (curve *D*) decreased notably between the two series of measurements.

Neither general nor detailed effects of earthquakes on the spring system can be predicted with certainty. One conclusion is that if an earthquake is near enough or strong enough, changes of some kind will occur. The overall effects are generally small for small disturbances and have not been particularly useful in analyzing the nature and characteristics of the spring system. If a major seismic shock occurs near a spring system, however, major changes can be expected. Marler (1964) has described the many changes that occurred in the springs and geysers of Yellowstone Park as results of the Hebgen Lake earthquake of 1959. One major factor, among others, clearly was distance from the epicenter.

GEOTHERMAL WELLS AND EFFECTS ON NATURAL VENTS

WELLS DRILLED BY PRIVATE INTERESTS

The first gothermal well was drilled at Steamboat Springs about 1920. Before this time the Steamboat Resort had depended on hot water piped from nearby natural springs to supply the hot baths and for other uses. As shown by plate 4, however, all springs of the area are highly variable in behavior. A well was drilled near the resort to obtain a more dependable supply. No record was kept of the results, but experience has demonstrated that drilled wells do provide a much more dependable supply than natural springs.

The first well at the present Reno Resort just north of the active springs was drilled about 1927, and since then, discharge from all commercial wells has probably ranged from 200 to 400 gpm of thermal water, the average being near 300 gpm. Two-thirds to three-fourths of the total discharge from wells was produced from the Reno Resort to provide an adequate source for rapid filling of a large swimming pool.

Measurements were obtained from all commercial wells drilled in the thermal area from 1945 to 1952. Special attention was given to temperatures, water compositions, nature of bedrock, and water levels as drilling progressed. The quality of the data differs greatly, depending upon the natural environment of each site and especially on the drilling method used and the speed of drilling. A rotary drill that circulates mud and is operated 24 hours per day can complete a well in minimum time, but very little information is obtained on the changing conditions with depth. A temperature log obtained after drilling is completed may have only a vague similarity at best to actual temperatures existing in the ground before drilling commenced. In contrast, drilling by cable-tool methods for 8 hours per day and 5 days per week provides much more opportunity to obtain meaningful data as drilling proceeds.

The most significant data obtained during the drilling of commercial wells are contained in tables 18–24, in order of location from south to north in the thermal area, and east to west.

Temperature data from Nevada Thermal Power Co. wells 4–6 were obtained from William W. Allen, who was supervising the drilling of these wells from 1960 to 1962. Data considered to be relatively reliable are shown in figures 24–28. Bottom-hole temperatures for these wells were obtained as drilling progressed, and the chloride content was determined at the indicated depths. Logs of cuttings from some of these wells have been described by White, Thompson, and Sandberg (1964, table 3).

In each figure that shows a significantly high range in temperature, a theoretical boiling-point curve for pure water is shown for reference. This reference curve is equivalent to curve A of figure 29, expanded in figure 30 to the depth scale generally used in this report. The data are contained in table 25 and assume a well filled to the ground surface with liquid water that is, at every point in the column, exactly at the boiling temperature for the existing pressure. The change in density from thermal expansion of water with temperature (fig. 30, curve C) is considered in the calculations.

TABLE 18.—*Data from South Steamboat well south of Low Terrace drilled by cable-tool rig*

[Measurements in parentheses are considered less reliable than others. Water samples from top of water column]

Date	Depth (feet)		Depth of casing (feet)	Bottom-hole temperature (° C)	Depth to water [2] (feet)	Temperature (° C, at water level)	Cl (ppm)	pH	Remarks
	Drilled [1]	Measured							
1946									
Dec. 6	6±	5		15.0					8 a.m.
7		14.2		27.2	7.3		320	7.44	7:20 a.m.
8		14.2		26.9	7.4				No drilling Dec. 8 and 9.
11		15.7	15.4	28.0	9.7		456		7:30 a.m.
12		15.9	16.5	27.5	6.5		540	8.14	8:30 a.m.
13	20	17.8	19.7	29.1	8.1		556	7.92	7:30 a.m.
14		19.2	20.6	29.4	7.5				8:30 a.m.
16	26	22.3	26	31.6	11.6		152	7.71	8 a.m.
18		31.6	29	37.7	12.4		480	8.47	8 a.m.
		32.8	29	(37.8)	(11.0)		(820)	(8.57)	5 p.m.
20		32.6	29	37.9	9.9				8 a.m.
					(7.5)				Outside of casing.
21		32.6	29	37.9	9.7		608	7.39	8 a.m.
26					9.1				No drilling since Dec. 21.
27	34	22.6	34	(31.1)	(+3.9)		(788)	(7.05)	Caving; not applicable to measured depth.
28		35.1	35	38.8	(28.0)		824	7.15	8 a.m.
29	43	41.2	40	42.0	10.7		592	7.26	8 a.m.
30		48.0	42	46.7	7.8				8 a.m.
1947									
Jan. 1		77.8	42	56.7	6.5		810	8.64	8 a.m.
2		96.8	42	61.1	5.2		840	8.55	8 a.m.
3	110	107.5	42	62.2	4.8		620	7.66	8 a.m.
5	119	114.3	42	62.7	4.5		854	6.74	8 a.m.
6		111.3	42	(61.7)	(4.0)				8 a.m.; caving.
10		105.6	115	(62.2)	(5.6)		(820)	(8.37)	8 a.m.; caving.
					5.2				Outside casing.
13			127		4.3		(852)	(7.12)	8 a.m.
15		122.2	127	63.7	9.0		872	8.12	8 a.m.
17		139.6	127	63.2	6.8				Sample bottle broken.
20	161	154.8	161	62.9	6.3		772	8.12	8 a.m.
23		161.3	163	62.0	10.4		616	7.19	8 a.m.
27		212.8	188	68.8	1.1		184		8 a.m.; high-chloride aquifer cased off, but some residual chloride still in well.
Feb. 3	275	267.7	188	74.3	+.1			7.48	8 a.m.
6					+.4				
10					+.5	43.3	12	7.45	Slight overflow.
17					+2.1	42.1	6.4	7.24	¼ gpm discharge.
19	275	257	188	74.0					

[1] Depth reported by driller. [2] Plus values, above ground surface.

TABLE 19.—*Data from Steamboat well 4, Steamboat Resort, Low Terrace*

[Measurements in parentheses are considered less reliable than the others. All water samples from top of column]

Date	Depth (feet)		Depth of casing (feet)	Temperature (° C)	Depth to water [2] (feet)	Temperature (° C, at water level)	Cl (ppm)	(pH)	Specific conductance (micromhos at 25° C)	Remarks
	Drilled [1]	Measured								
1947										
Sept. 21	------	19.2	------	82.8	13.5	------	550	7.7	2,500	
23	------	37.8	------	90.3	16.4	------	592	6.3	2,665	
25	------	31.1	------	(90.3)	(15.5)	------	(768)	(6.5)	(3,285)	
28	------	47.3	------	96.3	------	------	------	------	------	
30	------	46.4	------	96.3	6.7	------	826	6.4	3,275	
Oct. 5	------	54.2	------	96.7	18.5	------	872	6.5	3,370	
9	------	56.8	------	95.0	18.3	------	840	6.7	3,200	
12	------	68.0	------	98.1	13	95	960	7.8	3,645	Cl influenced by erupting Steamboat well 2(?).[2]
16	86	78	80	106.1	24	95½	842	7.9	3,010	
17	94	86.5	80	106.0	23	95	572	7.7	2,330	
18	------	104.5	80	(103.3)	18	(91½)	(88)	(7.3)	(675)	Cold water in to cool well.
22	------	107	93	103.9	19½	92	512	7.4	2,170	
23	------	111	93	108.9	19	92	600	7.2	2,580	
24	122	118	93	121.1	18	------	624	7.2	2,650	
26	------	131	93	141.9	21	93	628	7.2	2,740	
27	147	------	------	------	------	------	(624)	(8.5)	(2,545)	([3]).
1948										
Mar. 22	------	184	93	154.8	17	95	------	------	------	
24	225	------	------	------	------	------	------	------	------	Drilled and completed while well was erupting continuously; never able to get reliable bottom-hole temperature.

[1] Depth reported by driller.
[2] Steamboat well 2, only 6 ft to the northwest, was erupting nearly continuously during the drilling of well 4 and must have affected the measurements to some degree, in particular in depth to water level.
[3] Well 4 first erupted from depth of 142 ft; drilling continued through erupting column to 147 ft on Oct. 27; well was deepened to 225 ft in March 1948.

TABLE 20.—*Data from Rodeo well near crest of the Main Terrace* [1]

[Measurements in parentheses are considered less reliable than the others. Water samples from top of water column unless otherwise noted]

Date (1950)	Depth measured (feet)	Depth casing (feet)	Bottom temperature ° C	Depth to water (feet)	Temperature (°C, at water level)	Cl (ppm)	pH	Specific conductance (micromhos at 25°C)	Remarks
Feb. 9	1.0	------	17.4	------	------	------	------	------	Drilling, 10 a.m.[2]
	1.8	------	24.6	------	------	------	------	------	Drilling, 10:30 a.m.[2]
	13	------	------	------	60±	696	5.9	3,035	Drilling, 11:15 a.m.[2]
	23.7	19.5	------	(13.8)	75.6	872	6.7	3,495	Drilling, 1 p.m.[2]
	25.7	26.4	80	------	------	------	------	------	
	36.5	26.4	(93.4)	------	------	------	------	------	Drilling, 5:15 p.m.[2]
10	36.2	26.4	93.1	8.6	81.7	886	6.4	3,495	8 a.m.[3]
	46.7	32.4	(94.4)	(21.0)	------	892	6.4	3,560	Drilling, 2 p.m.[2]
	49.9	34.6	(92.2)	(10.8)	(80.3)	------	------	------	Drilling, 5 p.m.[2]
13	49.7	34.6	97.9	8.9	83.3	828	6.4	3,305	8 a.m.[3]
15	49.7	50.2	95.9	10.1	82.2	814	6.3	3,680	Removed and reset casing Feb. 13 and 14.
16	65.3	64.7	109.9	15.6	85.8	800	6.6	3,335	
17	81.9	84.1	(105.0)	(64.4)	(97.6)	792	(7.6)	3,160	Note low water 8 a.m.; vigorous boiling.
	96.0	88.7	(123.9)	(16.3)	------	------	------	------	Drilling, 5:30 p.m.[2]
18	89.0	88.7	124.3	13.6	83.4	------	------	------	9 a.m.[3]
20	86.3	88.7	124.3	12.0	84.3	848	6.7	4,960?	8 a.m.; in 5-ft mud.[3]
21	107.5	88.7	135.0	7.0	84.7	872	7.3	3,385	In 0.1-ft mud.[3]
23	116.9	88.7	136.3	6.9	85.9	876	7.5	3,400	8 a.m.[3]
	120±	88.7	------	------	------	(912)	(7.3)	(4,885)	Erupted red-brown water, 11 a.m.
	120±	88.7	------	8.3	------	888	6.7	4,875	Nonerupted, red.
24	120.1	88.7	138.9	7.3	86.7	864	6.8	4,025	8 a.m.[3]
25	131.5	88.7	143.6	7.6	97.8	862	7.3	3,805	8 a.m.; vigorous boiling.
27	131.5	88.7	144.3	7.4	97.9	894	7.6	3,615	8 a.m.; [3] vigorous boiling.
28	137.9	88.7	145.6	7.7	97.7	888	7.8	3,575	8 a.m.; vigorous boiling.

See footnotes at end of table.

TABLE 20.—*Data from Rodeo well near crest of the Main Terrace* [1]—Continued

Date (1950)	Depth measured (feet)	Depth casing (feet)	Bottom temperature °C	Depth to water[1] (feet)	Temperature (°C at water level)	Cl (ppm)	pH	Specific conductance (micromhos, 25°C)	Remarks
Mar. 1	146.6	88.7	146.9	7.0	98.2	932	8.2	3,590	8 a.m.; vigorous boiling.
2	--------	--------	--------	--------	--------	(920)	(8.9)	(3,540)	Water sample erupted from outside casing.
3	150.9	88.7	143.6	6.8	96.7	916	(9.6)	(3,580)	Casing cemented.
6	151.3	88.7	146.4	6.7	90.9	888	(8.0)	3,490	No drilling Mar. 4 and 5.
7	154.0	88.7	145.9	6.7	90.3	892	7.5	3,490	
8	--------	--------	--------	--------	--------	(964)	(9.0)	(3,630)	Erupted sample; eruption continuous Mar. 7 to noon Mar. 14 when eruption terminated. No new drilling.
15	164.3	88.7	126.1	7.9	91.0	832	7.1	3,195	
17	164.3	88.7	127.8	--------	--------	--------	--------	--------	Repeat.[3]
20	164.2	88.7	129.0	7.7	87.6	864	6.9	3,430	Repeat.[3]
21	164.1	88.7	130.1	7.6	87.9	864	6.8	3,400	Repeat.[3]
28	167.1	88.7	129.0	(16.0)	(77.1)	(128)	(11.5)	(1,340)	Rotary rig, circulating drill mud.
29	172.3	88.7	137.4	(17.8)	(81.9)	(124)	(11.4)	(1,150)	([4]).
30	177.2	88.7	138.3	(17.5)	(75.2)	(116)	--------	--------	([4]).
31	179.2	88.7	138.1	(12.8)	(72.1)	(130)	(11.6)	(2,010)	([4]). Drilled depth 180.1 ft.
Apr. 1	184.6	88.7	144.2	(14.8)	(74.6)	(100)	(11.4)	(1,350)	([4]).
2	190.0	88.7	147.4	(15.6)	(76.4)	(112)	(11.2)	(1,080)	([4]).
3	193.7	88.7	147.1	(16.0)	(76.6)	(110)	(11.0)	(971)	([4]).
6	200.8	88.7	152.3	(15.8)	(75.7)	(140)	(10.8)	(982)	([4]). Drilled depth 201.0 ft.
7	210.5	88.7	154.4	(15.9)	(74.4)	(140)	(10.5)	(921)	([4]). Drilled depth 211.0 ft.
8	221.1	88.7	157.8	(16.8)	(76.3)	(144)	(10.2)	(899)	([4]). Drilled depth 221 ft.
9	230.1	88.7	158.2	(14.2)	(76.1)	(140)	(10.8)	(1,080)	([4]). Drilled depth 230 ft.
10	235.4	88.7	155.4	(16.4)	(75.7)	(136)	(10.7)	(985)	([4]). Drilled depth 237 ft.
11	248.0	88.7	160.0	(17.7)	(77.2)	(124)	(10.4)	(855)	([4]). Drilled depth 248 ft.
14	260.3	88.7	160.3	(17.1)	(79.0)	--------	--------	--------	([4]). Drilled depth 261 ft.
15	271.1	88.7	163.4	(18.3)	(80.3)	(180)	(9.8)	(1,023)	([4]). Drilled depth 271 ft.
16	279.6	88.7	163.4	(17.7)	(80.3)	--------	--------	--------	([4]). Drilled depth 279.5 ft.
17	279.4	88.7	(125.3)	--------	--------	--------	--------	--------	2½ hr after circulation stopped. Drilled depth 282.5 ft; well completed.
18	279.3	88.7	159.3	8.3	71.2	(228)	(8.2)	(1,220)	20 hr after circulation stopped.
25	277.7	88.7	165.6	7.3	81.0	(356)	(7.2)	(1,630)	8 days after circulation stopped.
May 9	276.9	88.7	166.6	7.2	82.6	--------	--------	--------	22 days after circulation stopped.
20	276.9	88.7	166.8	7.2	84.1	(520)	(6.8)	(2,370)	33 days after circulation stopped; just before well erupted.
	--------	--------	--------	--------	--------	(924)	(8.5)	(3,600)	During eruption (induced by dry ice).
	--------	--------	--------	--------	--------	(520)	(7.1)	(2,410)	Side pipe during eruption, collected after previous sample; see text. Erupted to May 22.
22	281.5	88.7	(146.8)	(8.3)	(90.1)	(902)	(7.1)	(3,660)	6 hr after eruption ceased; bottom cleaned by eruption.
26	282.3	88.7	(160.3)	--------	--------	856	7.0	3,480	Pressure sampler, 5 ft below water level.
	--------	--------	--------	--------	--------	844	7.1	3,470	Pressure sampler, from 186 ft below water level.
June 2	282.3	88.7	(164.3)	(7.8)	85.2	832	6.7	3,440	
Aug. 17	282.3	88.7	168.8	7.6	86.4	836	7.0	3,440	

[1] For detailed log of well, see White, Thompson, and Sandberg (1964, table 3). Hole drilled by cable-tool rig to 164 ft, then completed by rotary rig, circulating mud.
[2] Measurement made during temporary cessation of drilling.
[3] No new drilling since previous measurement.
[4] Thermometer lowered through mud to or near drilled bottom immediately after drilling ceased and removed the following morning.

TABLE 21.—*Data from Harold Herz well 1, 4,000 feet northwest of the Main Terrace.*

[Measurements in parentheses considered less reliable than the others. Water samples from top of water column]

Date	Depth (feet)			Bottom-hole temperature (°C)	Depth to water (feet)	Temperature °C	Cl (ppm)	PH	Specific conductance (micromhos at 25°C)	Remarks
	Drilled [1]	Thermometer	Mud level							
1946										
Oct. 22	14.6	12.2	11.5	23.3						Drilling started Oct. 21.
		1.1								In dug hole 45 ft to southeast; used for shallow temperature.
		1.8		16.9						
	41	40.5	39.8	(37.4)						4:30 p.m., 1½ hr after drilling.
23	41	40.1	39.8	38.9						8:30 a.m.
24	52.5	51.1	50.2	44.9						8:15 a.m.
	58.3				55.8					5:30 p.m.
25	58.3	58.3	~57	48.6	55.9					10:25 a.m. Thermometer down since afternoon of previous day; water added on 25th.
26	58.3	57.0		47.8	(50.8)					
28	58.3	57.0		47.8	(52.7)					
30	58.3				53.6					
Nov. 1	58.3	56.9		47.7	53.9					
2	58.3				54.3					
3	58.3				54.5					Drilled on 3d; best figure for water table as measured in shallow open holes is 55–56 ft.
1947										
Feb. 10	(?)	74.0		56.4	55.0		144	7.8		
24	(?)	74.0		57.0	59.2	51.1	244	8.5?		Drilled Feb. 23; slight increase from deeper hole.
Mar. 10	(?)				58.6		(128)			Note Cl decrease; drill water probably added.
May 1	145	122.3		80.9	60.2		360	7.4	1,825	Drilled Apr. 27.
22	145	122.4		80.9	60.1		432	7.1	2,035	
30		145.1		89.7						Just bailed; drilling completed later on 30th.
June 5	155	133.5	133.5	84.9	60.2		388	7.1	1,830	At mud level, 6 days after drilling.
	155	140.0	133.5	87.8	60.2					In mud.
	155	154.5	133.5	93.2	60.2					Do.
	155	60	133.5	56.4	60.2					At water level.
	155	70	133.5	58.2	60.2					
	155	80	133.5	61.0	60.2					
	155	90	133.5	64.4	60.2					
	155	100	133.5	68.0	60.2					
	155	110	133.5	71.7	60.2					
	155	120	133.5	74.8	60.2					
	155	130	133.5	79.8	60.2					
July 23	155	132.7	132.7	81.8	60.1	56.7	356	7.1	1,705	At mud level, 54 days after drilled.
	155	133.3	132.7	83.0	60.1					Deepest probe.
	155	60.1	132.7	56.7	60.1					At water level.
	155	70	132.7	58.7	60.1					
	155	80	132.7	61.1	60.1					
	155	90	132.7	64.3	60.1					
	155	100	132.7	67.9	60.1					
	155	110	132.7	71.6	60.1					
	155	120	132.7	74.7	60.1					
	155	130	132.7	77.1	60.1					
1948										
Nov. 12	155	128.0	128.0	52.0	59.9	44.7	416	7.1	1,820	Cold water circulated in pipes for heating; note marked temperature decrease.

[1] Reported by driller.

TABLE 22.—*Temperature measurements in Nevada Thermal Power Co. well 4, west of Pine Basin*

[Data from William W. Allen, Supervisor. Drilled in 1960 by cable-tool rig. Measurements in parentheses are considered less reliable than the others]

Depth (feet)	Recorded temperature [1] (°C)	Depth (feet)	Recorded temperature [1] (°C)	Remarks
36	67	421	147	
85	81	433	149	
95	(127)	437	--------	Water level reported 217 ft below ground. (See text.)
100	97	445	149	
105	101	465	150	
106	96	495	152	
111	97	505	160	
120	106	520	--------	Water level reported 215 ft; water sample collected Sept. 8, 1960, by D. E. White contained in parts per million Na 212, K 111, Li 085, Cl 45, and B 3.0.
132	104			
134	107			
137	108			
141	108	535	166	
155	108	545	166	
176	(97)	570	177	
194	111	576	157	Tried to erupt with 25 lb dry ice; unsuccessful.
200	111	597	160	
212	112	604	168	
223	113	615	168	
241	116	622	171	
268	118	625	172	
282	124	634	172	
310	127	690	182	
321	129	700	177	
342	130	709	--------	Losing cuttings in crevice?
350	134	720	179	
371	138	726	186	
390	138	726	177	Well erupted Sept. 22, 1960, by airline to 315 ft, continuing to Oct. 17, 1960; erupted sample of Sept. 26 contained in parts per million: Na 660, K 65, Li 7.0, Cl 874, B 48.
		720	176	

[1] Drillers' measurements made by lowering maximum-recording mercury thermometer to bottom of hole, generally at start of morning shift 8 hr after previous drilling. Recorded to nearest degree Fahrenheit, here converted to nearest degree centigrade.

Liquid water under pressure exceeding that of the local atmosphere and heated to its boiling point at that pressure has much thermal energy that can be converted into steam when pressure is suddenly reduced to atmospheric. Formation of steam by sudden reduction of pressure on hot water is commonly known as "flashing." The pressure of flashing can be either atmospheric, as in a geyser or a freely erupting well, or above atmospheric pressure, as in the steam-water separator of a geothermal powerplant. Table 26 shows the enthalpy of liquid water at various temperatures from 100°–374°C and the energy exceeding that of liquid water at 100°C, available for formation of steam by flashing from higher initial temperatures. The weight percent of steam formed by flashing of water from temperatures determined from curve *A* of figure 30 is shown in curve *B*.

DRILL HOLES, U.S. GEOLOGICAL SURVEY

Diamond drilling for research purposes was done under contract for the Geological Survey from June 1950 to February 1951. The detailed logs of these holes were published by White, Thompson, and Sandberg (1964, table 3); the hydrothermal alterations have been studied by Sigvaldason and White (1961, 1962) and by Schoen and White (1966).

TABLE 23.—*Temperature measurements in Nevada Thermal Power Co. well 5 near clay quarry*

[Data from William W. Allen, Supervisor. Drilled in 1961 by cable-tool rig. Measurements in parentheses probably less reliable than the others]

Depth (feet)	Recorded temperature [1] (°C)	Remarks	Depth (feet)	Recorded temperature [1] (°C)	Remarks
53	60		378	148	"390 ft, soft, much mud."
76	60				
87	----------		397	151	
92	77	"Entering granodiorite"; (probably porous white acid-leached granodiorite at higher levels).	405	----------	"Hole trying to erupt."
			419–430	----------	"Soft, friable."
			482	(146)	"Water level, 282 ft." (See text.)
			513	157	
103	83		533	171	"Water level, 234 ft." (See text.)
120	93				
133	98		556	168	
157	100		572	171	"Mostly soft."
170	101		600	171	
186	103		618	171	
207	105		684	175	
228	108		709	175	"Hard granite."
238	(101)		741	169	"Thermometer replaced."
246	114	"Some water in hole."			
			782	169	"Fresh coarse biotite, 770 ft, and pea-sized pieces black lava."
272	116	"Some calcite 260 ft."			
291	124				
312	(122)				
325	128	"18 in. crevice at 317 ft."	801	169	
			826	163	"Completed July 15, 1961." Temperature series taken July 19 shown in fig. 28.
338	139	"Water level, 250 ft." (See text.)			
358	142				

[1] Drillers' measurements made by lowering maximum-recording mercury thermometer to bottom of hole, generally at start of morning shift, 8 hr after previous drilling. Recorded to nearest degree Fahrenheit, here converted to nearest degree centigrade.

TABLE 24.—*Temperature measurements in Nevada Thermal Power Co. well 6 in Pine Basin*

[Data from William W. Allen, Supervisor. Drilled in 1961 by cable-tool rig. Measurements in parentheses probably less reliable than the others]

Depth (feet)	Recorded temperature [1] (° C)	Remarks	Depth (feet)	Recorded temperature [1] (° C)	Remarks
20	71	"Opalized at 55 ft."	325	146	
61	86	"All opalized."	336	146	
72	90		355	150	
86	93	"Some granite at 86 ft."	370	152	
			399	157	"Calcite at 405 ft, water level, 115 ft."
94	104	"Making water at 95 ft."			
			420	157	
102	110		445	163	
118	114	"Harder at 117 ft."	472	171	Water added.
131	114		507	171	
142	116		551	169	
166	(104)	Set casing, may lack water.	576	169	"Soft gray mud 596 ft."
189	117		607	174	"Static well-head pressure 7 lbs."
205	121				
209	122		624	178	
237	127		634	177	
259	132	"Mud is reddish."	653	178	"Seven lbs well-head pressure."
264	129				
272	135		688	177	
286	(124)	Cold water introduced.	700	178	
			716	179	Well completed Sept. 11, 1961.
297	121				
311	146	"Water level, 111 ft."			

[1] Drillers' measurements made by lowering maximum-recording mercury thermometer to bottom of hole, generally at start of morning shift 8 hr after previous drilling. Recorded to nearest degree Fahrenheit, here converted to nearest degree centigrade.

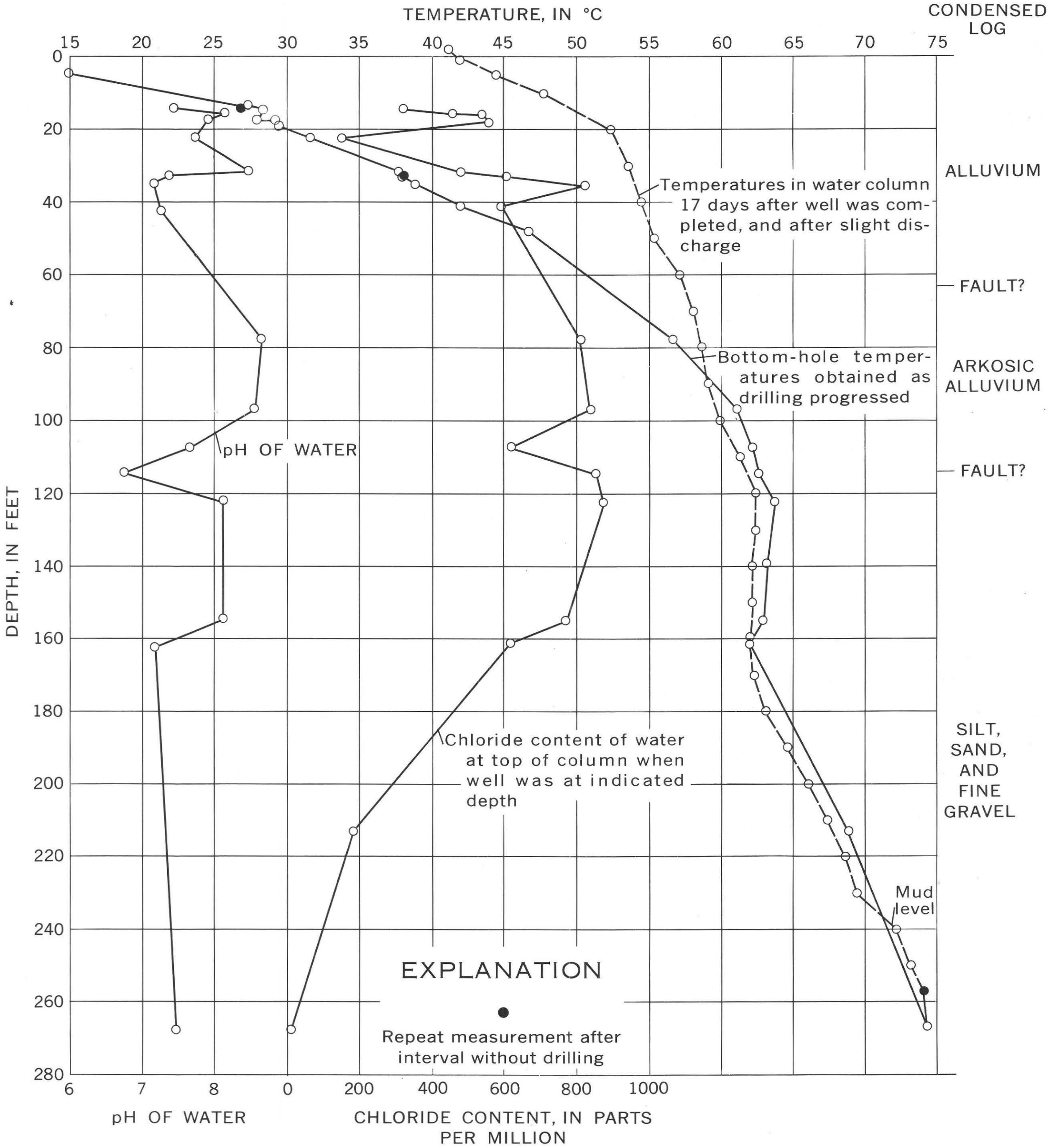

FIGURE 24.—Temperatures, pH, and chloride content of water in the South Steamboat well, south end of the thermal area. Drilled by cable-tool rig.

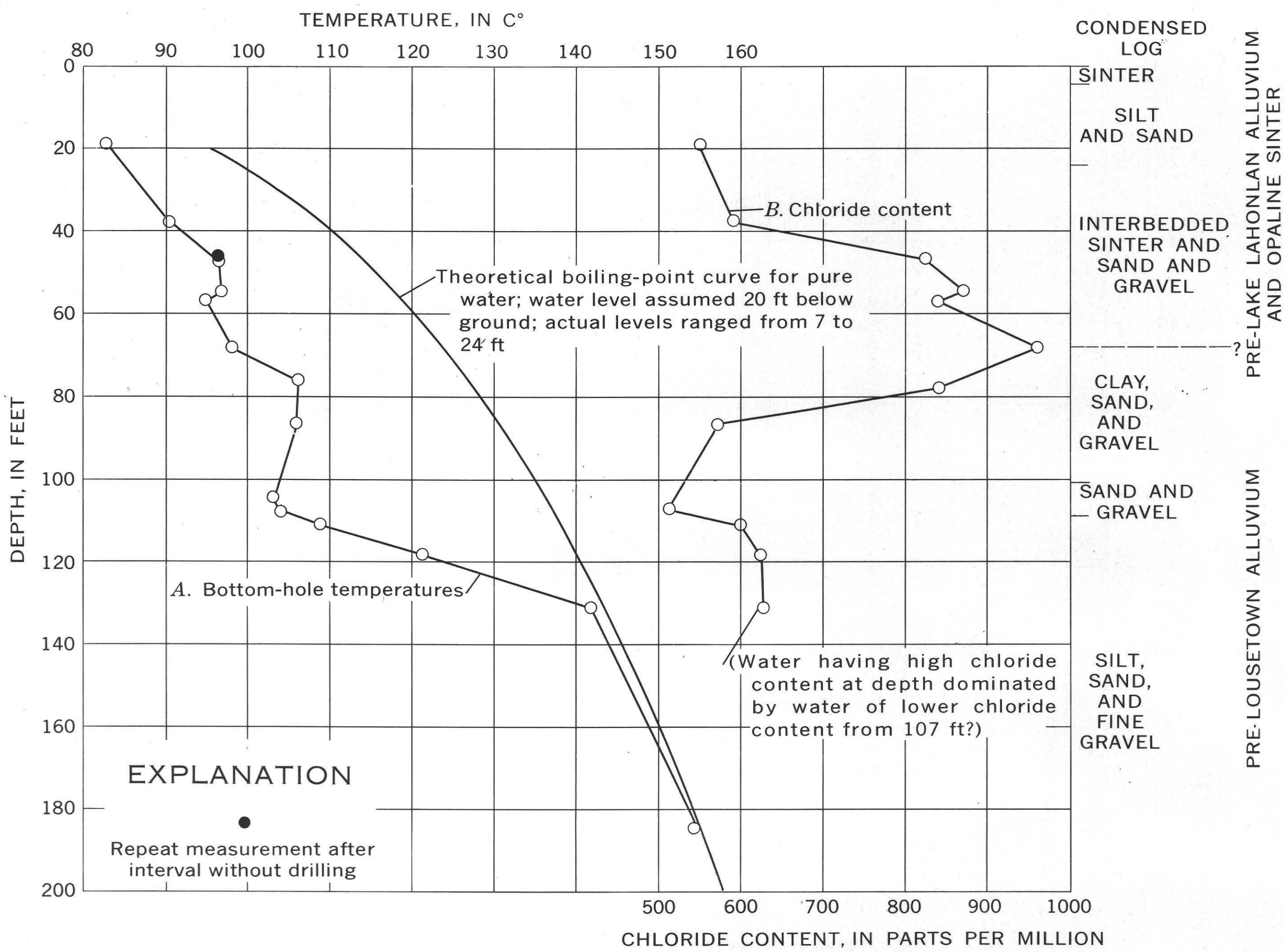

FIGURE 25.—Bottom-hole temperatures and chloride content of water in the Steamboat well 4, Steamboat Resort, Low Terrace. Drilled by cable-tool rig; 6-inch hole, cased to 93 feet.

Diamond drilling excels in yielding core for petrographic and chemical study. Data obtained from the eight diamond-drill holes are shown in tables 27–34 and figures 31–38. The data suggest that repeat measurements made over a number of weekends were normally within several degrees centigrade of the daily temperature measurements and the probable original ground temperatures prior to entry of the hole. Some repeat measurements, however, demonstrate that no universal rules of behavior are applicable.

Water samples obtained from diamond-drill holes are generally not reliable because of introduction of drill water. Where positive pressures existed, water was leaked off and sampled, especially from drill holes GS-1, 4, and 5. A deep-hole water sampler was constructed during the course of the study, but did not operate satisfactorily. Water samples, unless otherwise noted, were obtained from the top of the water column, but in the tables and figures of this report, the analytical data from these samples are referred to the drilled depth of the time of sampling. This obviously does not give a true picture of actual water compositions at these referred depths. Each sample is actually a product of mixing of waters from various depths, plus contamination from any water added from previous drilling and still remaining in the hole.

Table 35 is a summary of the most reliable chloride contents from diamond-drill holes and other wells drilled in the thermal area, listed by 100-foot intervals in depth.

GENERAL TEMPERATURE RELATIONSHIPS AND THE ERUPTION PROCESS

Experience gained from recent exploration for geothermal power has shown that a well penetrating a permeable aquifer where temperatures are close to boil-

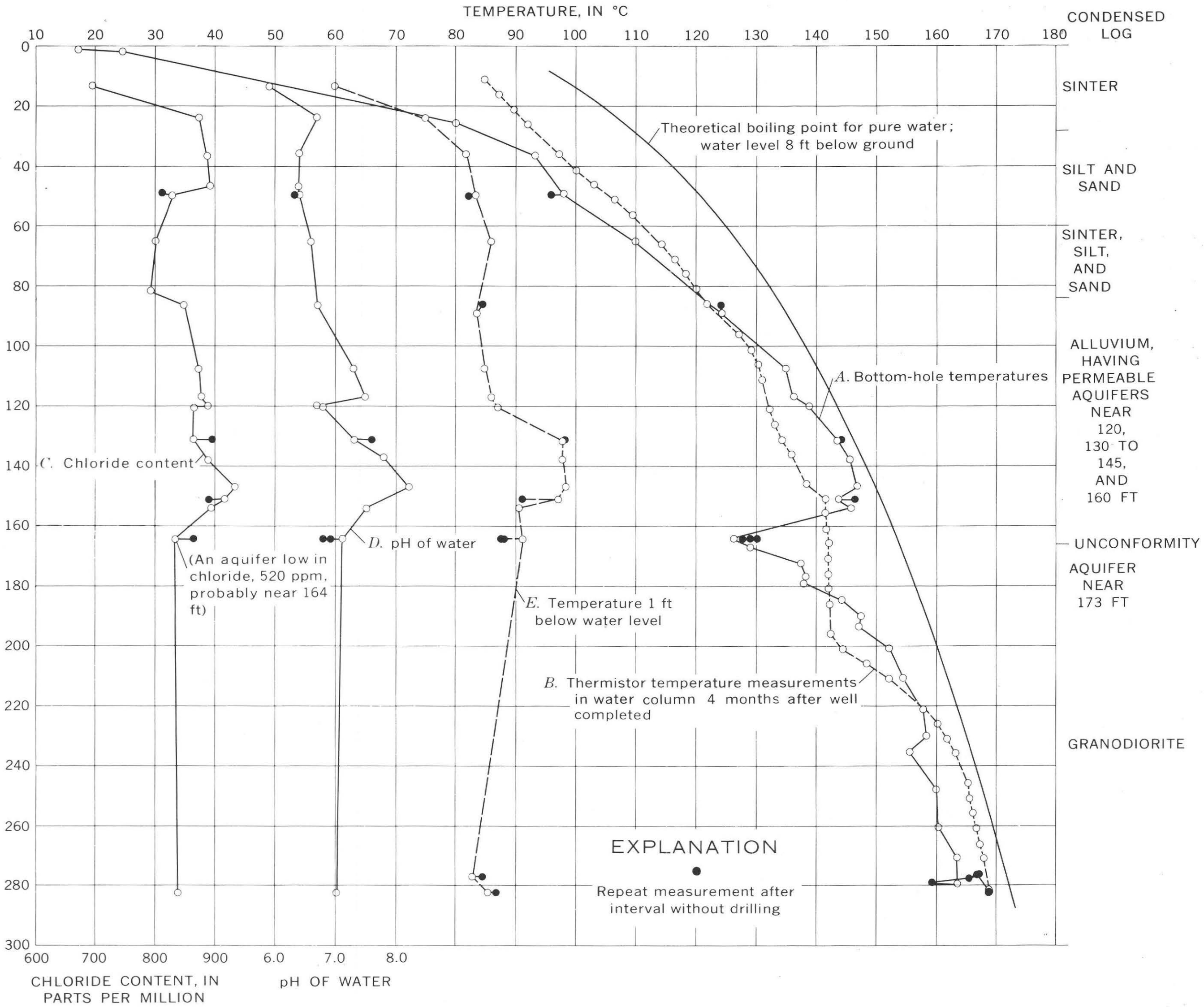

FIGURE 26.—Temperatures, pH, and chloride content of water in the Rodeo well near crest of the Main Terrace. Drilled by cable-tool rig to depth of 164 feet and completed by rotary rig, circulating mud; cased to 88.7 feet.

ing for prevailing depths and pressures can be made to erupt much like a geyser. Some shallow wells are true geysers in every sense, except that they are drilled wells rather than natural vents. No. 32 Geyser well on the Low Terrace (pl. 3), only 43.0 feet deep, is an outstanding example (White, 1967b). Other wells drilled to greater depths and higher temperatures may erupt continuously rather than periodically, once the process has started. Eruption from any of the wells at Steamboat Springs is continuous, except that chemical changes occur in the erupting water that result in deposition of calcium carbonate in the casing (p. C59–C62). As this occurs, the rate of discharge decreases, and eventually the well ceases to flow. In actual practice, when discharge has decreased to the minimum requirements of the establishment, the well is shut down and cleaned by drilling out the carbonate deposit.

Several relationships summarized above require more detailed consideration. Measurements in all wells and drill holes illustrated in figures 24–38 show that temperatures generally increase rapidly with depth; for the first few hundred feet below the water table, the temperatures of most wells are usually close to the reference boiling-point curve for pure water (figs. 29, 30, curve *A*). A well having such a temperature distribution is very unstable, because cooler relatively heavy water overlies deeper water that is lower in density, owing to its higher temperature (fig. 30, curve *C*). Water at temperatures above boiling for the atmospheric pressure of the area contains excess energy that will be

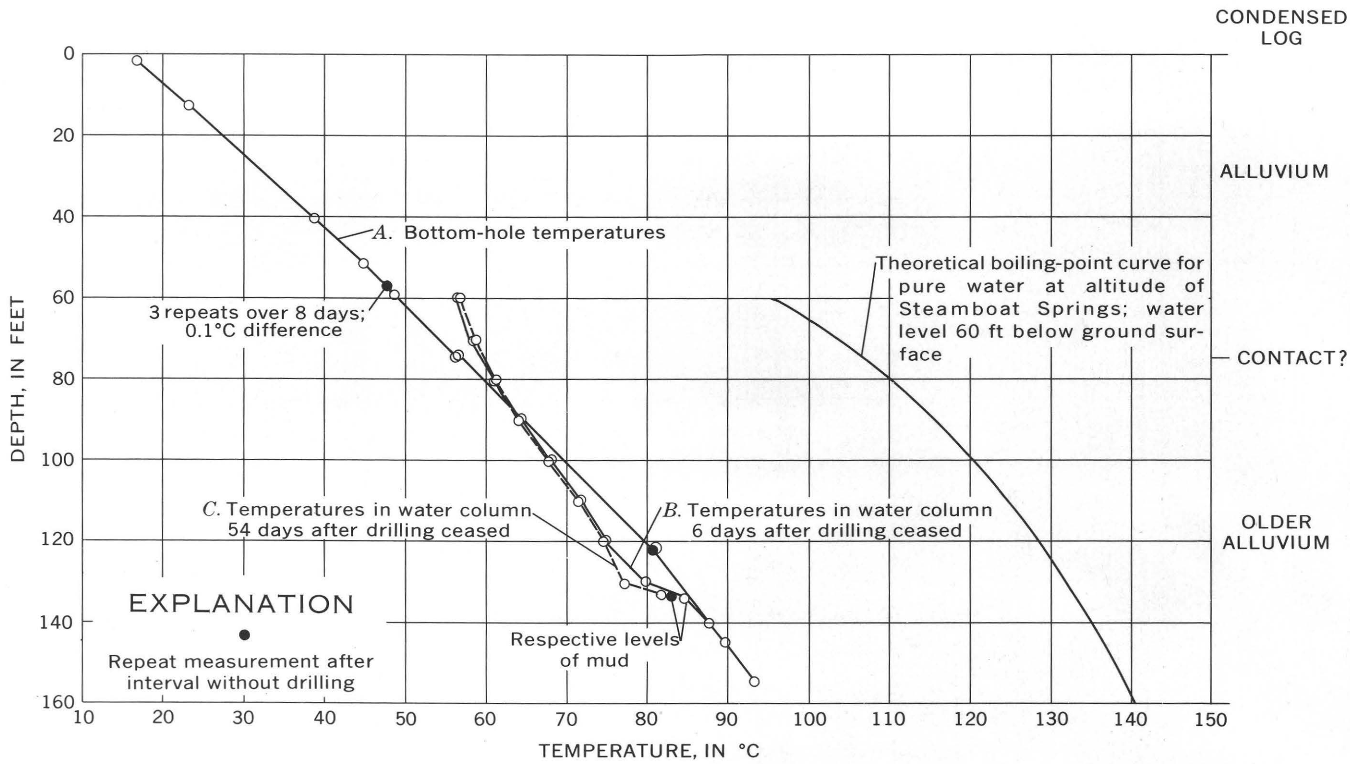

Figure 27.—Temperatures in Harold Herz well 1, 4,000 feet northwest of Main Terrace. Drilled by cable-tool rig, 6-inch hole uncased.

released to form steam if the pressure is lowered (see table 26 for decrease in pressure to 1 atm). If the hotter water at depth in a near-boiling column is displaced upward into a lower pressure environment, boiling starts to occur, and the column expands from the steam that is formed. In adiabatic cooling with no loss of heat by conduction, all excess heat is used in converting water to steam. If the water table is already at the ground surface, expansion of the column forces water at the top to overflow. If the water table is below the surface and the rate of boiling is sufficiently great, the column expands until water reaches the surface and is discharged. Hydrostatic pressure on the whole column then starts to decrease, rates of boiling increase further, and the well flashes into eruption. For a more detailed discussion of the above, see White (1967b).

Some wells erupt spontaneously when suitable aquifers are penetrated during drilling. This occurred in Steamboat well 4 (fig. 25) below 140 feet and in drill hole GS–1 (fig. 31) near 150 feet in depth. Other wells that are not sufficiently unstable do not erupt spontaneously. Convection and perhaps even some boiling may occur within the column, expanding it somewhat by steam that is formed, but not enough to discharge water at the surface and initiate an eruption.

If temperatures at depth are near enough to the boiling-point curve, a well can be induced into eruption in several different ways. One common practice in wells drilled by cable-tool rig is to withdraw the bailer rapidly after penetrating a zone that is to be tested. The rapid withdrawal of a bailer that is only slightly smaller in diameter than the casing lifts the upper part of the water column; hydrostatic pressures are then decreased at all depths below the bailer. Because of this reduction in pressure below the bailer, water formerly just below its boiling point suddenly starts to boil; as the bailer is raised, the column below the bailer expands from water vaporizing into steam. If the top of the column is lifted to the surface and overflows, the mixture of water and steam that is rising below the bailer also starts to discharge as the bailer is removed from the well and eruption commences. The previously existing steady-state conditions have been upset, and the vapor pressure in hot water flowing into the well is no longer balanced or exceeded by the weight of water in the column.

Curve *B* of figure 30 shows the percent of water that flashes into steam when water at each temperature (from curve *A*) is erupted to a pressure of 1.0 atmosphere. The data are from tables 2 and 4 of White (1955b), which also include ratios of steam to water.

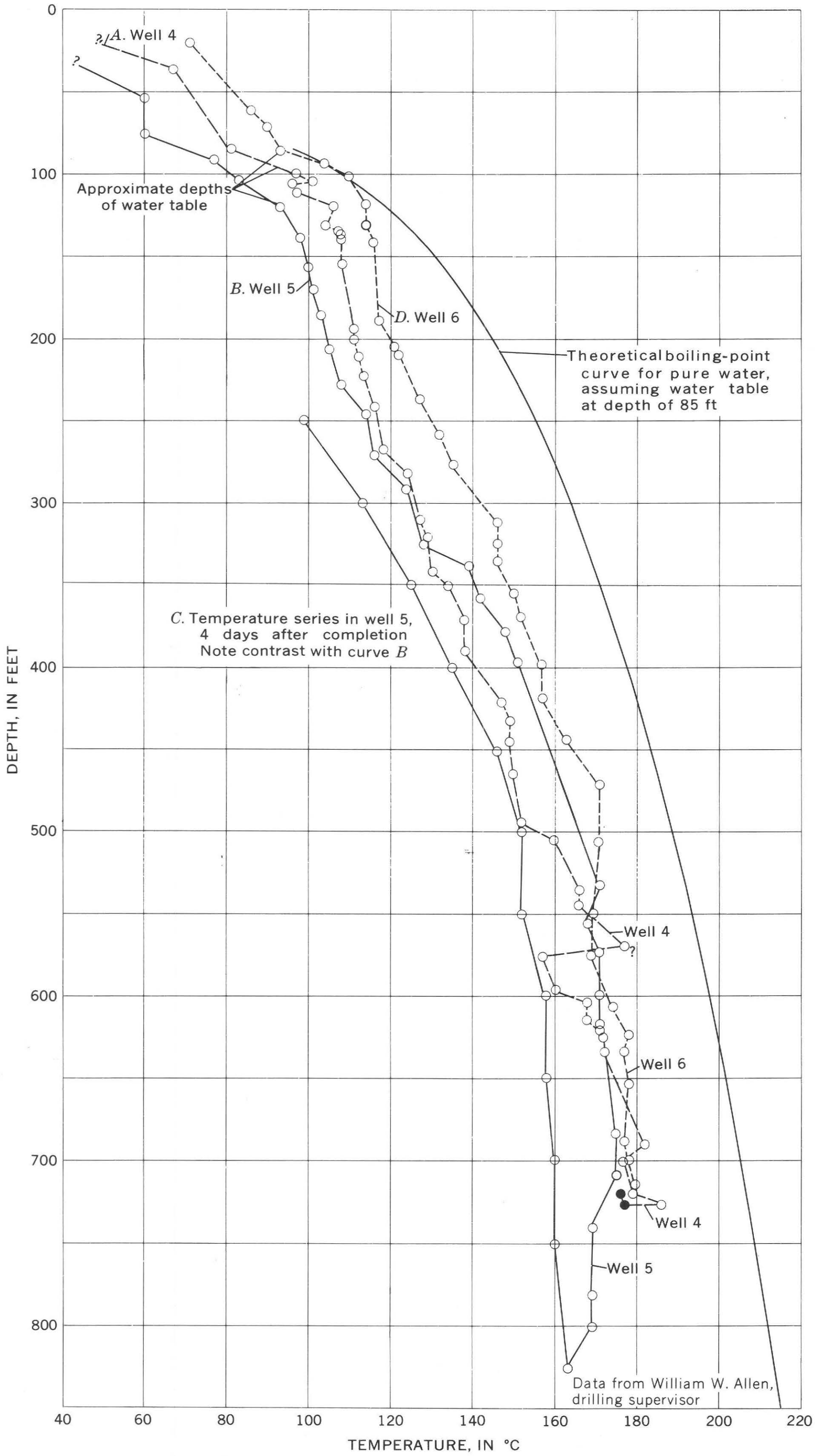

FIGURE 28.—Bottom-hole temperature of Nevada Thermal Power Co. wells 4–6 in western part of the thermal area.

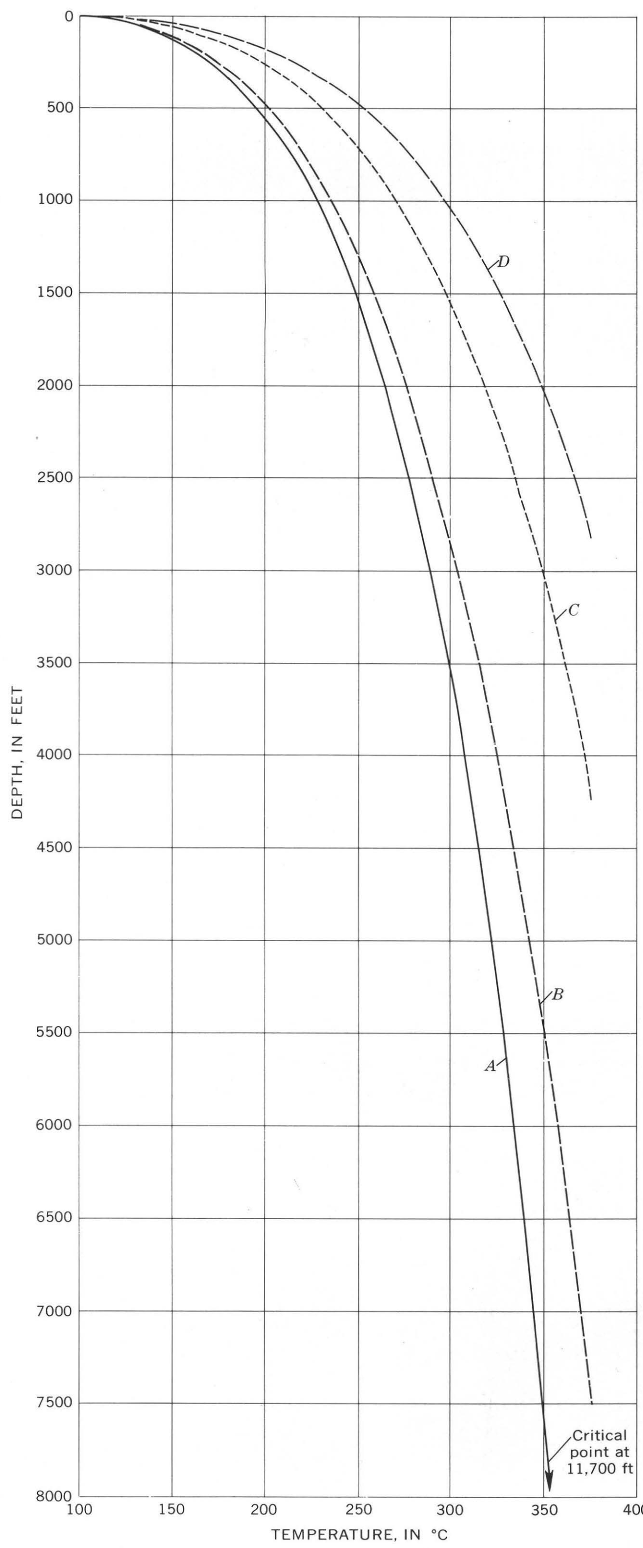

FIGURE 29.—Boiling-point curves for pure water. *A*, Hydrostatic pressure, water column at boiling point throughout; *B*, Water column at 15°C (except for calculated point); *C*, Lithostatic pressure, rocks of density 1.8; *D*, Lithostatic pressure, rocks of density 2.7.

TABLE 25.—*Boiling points and pressures in a column of pur water, every part of which is just at its boiling point*

[Calculated for sea level; corresponding values shown for Steamboat Springs, Nev.

Depth below water		Average temperature °C, for interval [1]	Average density for interval [1]	Pressure increment for interval (psi)	Total water pressure (psi)	Total pressure (psi)	Temperature °C, for water level at sea level [1]	Temperature °C, for Steamboat Springs [2]
Feet	Meters							
0	0	-------	-------	-------	-------	14.696	100.0	95.3
		103.6	0.9557	4.14				
10	3.05				4.14	18.84	107.1	103.5
		110.0	.9510	4.12				
20	6.10				8.26	22.96	113.0	110.0
		117.8	.9451	8.19				
40	12.2				16.45	31.15	122.5	120.2
		126.3	.9382	8.13				
60	18.3				24.58	39.28	130.1	128.2
		133.3	.9323	8.08				
80	24.4				32.66	47.36	136.5	134.8
		139.2	.9271	8.04				
100	30.5				40.70	55.40	142.0	140.5
		144.4	.9226	8.00				
120	36.5				48.70	63.40	146.8	145.6
		149.0	.9182	7.96				
140	42.7				56.66	71.36	151.2	150.1
		153.2	.9143	7.92				
160	48.8				64.58	79.28	155.2	154.2
		157.0	.9104	7.89				
180	54.9				72.47	87.17	158.9	157.9
		160.6	.9069	7.86				
200	61.0				80.33	95.03	162.3	161.4
		165.4	.9021	15.63				
240	73.2				95.96	110.66	168.5	167.6
		171.2	.8962	15.53				
280	85.4				111.49	126.19	173.9	173.2
		176.4	.8906	15.44				
320	97.5				126.93	141.63	178.9	178.2
		181.1	.8853	15.34				
360	109.8				142.27	156.97	183.4	182.7
		185.5	.8802	15.25				
400	122.0				157.52	172.22	187.5	186.9
		189.9	.8751	18.97				
450	137.2				176.49	191.19	192.3	191.7
		194.5	.8696	18.85				
500	152.4				195.34	210.04	196.6	196.1
		200.6	.8621	37.35				
600	182.9				232.7	247.4	204.5	204.1
		207.9	.853	37.0				
700	213.4				269.7	284.4	211.4	211.0
		214.5	.844	36.6				
800	243.8				306.3	321.0	217.6	217.2
		220.4	.836	36.2				
900	274.3				342.5	357.2	223.2	222.8
		225.7	.829	35.9				
1,000	304.8	-------	-------	-------	378.4	393.1	228.3	227.9

[1] Calculated from tables, Handbook of Chemistry and Physics, 44th ed., 1962–63 p. 2198, 2422–2425, by successive approximations. In the first 10-ft interval for exam ple, the problem is to determine the average temperature and density of boilin water within the interval and from this, the total pressure and boiling point at 10 f in depth.

[2] Calculated separately by same method as for sea-level data, but assuming alti tude of 4,600 ft equivalent to air pressure of 12.47 psi, 25.40 in. Hg, or 645 mm Hg The boiling-point curve for Steamboat could also be constructed by extending th sea-level curve upward to 4.93 ft for a new reference point. In a boiling column a Steamboat Springs, the total pressure at a depth of 4.93 ft below the water table i 14.7 psi, or 1 atm.

Eruption of geothermal wells has also been induce by placing an airline sufficiently below the water leve of a well with temperatures close to the boiling-poin curve. Compressed air (or some other gas) is ru through the line at sufficient rates to expand the wate column to the point of overflow. The cooler water on to of the column is discharged, the pressures throughou the column then decrease, and hotter water from dee aquifers flows in. If the inflowing hot water is suff ciently above the surface boiling temperature, boilin starts and then increases in the upper part of the col umn, eventually initiating an eruption.

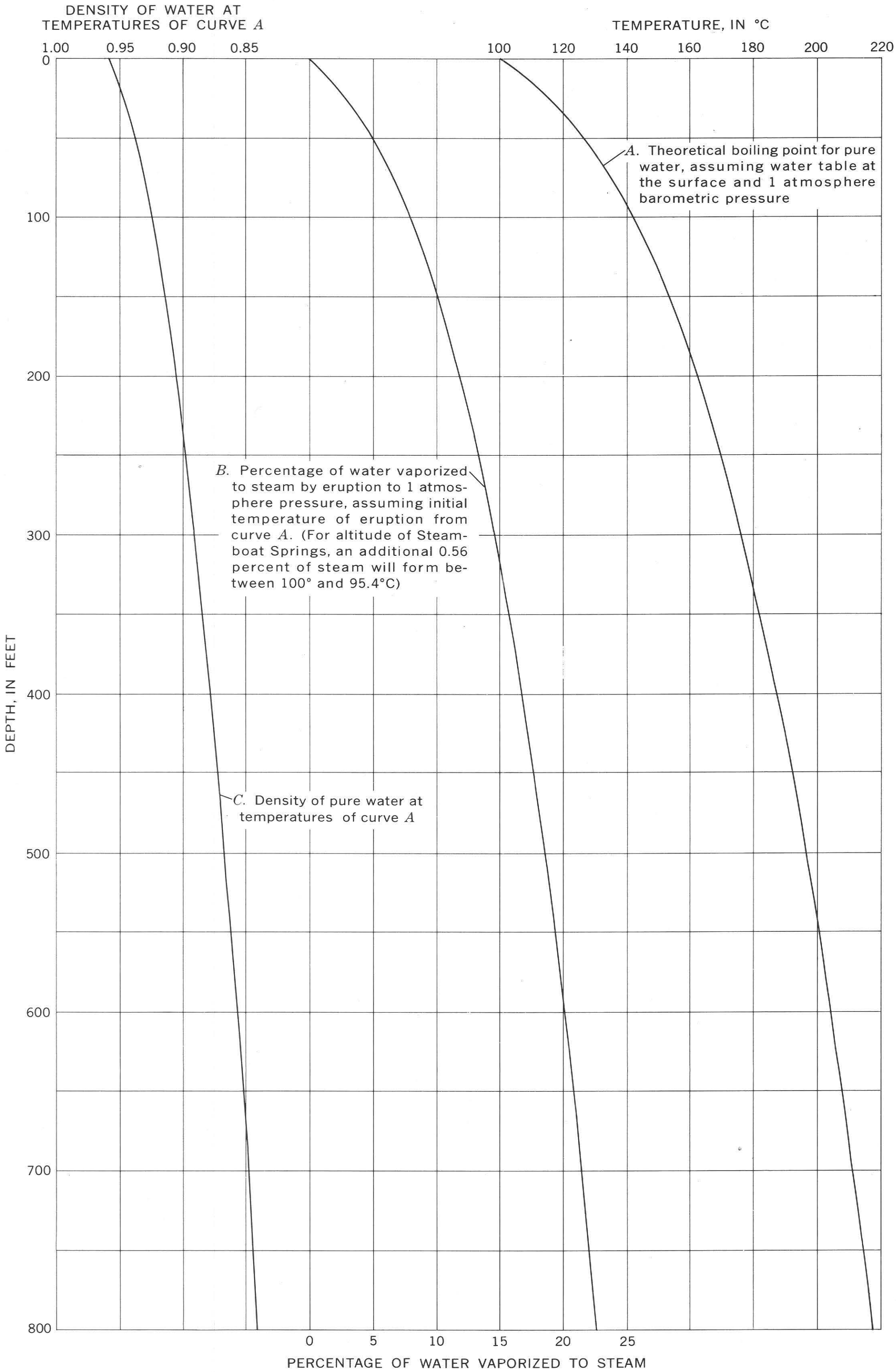

FIGURE 30.—Theoretical boiling-point curve, density of water at indicated temperatures, and percent of water vaporized to steam if erupted to atmospheric pressure.

TABLE 26.—*Energy and steam available in a thermal system with temperatures at depth and controlled by the boiling-point curve of a water column*[1]

[One pound of water from the specified depth or temperature assumed to form appropriate proportions of water and steam at 100° C and 1 atm pressure]

Depth (feet)	Temperature (°C)	Enthalpy [2] (Btu per pound)	Excess [3] (Btu per pound)	Steam [4] (weight percent of original water)
0	100.0	180.1	0	0
4.7	[5] 103.5	186.4	6.3	.65
25	115.6	208.3	28.2	2.91
50	126.5	228.3	48.2	4.97
100	142.8	256.8	76.7	7.91
134.1	150.0	271.7	91.6	9.44
200	162.3	294.6	114.5	11.82
300	176.4	321.2	141.1	14.54
500	196.6	359.7	179.6	18.5
541	200.0	366.3	186.2	19.2
1,520	250.0	467	287	29.6
3,570	300	578	398	41.1
7,630	350	718	538	55.5
11,520	374.0	903	723	74.6

[1] From White, 1955b, p. 1124–1125.

[2] Keenan and Keyes, 1936, p. 31–32. Btu=251.98 cal (mean)=2.930×10^{-4} kwhr=777.97 ft lbs=1054.8 joules=1.0548×10^{10} ergs 1 Btu per lb=0.5556 cal per g.

[3] Energy content in excess of that of liquid water at 100°C; available for formation of steam by flashing from indicated initial temperatures to 100°C and 1 atm pressure.

[4] From excess Btu relative to 970.3 Btu (heat necessary to convert 1 lb water to steam at 100°C and 1 atm pressure). This table is for general use for eruption to 1 atm pressure; for barometric pressure at Steamboat Springs, add 0.55 wt percent of steam that will form in decreasing pressure from 14.70 to 12.47 psi.

[5] Equivalent to maximum superheat recorded in natural thermal water at surface, 3.5°C in Giant Geyser (Allen and Day, 1935, p. 20).

An even more dramatic method of inducing an eruption, seemingly anomalous, is to drop chunks of dry ice (solid CO_2 with a density of 1.56 and a temperature of −78.5°C) into the well. The dry ice has a very pronounced cooling effect, but thermal energy from the hot water produces a rapid rate of evolution of CO_2 vapor. If the rate of evolution of CO_2 gas is too low, the gas bubbles rise through the column of hot water, with only slight expansion of the column. If evolution is fast enough, the column expands to the surface and water is discharged. Hydrostatic pressure throughout the column then decreases enough for water to boil vigorously below the dry ice; eruption then occurs.

Temperatures within a well may be very close to the boiling-point curve but continuing in a steady state of disequilibrium. The system may be so delicately balanced that an eruption can be induced by one of several seemingly insignificant changes. Eruptions of No. 32 Geyser well have been induced by throwing a small handful of sand or gravel into the water. To understand the results, we must realize that hot-spring waters are commonly supersaturated with CO_2 and other gases. These gases are in solution in water deep in the system where pressures are high. When this water rises to shallow depths where pressures are low, the total potential vapor pressure of the gases may considerably exceed the hydrostatic pressure; some "boiling" may occur, but at slow rates that do not attain equilibrium with the changing pressure. Sand thrown into the hot water has a cooling effect, but this is more than counterbalanced by providing surfaces, edges, and corners where nucleation of the supersaturated gases can occur rapidly. Existing temperatures may be a few degrees below the boiling points of pure water for prevailing pressures, but hot-spring water with dissolved gases has an effective "boiling range," [3] rather than a single boiling point for each pressure.

The water may be strongly supersaturated with gases when this low-temperature "boiling" first starts; gas bubbles may then form rapidly enough to expand the water column to the point of overflow, initiating an eruption. An eruption is obviously much easier to induce from an initial water table 1 inch below ground surface than it is from 1–5 feet or more below the surface.

The Rodeo well (fig. 26) and drill hole GS–3 (fig. 33) are examples of wells that did not erupt spontaneously but could be induced into eruption by one of the several methods described above. The South Steamboat well (fig. 24), Harold Herz well 1 (fig. 27), and drill hole GS–6 (fig. 36), on the other hand, are too low in temperature to erupt under any ordinary conditions.

WATER SUPPLY

In addition to suitable temperature relationships, an equally important requirement for continuous eruption is adequate reservoir permeability and water supply. Water must flow into the well from connecting channels or aquifers in sufficient quantity and temperature to maintain continuous eruption. Senges well, located midway between the Main and High Terraces (pl. 1), had a temperature profile very close to the boiling-point curve; it is an example of a well that is high enough in temperature but without a sufficient water supply to sustain continuous eruption. Several times before 1949 the well was induced into vigorous eruption by the dry-ice method; after a minute or so of eruption, the water originally contained in the well and in nearby interconnecting pore spaces had been expelled, and the eruption changed into a steam phase similar to that of many geysers. No quantitative data were obtained on the rate of discharge that is necessary to maintain continuous eruption, but the critical rate is a function of temperature, depth of lift above the water table (or the potential level from the producing aquifer), and diameter of the well. For a well 6 inches in diameter,

[3] At constant pressure, the solubility of a gas in water decreases with increasing temperature. Depending on the content of dissolved gases, a vapor phase starts to form at some temperature below the boiling point of pure water for the prevailing pressure. The vapor phase is enriched in gases, which thereby decrease in the remaining water. At constant pressure the "boiling range" spans the temperatures from first bubble formation to that of boiling of pure water or even higher for superheated (unstable) waters.

TABLE 27.—*Data from drill hole GS-1*

[Near the south end of the Low Terrace, Diamond drilled, circulating cold water. Measurements in parentheses are considered less reliable than the others]

Date	Depth drilled (feet)	Depth of thermometer (feet)	Bottom temperature (°C)	Depth to water [1] (feet)	Temperature (°C at water level)	Cl [2] (ppm)	pH [2]	Specific conductance (micromhos at 25°C) [2]	Remarks
1950									
June 23	29.5	27.1	57.7	4.3	37.8	(244)	6.8	(1,330)	Cased 9 ft, 5 in ½ hr after drilling ceased.
	68.5	67.7	(66.1)	----	----	----	----	----	
24	68.5	67.6	91.2	6.0	37.1	544	6.4	2,550	
25	103	100.8	112.3	5.6	37.2	588	6.3	2,640	
26	103	100.8	113.7	5.5	38.6	600	6.3	2,720	
29	133	62.3	(87.9)	(4.3)	(39.2)	----	----	----	Cased 47 ft (NX casing), hole cemented June 28.
30	135	128.6	122.2	1.5	33.9	(4)	(11.2)	(810)	Water sample still mostly drill water.
July 1	150	136.8	130.0	1.1	36.6	(120)	(10.0)	(750)	
2	185	127.5	(114.7)	.2	35.0	(872)	(11.8)	(6,220)	Hole cemented July 1.
3	185	127.4	(121.0)	+0.5	33.0	532	(10.6)	(2,700)	No new drilling.
6	190	148.1	154.8	+14±	60.0	818	(9.8)	3,250	Caving in hole; note excess pressure.
7	195	193.2	158.1	1.7	73.3	(75)	(7.5)	(610)	47 ft NX casing, 195 ft BX casing.
8	230	222.4	155.6	+20±	----	808	6.4	3,270	Note temperature reversal; may not have recorded minimum, the thermometer not insulated.
9	240	223.5	(154±)	----	----	676	6.6	2,910	Cased 215 ft BX.
10	245	241.2	(154±)	+18±	----	596	6.2	2,640	Cased 220 ft BX.
10	----	----	----	----	----	(672)	(6.6)	(2,950)	New spring 6 ft to north.
11	275	273.0	(154±)	+18±	----	816	6.4	3,510	Note probable temperature reversal 230–340 ft but not recorded by uninsulated thermometer.
12	305	302.5	(154±)	+17±	----	768	6.4	3,530	
13	340	337.8	(154±)	+47±	----	770	7.7	3,070	
14	365	365.0	154	----	----	----	----	----	Thermometer insulated.
14	375	375.0	(155±)	----	----	----	----	----	1 hr after drilling.
July 16	399.5	399	156.6	+48±	----	----	----	----	Thermometer insulated.
Sept 12	399.5	397.9	156.7	+52±	----	810	6.5	3,240	Do.
12	----	----	----	----	----	(20)	(5.6)	(324)	Water condensed(?) in compartment above valve.
19	----	----	----	----	----	(860)	(7.9)	(3,330)	Water between casings 47–220 ft, erupting.
1951									
Mar. 27	----	----	----	[1] 64 lb	----	----	----	----	
1952									
May 15	399.5	398.4	156.9	----	----	816	6.1	3,310	Water below 220 ft; sample warm.
15	----	----	----	----	----	(824)	(5.9)	(3,240)	Between casings, 47–220 ft.
Mar. 30	----	----	----	[1] 68 lb	----	840	6.4	3,320	Sample drawn off hot.

[1] Plus values are above ground level, generally indicated by pressure gage readings; final pressure readings (lb P) in pounds per square inch.

[2] Water samples obtained from top of column; when water level was above value well was allowed to discharge as long as feasible before sample was collected.

the critical rate of discharge is on the order of 5–50 gpm, depending on the lift required.

Geothermal exploration wells were drilled in the western part of the thermal area (pl. 1 and fig. 28). Several of these wells sustained impressive rates of production for several weeks or months. Nevada Thermal Power Co. 4, for example, erupted continuously at rates of discharge estimated at more than 200 gpm for at least 2 weeks. Discharge eventually decreased greatly because of deposition of $CaCO_3$ in the pipes and also, it is now clear, because of decreasing water supply. Granodiorite constitutes the bedrock throughout this western area. The local reservoir capacity consists of open spaces in fractures and zones of brecciated granodiorite. Permanent discharge of 200 gpm could be maintained only if the system could adjust to accelerated withdrawal by increasing the rate of upflow to 200 gpm, but channel permeabilities were evidently too low to permit this adjustment. The local reservoir became exhausted, water levels fell, temperatures and vapor pressures decreased, and the depth of lift increased until eruption could no longer be sustained.

DEPOSITION OF CALCIUM CARBONATE

Chemical changes induced by loss of CO_2 to the vapor phase result in deposition of calcium carbonate in pipes and casings. In the eruption process, hot water moves rapidly from a region of relatively high temperature and pressure up to the surface, where pressure is close to 1 atmosphere. Curve *B* of figure 30 indicates that if water at 170°C is erupted rapidly to the surface at sea level where the boiling point is 100°C, 13.4 percent of the original water will be converted to steam, assuming no loss of heat in other ways. At Steamboat Springs where

TABLE 28.—*Data from drill hole GS-2*

[Crest of High Terrace. Diamond drilled, circulating cold water. Measurements in parentheses considered less reliable]

Date	Depth drilled (feet)	Depth of thermometer (feet)	Bottom temperature (°C)	Depth to water (feet)	Temperature (°C at water level)	Cl [1] (ppm)	pH [1]	Specific conductance (micromhos at 25°C) [1]	Remarks
1950									
July 24	7	6.8	36.9	--------	--------	--------	--------	--------	Drilled July 23.
25	20	19.5	42.1	--------	--------	--------	--------	--------	3 ft NX casing.
26	27	19.9	(37.5)	--------	--------	--------	--------	--------	Aquajel in hole.
27	31	25.8	(37.9)	--------	--------	--------	--------	--------	Cemented July 26.
28	42	41.3	71.4	(37.1)	--------	--------	--------	--------	
29	52	6.7	(32.2)	--------	--------	--------	--------	--------	Cemented July 28.
30	60	58.9	99.0	39.1	--------	--------	--------	--------	
31	70	61.6	(89.4)	41.0	--------	--------	--------	--------	
	87	86.4	(87.5)	(39.3)	--------	--------	--------	--------	4 p.m., 15 min after drilling ceased.
Aug. 1	87	86.3	114.1	41.3	78.6	(4)	(9.8)	(497)	Drill water affected by cement near 50 ft?
2	111	110.4	111.4?	44.8	81.9	(2)	(6.3)	(575)	
4	132	130.9	126.7	41.6	75.6	(4)	(7.1)	(499)	Cemented to 129, Aug. 2, drilled out Aug. 3.
5	148	147.6	115.3?	50.4	80.6	(2)	(6.5)	(676)	Near fissure, cooled by drill water, no water return.
6	164	163.7	118.3	50.9	81.9	(2)	(6.2)	(703)	Do.
7	164	163.7	129.7	50.3	86.6	(2)	(6.2)	(702)	Repeat measurement; no new drilling.
8	180	179.5	121.7	51.1	86.1	(2)	(6.0)	(651)	
9	203	202.1	144.6	51.7	86.8	(8)	(7.9)	(225)	185 ft BX casing in Aug. 8.
10	227	225.8	145.0	51.6	87.8	(8)	(6.9)	(305)	
11	255	254.2	141.0	51.5	88.3	(10)	(6.8)	(263)	Excessive cooling from loss of drill water?
12	280	278.4	159.1	51.6	88.7	(6)	(6.9)	(234)	
13	305	303.2	160.4	51.3	89.7	(8)	(6.8)	(201)	
14	305	303.1	161.4	51.2	92.2	(5)	(6.9)	(216)	Repeat measurement; no new drilling.
15	326	323.4	150.6	51.5	91.2	(8)	(6.8)	(198)	Excessive cooling from loss of drill water?
16	356	354.4	154.7	51.5	91.6	(6)	(6.9)	(166)	
17	376	374.9	157.5	51.3	92.6	(8)	(6.8)	(171)	
18	398	396.2	152.2	51.3	95.4	(8)	(5.9)	(541)	Hole completed, casing pulled.
19	398	396.1	152.7	--------	--------	--------	--------	--------	Repeat measurement; no new drilling.
23	398	396.0	154.4	48.3	--------	(24)	(6.7)	(574)	Repeat measurement; no new drilling. Thermistor series. (See fig. 32.)
Sept. 20	398	--------	--------	--------	--------	(856)	(8.3)	(3,370)	Erupted sample.
21	398	--------	--------	--------	--------	564	6.0	2,730	Nonerupted sample.
1951									
Mar. 27	398	--------	--------	+12–16 ft [2]	--------	--------	--------	--------	
1952									
May 16	398	85.8	114.7	+3–7 ft [2]	--------	(904)	(8.8)	(3,380)	Erupted sample. $CaCO_3$ deposited below 86 ft.

[1] Water samples are clearly from introduced drill water before Sept. 20, 1950; all are from top of water column.
[2] Pressure-gage reading, equivalent feet of water.

the average boiling point is 95.4°C, an additional 0.56 percent of steam forms, for a total of 14.0 percent. As boiling occurs, CO_2 is relatively enriched in the vapor phase because it is very volatile and its solubility in near-boiling water is low. The pH of the remaining liquid water is generally considerably higher than that of water reaching the surface by normal upflow. At Steamboat Springs, the pH's of erupted samples are generally from 8.5 to 8.9 (White and others, 1953, p. 496–498; Ellis, 1959, 1962, p. 439–444), in contrast to the natural springs, which have pH's of 6.0–8.2. The highest pH's in the natural springs are always found in vigorous spouters and geysers. As Co_2 is vaporized from the water, the equilibria shift, and much bicarbonate is converted to carbonate. As a result, the solubility of calcium carbonate is then exceeded. Calcite is deposited in most erupting wells of the Steamboat area, but aragonite is the dominant carbonate of the West Reno well.

The rate of deposition is not constant from well to well, nor within a single well with depth or time. Most carbonate is deposited in the upper 50–100 feet of the well casing and in discharge pipes leading to the storage tanks. The detailed distribution of carbonate within the pipes is influenced by irregularities that presumably localize turbulence and differences in pressure. The effective diameter of the pipes just above and below the main valves of the Steamboat wells is ordinarily re-

TABLE 29.—*Data from drill hole GS-3*

[West of crest of the Main Terrace (4 E., traverse 3). Diamond drilled, circulating cold water. Measurements in parentheses are considered less reliable. Water samples from top of column unless otherwise noted]

Date	Depth drilled (feet)	Depth thermometer (feet)	Bottom temperature (°C)	Depth to water (feet)	Temperature (°C at water level)	Cl (ppm)	pH	Specific conductance (micromhos, 25°C)	Remarks
1950									
Aug. 20	45	39.0	83.8	10.1	49.8	(60)	(6.6)	(637)	Drilled Aug. 19; 16 ft of 5-in casing.
21	79	78.2	112.8	8.3	49.7	(16)	(6.3)	(580)	55 ft NX casing.
22	125	123.5	127.4	8.7	54.0	(12)	(6.3)	(837)	Drilled with fresh water until Oct. 31.
23	174	172.6	147.8	8.6	52.7	(44)	(7.6)	(597)	
24	223	220.5	156.3	8.9	53.2	(40)	(6.2)	(847)	
25	258	254.3	158.0	8.9	51.2	(140)	(6.1)	(1,034)	
26	294	292.2	160.1	9.0	52.8	(40)	(6.0)	(693)	
27	300	299.4	165.6	8.6	53.7	(92)	(6.2)	(1,045)	
28	300	299.4	165.7	----	----	----	----	----	
Sept. 20	300	----	----	7.2	66.3	(608)	(6.2)	(2,630)	
Oct. 16	300	----	----	Erupts	----	(796)	(6.1)	(3,170)	Erupts when open. Sample from valve.
21	300	296.4	165.9	6.3	51.1	776	6.1	3,090	Sampled under slight pressure; first reliable sample.
24	334	332.6	158.1	8.6	56.8	(232)	(6.1)	(1,350)	Drilled Oct. 23, first in GS-3 since Aug. 26.
25	369	367.8	164.4	11.6	58.3	648	6.2	2,770	
26	415	413.9	159.1	13.4	58.6	680	6.2	2,845	
27	435	431.6	166.1	13.0	61.9	720	6.2	2,970	
28	473	472.5	168.6	13.1	58.3	680	6.1	2,850	
29	473	472.4	169.1	12.2	63.9	760	6.2	3,125	Repeat measurement.
30	473	472.3	169.2	12.0	65.6	756	6.3	3,100	Repeat; see thermistor measurement, fig. 33.
						(908)	(7.0)	(3,545)	Composite sample from springs 3–5, pumped for use as drill water.
Nov. 1	515	512.1	168.2	12.8	63.7	744	6.2	3,085	Drilled Oct. 31, using spring water through completion of hole.
2	533	529.9	169.1	12.7	66.2	764	6.2	3,180	
4	563	561.4	167.7	12.3	66.7	760	6.2	3,160	Note slight temperature reversal.
5	563	561.4	167.6	12.1	67.2	768	6.3	3,170	Repeat measurement; thermometer not insulated. Core barrel lost Nov. 2.
12	563	550.5	165.6	11.3	70.0	812	6.5	3,245	Thermometer insulated. Water sample from bottom.
						752	6.4	3,160	Normal sample from top.
1951									
Jan. 10	563	550.3	165.6	11.2	69.1	784	6.2	3,205	No drilling since Nov. 3.
14	573	572.8	164.2	12.5	68.3	760	(7.0?)	3,250	Drilling AX inside lost core barrel.
16	603	602.2	(165.4)	12.5	62.2	772	6.2	3,075	
18	620	618.7	(164.9)	12.3	62.0	712	(7.0?)	3,130	Core barrel recovered; NX drilling renewed.
23	633	629.7	163.9	12.8	62.2	772	6.2	3,255	Thermometer well insulated.
			(164.2)	----	----	----	----	----	Normal uninsulated.
25	----	658.8	163.8	12.9	61.7	768	(7.2?)	3,235	Thermometer well insulated.
27	686	684.0	163.7	12.5	64.6	792?	6.9	3,395	Thermometer down all night, insulated.
			(164.2)	----	----	----	----	----	Uninsulated.
Feb. 16	686	683.6	164.0	11.8	66.8	792	6.4	3,290	Thermometer well insulated.
			(165.4)						Thermometer uninsulated. Excess gas pressure, equal to +41 ft water.
1952									
May 15	686	615.9	157.7	14.4	74.4	792	6.5	3,180	Hole has cooled by convection.

duced from 6 inches to 1 or 2 inches within a period of 4–5 months (fig. 39). The thickness in some other near-surface parts of the pipe may be only half as much.

Calcite is deposited in the East Reno well at only half the rate of the Steamboat wells. In striking contrast the West Reno well is nearly choked with aragonite within a period of only 4–5 days (fig. 40). The presence of aragonite rather than calcite in this well is evidently controlled largely by rate of deposition, because water compositions, pH relationships, and temperatures are all very similar; the deep water of the West Reno well is probably slightly higher in calcium than that of the Steamboat wells.

When an erupting well is so nearly filled with calcium carbonate that the discharge decreases to the minimum required by the establishment, the main valve is closed and the well is shut down for cleaning. When the valve is first closed, the well is still very hot in its upper part, with much steam and only a small proportion of liquid water. If reopened immediately, the well again flashes into eruption, but if the valve is kept closed and the well is permitted to cool, the steam condenses and water rises in the casing. In many wells, the valve can be reopened without eruption after a sufficient lapse of time; the water has again "gained the upper hand," and its pressure equals or exceeds the vapor pressure of hot

TABLE 30.—*Data from drill hole GS-4*

[East of crest of Main Terrace, 7E., traverse 3. Diamond drilled, circulating cold water. Measurements in parentheses considered less reliable. Water samples from top of column]

Date	Depth drilled (feet)	Depth thermometer (feet)	Bottom temperature (°C)	Depth to water[1] (feet)	Temperature (°C at water level)	Cl (ppm)	pH	Specific conductance micromhos (25°C)	Remarks
1950 Aug. 30	------	59.5	103.2	3±	------	------	------	------	Drilled Aug. 29; 12 ft of 5-in casing, 55 ft. NX casing.
31	------	106.0	135.6	+1.8	56.4	(768)	(6.6)	(2,970)	
Sept. 1	------	129.7	140.9	+1.2	54.3	(532)	(6.6)	(2,130)	
2	------	152.9	145.3	+1.9	48.9	(352)	(6.6)	(1,520)	
3	172	166.7	149.7	+1.8	52.2	(392)	(6.6)	(1,660)	
4	172	166.7	152.3	+2.5	52.2	(628)	(6.5)	(2,500)	Repeat measurement.
5	172	166.7	153.2	+2.7	56.2	(688)	(6.4)	(2,760)	Do.
6	192	174.5	(147.8)	(+2.5)	(41.2)	(120)	(11.2)	(1,340)	Cemented Sept. 5.
7	216	209.9	156.6	+2.3	35.6	(96)	(7.7)	(556)	
8	254	249.7	164.8	+2.8	56.7	(368)	(6.5)	(1,760)	
9	283	277.6	168.1	+11	71±	(828)	(8.8)	(3,150)	Had been erupting; measurements under pressure.
10	322	318.4	169.4	+5	55½	(596)	(6.7)	(2,520)	Leaking through valve.
11	340	338.2	170.4	+6– +52	95±	(852)	(7.7)	(3,340)	330 ft BX casing; erupting through leaking valve.
						768	6.7	3,120	10 a.m. Collected under pressure; first reliable sample.
12	340	338.2	166.0	+5½	37	768	5.9	3,160	Note decrease in bottom temperature—downward circulation from higher levels proved by series of measurements Sept. 13.
13	393	389.8	163.0	+5	49	772	6.0	3,160	7:30 a.m. Maximum thermometer.
			164.0	+7	------	------	------	------	11 a.m. Corrected thermistor temperature.
			166.3	+65	------	------	------	------	11:15 a.m. Valve leaking 2½± gpm since 11:05 a.m., downward circulation reversed by leakage.
			169.0	+48	------	------	------	------	11:25 a.m. Valve shut, cap leaking 1± gpm.
			168.9	+10	------	------	------	------	Cap leak decreasing, 11:30 a.m.
			168.0	+6	------	------	------	------	Seeping discharge, 11:40 a.m.
			165.8	5½	------	------	------	------	Seeping, 12:55 p.m. Note temperature reversal again from downward circulation probably from 150- to 200-ft depth.
18	393	389.8	170.7	+8	48	796	6.0	3,140	No drilling since Sept. 12; note natural temperature recovery from lower values.
				−5.8	------	------	------	------	Water level after 10 minutes pumping cold water.
19	445	440.3	169.6	+3	46	(344)	(6.2)	(1,495)	Bottom temperature probably slightly low from downward circulation.
						(940)	(7.1)	(3,700)	Drill water used day of Sept. 19, from springs 2–5.
20	------	464.5	170.6	+2	49	804	6.4	3,170	8 a.m. Drill water used again from ditch during day.
			171.1	+61	------	------	------	------	Temperature by maximum thermometer, controlling leakage, as on Sept. 13.
21	------	482.9	170.6	+4½	43	(580)	(6.1)	(2,620)	Ditch water influence.
22	505	503.3	170.6	+3	------	(724)	(5.9)	(2,850)	Hole completed. BX casing removed.
Oct. 30	505	502.9	171.1	+19	------	784	5.7	3,100	Thermometer uninsulated.
1951 Jan. 29– Feb. 3.	------	------	------	------	------	(928)	(8.5)	(3,725)	Erupted water being used in drilling GS-8.
1952 May 30	------	------	170.8	+16	------	816	6.6	3,180	Thermometer insulated.

[1] Positive values after Sept. 8, 1950, are feet above ground level, determined by pressure gage; water in casing above ground level greatly cooled.

water deeper in the well. A cable-tool drilling rig is then used by resort personnel to drill out the deposit (fig. 41). Slowly formed dense calcite is hard and may require a day or more of continuous drilling to clean a well. The rapidly formed aragonite of the West Reno well is soft and porous; it is ordinarily cleaned every 4 or 5 days by repeatedly dropping a heavy iron rod manipulated by cable attached to a jeep. Periodically, this method becomes ineffective and a cable-tool rig is used for a more thorough cleaning.

At times it was not possible to shut down the Steamboat Resort wells; even after a day of inactivity, a well would flash into eruption as the valve was opened. At such times, cleaning was accomplished with greater

TABLE 31.—*Data from drill hole GS–5*

[East of crest of Main Terrace and 418 feet north of GS-4. Diamond drilled, circulating cold water. Measurements in parentheses considered less reliable than the others. Water samples from top of column unless otherwise noted]

Date	Depth drilled (feet)	Depth thermometer (feet)	Bottom temperature (°C)	Depth to water[1] (feet)	Temperature (°C at water level)	Cl (ppm)	pH	Specific conductance (micromhos, 25°C)	Remarks
1950									
Sept. 24	37	34. 2	70. 0	5. 3	45. 4	(304)	(5. 5)	(1, 400)	Drilled Sept. 23. 7 ft of 5-in. casing.
25	58½	58. 3	95. 3	6. 4	44. 2	(12)	(7. 2)	(143)	50 ft NX casing.
				5. 5	42. 3	------	------	------	Water level outside NX casing.
26	90	89. 9	129. 8	+. 6	34. 6	(44)	(7. 0)	(293)	
27	115	114. 5	138. 6	+3. 1	30	(36)	(8. 6)	(255)	
28	139	138. 3	141. 1	+3. 1	29	(160)	(6. 9)	(870)	Well erupts if permitted to discharge.
29	158	157. 3	141. 0	+3. 3	24	(156)	(7. 4)	(769)	Temperature 1 ft below ground, 32°C; cooling in pipe above ground, convection.
30	175	174. 7	151. 6	+3. 4	25	(524)	(7. 0)	(2, 160)	
Oct. 1	175	174. 6	153. 7	+3. 9	24	(764)	6. 8	(3, 050)	Repeat measurement.
2	175	174. 6	153. 8	+3. 6	20	------	------	------	Do.
3	175	174. 0	154. 0	+4. 0	38	820	(7. 3)	3, 290	Sampled after some discharge; first reliable sample.
4	208	206. 7	159. 2	+3. 5	27	------	------	------	Discharge 1 min, sampled.
				+4. 2	61	(780)	(7. 1)	(3, 050)	
5	238	236. 4	164. 7	+4. 0	33	(736)	(7. 2)	(2, 900)	
6	238	236. 4	165. 2	+3. 7	26	808	7. 1	3, 120	Repeat measurement.
7	256	254. 6	165. 9	+3. 4	29	(772)	(6. 8)	(3, 060)	
8	281	279. 1	168. 0	+3. 8	26	(770)	(6. 8)	(3, 030)	
9	300	298. 6	169. 7	+3. 8	------	(768)	(7. 1)	(3, 020)	No water return 250–308 ft.
10	315	313. 6	170. 8	+3. 7	------	(408)	(6. 2)	(1, 840)	308 ft BX casing.
				+(2. 6)	------	(3)	(7. 5)	(101)	Steamboat ditch—drill water, 4 p.m.
11	332	330. 1	------	+6	------	794	6. 1	3, 170	
12	340	338. 8	(166. 9)	+16	------	(688)	(6. 7)	(2, 810)	Thermometer bounced coming up.
13	363	362. 8	171. 8	+14	------	810	6. 3	3, 270	
14	379	378. 9	172. 2	+33	95±	780	(7. 7)	(2, 910)	Valve leaking considerably all night.
15	412	410. 9	171. 8	+36	Hot	808	(7. 2)	(3, 340)	Do.
16	430	------	------	+48	95+	(882)	(8. 0)	(3, 440)	Strong leak, high temperature discharge, no measurements.
				—10–14	------	------	------	------	After 10 min, pumping cold water.
17	465	462. 2	(172. 1)	+18	------	824	6. 3	3, 270	Valve had been completely closed.
18	505	399. 2	(172. 2)	+17	------	820	6. 3	3, 270	Thermometer stuck at 399 ft; uninsulated.
19	545	543. 1	(172. 2)	+17	------	826	6. 00	3, 280	Thermometer uninsulated.
22	575	572. 2	(172. 2)	+20	------	820	6. 2	3, 270	Hole completed Oct. 19.
				8	------	------	------	------	Pressure decreased after escape of some gas.
23	575	572. 2	169. 6	+9	------	826	6. 0	3, 270	Bx casing left in but space between NX–BX not sealed. Thermometer insulated proving slight reversal.
1951									
Apr. 13	------	------	------	------	60±	820	6. 2	3, 310	
						832	6. 9	3, 345	Sample from top of column before eruption; thermometer blocked at 400 ft.
1952									
May 16	575	400	172. 6	------	------	824	(7. 8?)	3, 230	
16	575	469. 6	172. 6	±23	47	------	------	------	Bottom temperature?after eruption; 105 ft sediment in hole.
30	575	469. 6	------	+13	------	828	6. 2	3, 200	Before eruption.
	------	------	------	+60	------	------	------	------	After eruption.
1953									
June 20	575	------	------	------	40±	824	6. 9	3, 290	After slight discharge permitted.
					95+	864	8. 7	3, 340	Erupted sample.

[1] Positive values mean excess pressure in feet of water, as measured by pressure gage.

difficulty by drilling through the erupting column of boiling water.

Some characteristics of Steamboat well 2 are shown graphically in figure 42 through a complete production cycle. The rate of discharge had decreased to about 12 gpm on September 22, 1945, when the well was shut down for removal of calcite. Cleaning was continued on the 23d, but the well erupted spontaneously before the work was completed. A common practice with these Steamboat wells was to clean them in two stages, with the second stage about 1 week to 10 days after the first, as shown graphically in figure 42. The lower part of the well was much easier to clean after it had subsided a little from its very vigorous initial eruption.

Through early October, discharge decreased at a rapid and almost constant rate until October 17 and then decreased at a slower rate until cleaning was again necessary.

TABLE 32.—*Data from drill hole GS–6*

[Near crest of Sinter Hill. Diamond drilled, circulating cold water. Measurements in parentheses considered less reliable than the others. Water samples from top of water column]

Date	Depth drilled (feet)	Depth thermometer (feet)	Bottom temperature (°C)	Depth to water (feet)	Temperature, °C at water level)	Cl (ppm)	pH	Specific conductance (micromhos 25°C)	Remarks
1950									
Nov. 20	70	68. 9	33. 3	(67. 9)					33ft BX casing.
22	89	86. 9	40. 6						
24	113	113. 0	68. 3						
26	160	158	77. 8						
27	212	208. 4	100						
29	212	206. 6	100. 8						
Dec. 1	212	205. 3	102. 5						
2	212	205. 3?	102. 3						
13	212	198. 0	102. 1	154. 0		(14)	(6. 6)	(1, 137)	Still mostly drill water?
1951									
Mar. 27	212	169. 7	92. 2	147. 3	86. 3	(19)	(6. 4)	(360)	Some gas escaping.
1952									
May 22	212	166. 8	93. 8	142. 7	85. 4	12	6. 7	372	For the most part meteoric water but presumably no longer largely drill water.

TABLE 33.—*Data from drill hole GS–7*

[Silica pit area. Diamond drilled, circulating cold water. Measurements in parentheses considered less reliable than the others. Samples from top of water unless otherwise noted]

Date	Depth drilled (feet)	Depth of thermometer (feet)	Bottom temperature (°C)	Depth to water (feet)	Temperature (°C at water level)	Cl (ppm)	pH	Specific conductance (micromhos 25°C)	Remarks
1950									
Nov. 29	20								11 ft of 6-in. casing. Acid-altered granodiorite is friable, very porous.
30	20	10. 4	21. 4	(8)					Much less drill water.
Dec. 1	37	36. 5	43. 9						Do.
2	85								Much less drill water. Hole cemented.
5	110								Much less drill water. Hole cemented again.
6	115	49. 6	44. 4						Measurement temperature probably much too low; 18 in. free-fall drill rods near 110 ft; hole cemented.
11	115	100±	80						Core barrel stuck in hole Dec. 6–11.
13	122	118. 9	79. 7						Core barrel recovered Dec. 11. 120-ft NX casing. First full water return in hole.
14	166	153. 2	111. 3	107. 8	82				Thermometer in 8 ft mud.
15	166	147. 5	118. 4	110. 1	90½	(16)	(7. 6)	(696)	Thermometer in 2 ft mud; repeat measurement.
18	196	192. 7	130. 8	106. 2		(6)	(7. 5)	(397)	Thermometer in 4 ft mud.
19	222	209. 9	130. 4	106. 4	84. 2	(14)	(6. 7)	(591)	Thermometer in 2 ft mud; considerable water loss.
21	230	227. 5	135. 9	106. 3	82. 2	(8)	(11. 5)	(968)	Cemented Dec. 19; drilled out Dec. 20.
22	258	253. 3	134. 8	114. 2	82. 2	(6)	(10. 4)	(386)	
24	285	283. 4	150. 2	115. 3	82. 0	(10)	(7. 1)	(603)	Thermometer in 2 ft mud.
26	285	283. 0	152. 9	116. 3	92. 4	(10)	(7. 2)	(655)	Thermometer in 2 ft mud (boiling at water level).
27	305	299. 0	145. 6	117. 8	89. 0	(10)	(6. 8)	(520)	Do.
28	330	319. 6	146. 4	121. 1	90. 0	(12)	(6. 9)	(524)	
29	340								Lost 82-ft BX casing in hole, cemented.
30	330	187. 7	(130. 3)						New hole in bottom, bypassing casing.
1951									
Jan. 2	340	316. 4	153. 2	119. 5	94. 4	(8)	(7. 6)	(668)	Boiling vigorously; 325-ft AX casing cemented in.
4	364	363. 0	151. 3			(12)	(7. 2)	(504)	No water return; bottom cooled?
5	389	386. 2	152. 2	117. 4	94. 4	(6)	(7. 3)	(465)	Not boiling so vigorously.
	410	409. 0	(110. 0)						5 p.m., 1 hr after drilling ceased.
6	410	408. 5	148. 4	112. 2	89. 9	(8)	(6. 9)	(513)	9 a.m., in 6 in. mud, boiling.

TABLE 33.—*Data from drill hole GS–7*—Continued

Date	Depth drilled (feet)	Depth of thermometer (feet)	Bottom temperature (°C)	Depth to water [1] (feet)	Temperature (°C at water level)	Cl (ppm)	pH	Specific conductance (micromhos 25°C)	Remarks
1951									
7	410	408.4	156.1	+70 P	----	----	----	----	Ready to erupt. Valve not closing; had to pump cold water ½ hr to get thermometer in hole; cold water must be flowing into walls above bottom of hole; note high temperature.
8	----	----	----	----	----	----	----	----	New valve; cemented between AX and NX casings after steam started escaping outside casings.
	----	----	----	----	----	----	----	----	Pressure as much as 82 psi, hole almost got out of control.
9	410	407.6	160.6	+82 P	----	----	----	----	
Mar. 29	410	407.0	160.9	+76 P	----	(10)	(8.2)	(968)	No water level—hole filled with vapor and water droplets with slight condensation of vapor in upper part of hole; remaining vapor escaping into walls.
1952									
May 22	410	407.8	161.0	+80½P	----	----	----	----	Before measurements.
				+71½P	----	0.5	6.5	158	Sampler in hole, slight leakage of gas. Water sample is condensate?
						(24)	(9.7)	(1,640)	Erupted from side valve, 120–325 ft; pH probably from cement.

[1] Positive values, excess pressure (P) in pounds per square inch.

TABLE 34.—*Data from drill hole GS–8*

[East base of Main Terrace, 10 E, traverse 3. Diamond drilled, circulating cold water. Measurements in parentheses considered less reliable than the others]

Date	Depth drilled (feet)	Depth thermometer (feet)	Bottom temperature (°C)	Depth to water (feet)	Temperature °C at water level)	Cl [1] (ppm)	pH [1]	Specific conductivity, $K \times 10^6$	Remarks
1951									
Jan. 30	41	37.0	75.2	27.2	----	856	6.5	3,660	Drilled Jan. 29. 21 ft NX casing. Gas escaping above water level.
Feb. 1	75	74.9	103.3	30.0	70.0	876	(9.3)	3,600	Pumping erupted GS–4 water. Cemented.
						(928)	(8.5)	(3,725)	Water from GS–4 used in drilling, Jan. 29 to Feb. 1.
2	110	102.8	120.7	21.1	68.9	924	(9.1)	3,780	
						(956)	(8.1)	(3,800)	Drill water from GS–4 and spring 6, Feb. 1–Feb. 3.
3	130	115.3	123.2	16.6	55.8	944	(8.7)	3,790	Thermometer in 6 in. fine sand.
4	130	121.8	124.8	15.1	53.3	928	(8.5)	3,790	Hole flushed out, completed.
5	130	121.8	126.6	16.4	55.6	912	7.2	3,810	Repeat measurement.
15	130	121.8	128.7	20.4	64.1	892	6.9	3,800	Do.
1952									
May 15	130	121.8	128.7	25.6	80.2	896	6.6	3,330	Do.

[1] Water samples from top of water column.

COMPLICATIONS OF SIMPLE BEHAVIOR

Nearly all erupting wells at Steamboat Springs have had limited periods of productivity and are considered to be short lived. The West Reno well was drilled in 1939, and by 1965 it was much the oldest producing well in the area. The useful life of some other wells is less than 6 years.

The well here called Steamboat well 1 (see pl. 3 for location) became unreliable early in 1945. Well 2 was being drilled in March 1945 when White and P. F. Fix visited the area briefly before systematic observations were started. Well 2 was temporarily completed at a depth of 120 feet, and production was satisfactory until the spring and summer of 1947, when obstructions of unknown nature forced abandonment of the hole. Two new wells (3a and 3b) were started near the previous wells, but both were abandoned at shallow depths. Steamboat well 4 was then started in September 1947, in the same small area; it was first drilled to a depth of 147 feet (table 19) and later deepened to 220 feet. This well produced satisfactorily until the field study was completed in 1952, but was abandoned during the winter

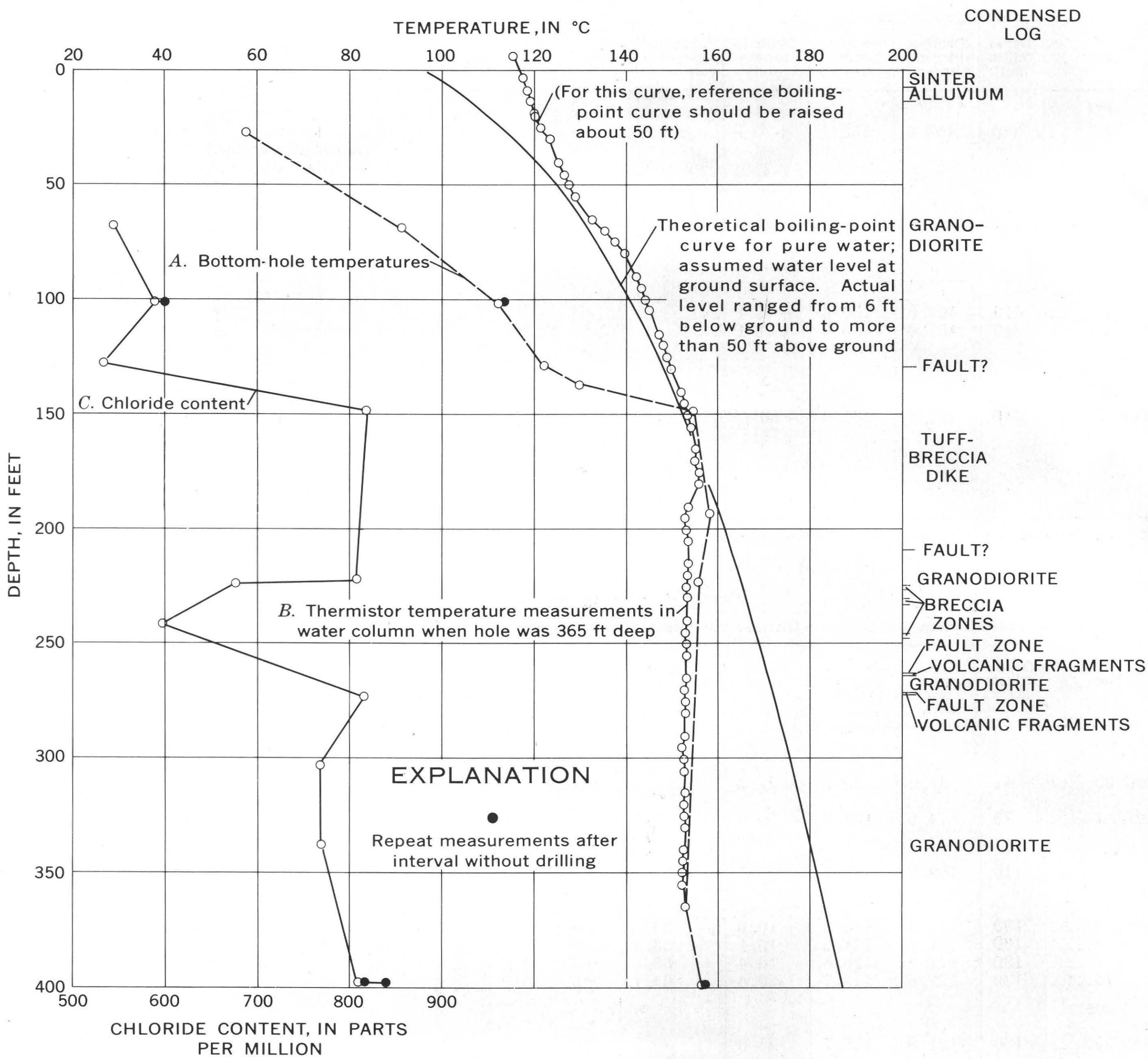

FIGURE 31.—Temperature and related data from drill hole GS-1 near the south end of the Low Terrace.

of 1953–54. A new well was then drilled about 100 feet northeast of the previous wells.

When a producing well is abandoned for any reason and removal of carbonate is no longer attempted, the well does not erupt continuously until choked completely with calcite. If the valves are left open, the well eventually changes from continuous eruption to intermittent, geyserlike eruption. This evolution in behavior took place in each of the producing Steamboat wells that was observed after abandonment.

The behavior of an abandoned geysering well, Steamboat 2, changed with the changing discharge pattern of nearby continuously producing Steamboat well 4, between its periodic cleanings. The interrelationships are illustrated graphically (fig. 43) from water-stage records of the overflow stream from the storage tanks; the records measure the combined discharge from both wells. The discharge from Steamboat well 4 decreased during September and October 1949, as the casing of this well was filling with calcite. Abandoned well 2 was inactive during the early part of this cycle, but probably on September 29 geyser eruptions of very short duration started to occur, first at intervals of 2–3 hours. By October 2 the interval between most eruptions had

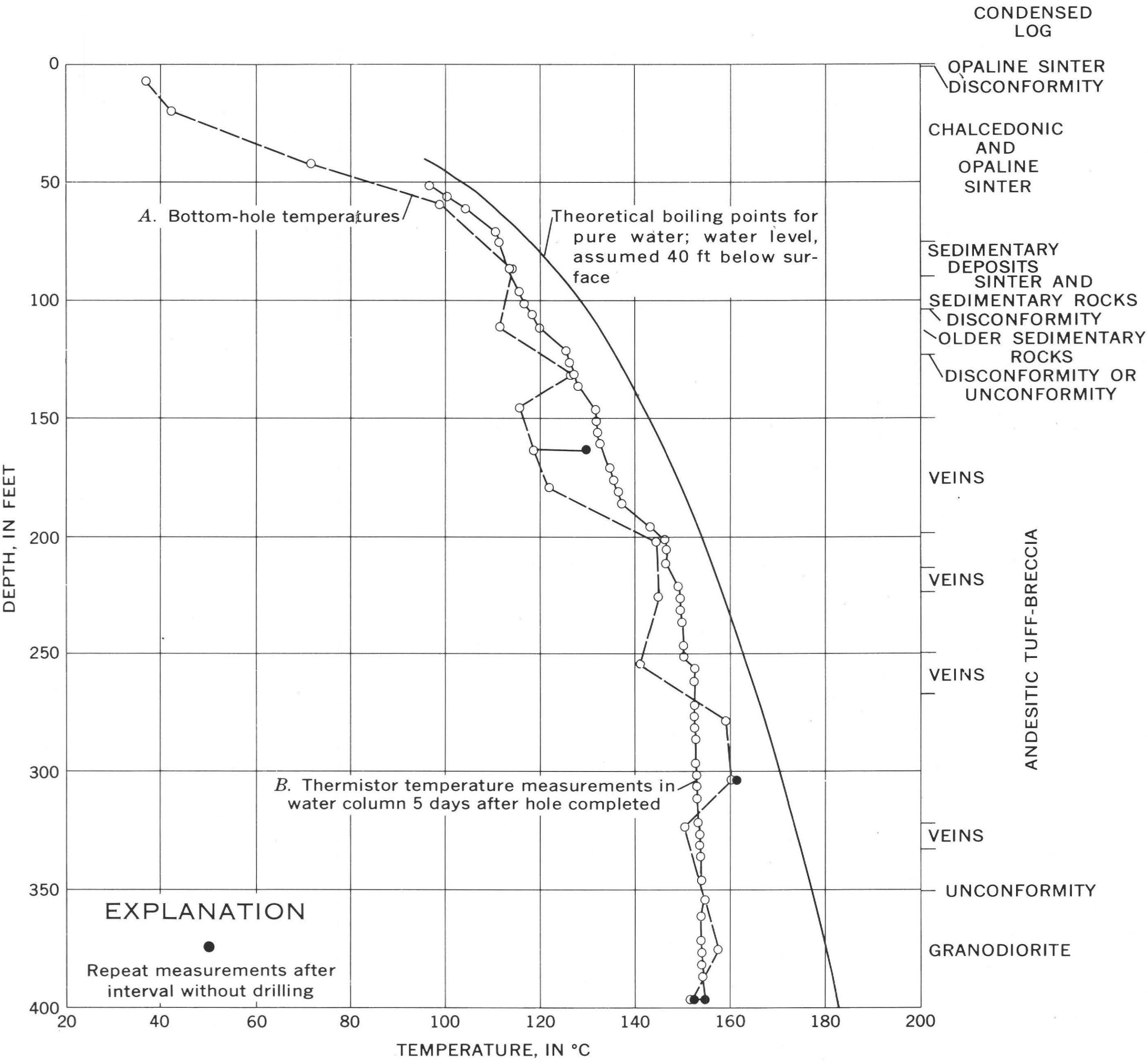

FIGURE 32.—Temperature and related data from drill hole GS–2 crest of High Terrace.

increased to 5 or 6 hours, and the duration and intensity of each eruption also increased, one eruption continuing for about 40 minutes. The intervals and durations both became much longer until October 9, when well 2 started to erupt continuously; each well then contributed about half of the total discharge until well 4 was cleaned in January 1950. The explanation for the change in behavior of well 2 is not known with certainty. Presumably, because of gradual plugging and buildup of pressure in well 4 or in a specific aquifer, water and heat were diverted into well 2.

One characteristic of most recorded erupting thermal wells at Steamboat Springs is a short-term fluctuation in discharge, typically about 10–25 percent of the average discharge. This type of fluctuation is shown on figure 43 as a broad irregular line because of the combination of gears and time constants used on these recorders. Some wells have had even larger fluctuations, best described as geyser eruptions superposed on normal eruption. A single well may have two or more types of superposed eruption, each of different magnitude, duration, and interval between eruptions, as illustrated in figure 44 for Steamboat well 4. Steamboat well 2 was completely inactive at that time (1951). The

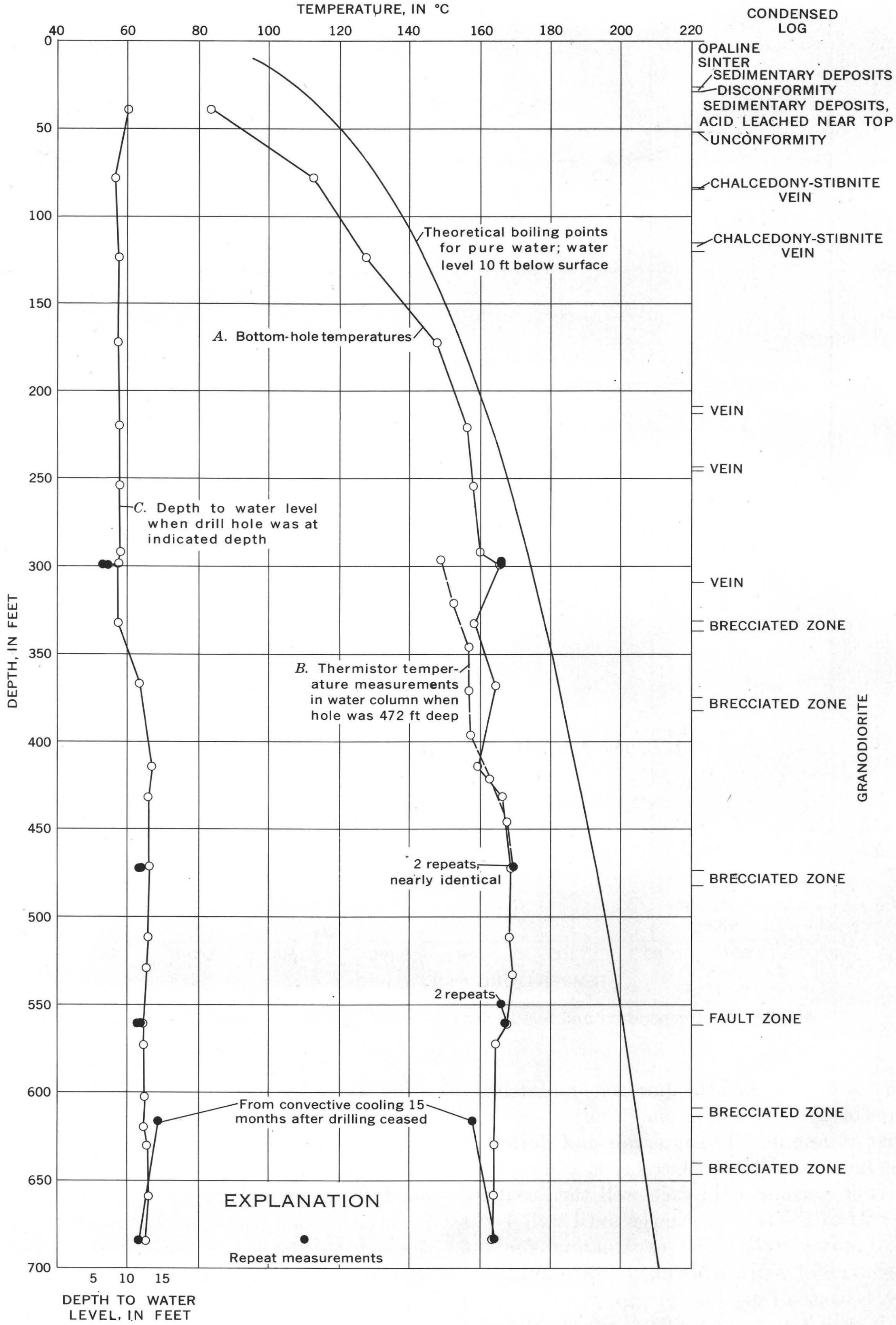

Figure 33.—Temperature and related data from drill hole GS-3 west of crest of the Main Terrace.

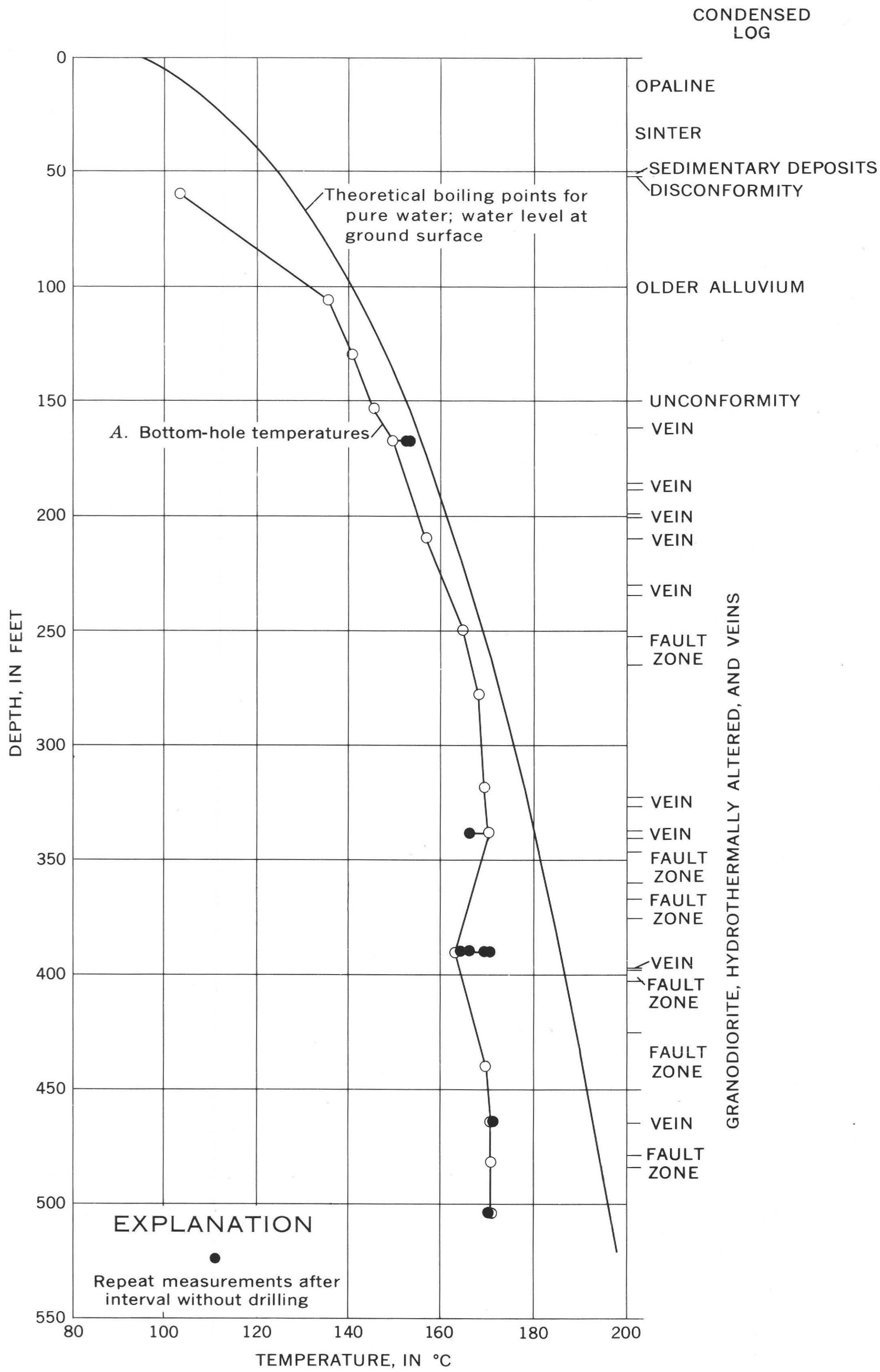

FIGURE 34.—Temperature and related data from drill hole GS-4 east of crest of the Main Terrace.

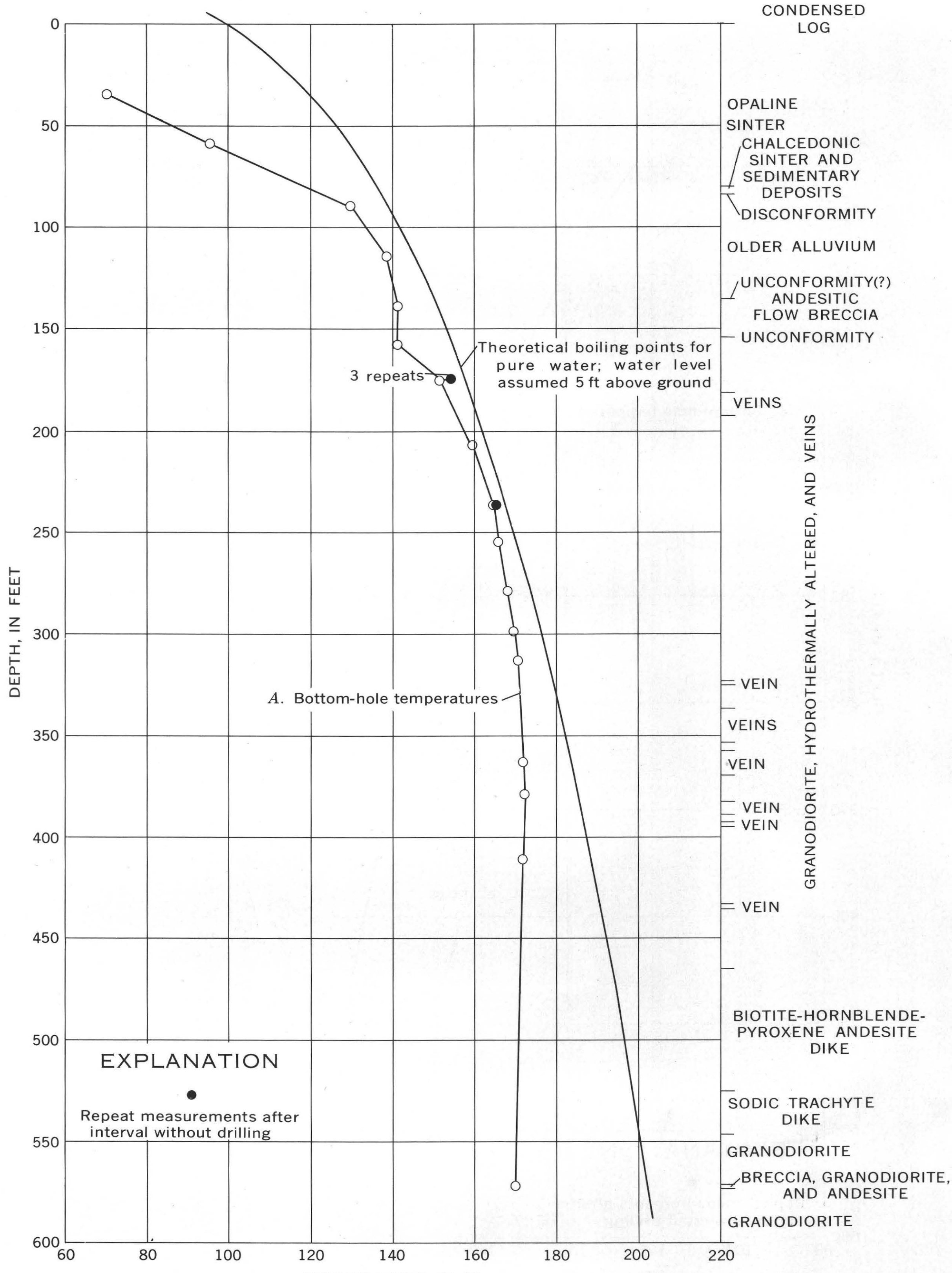

Figure 35.—Temperature and related data from drill hole GS-5 east of crest of the Main Terrace (and 417 ft north of GS-4).

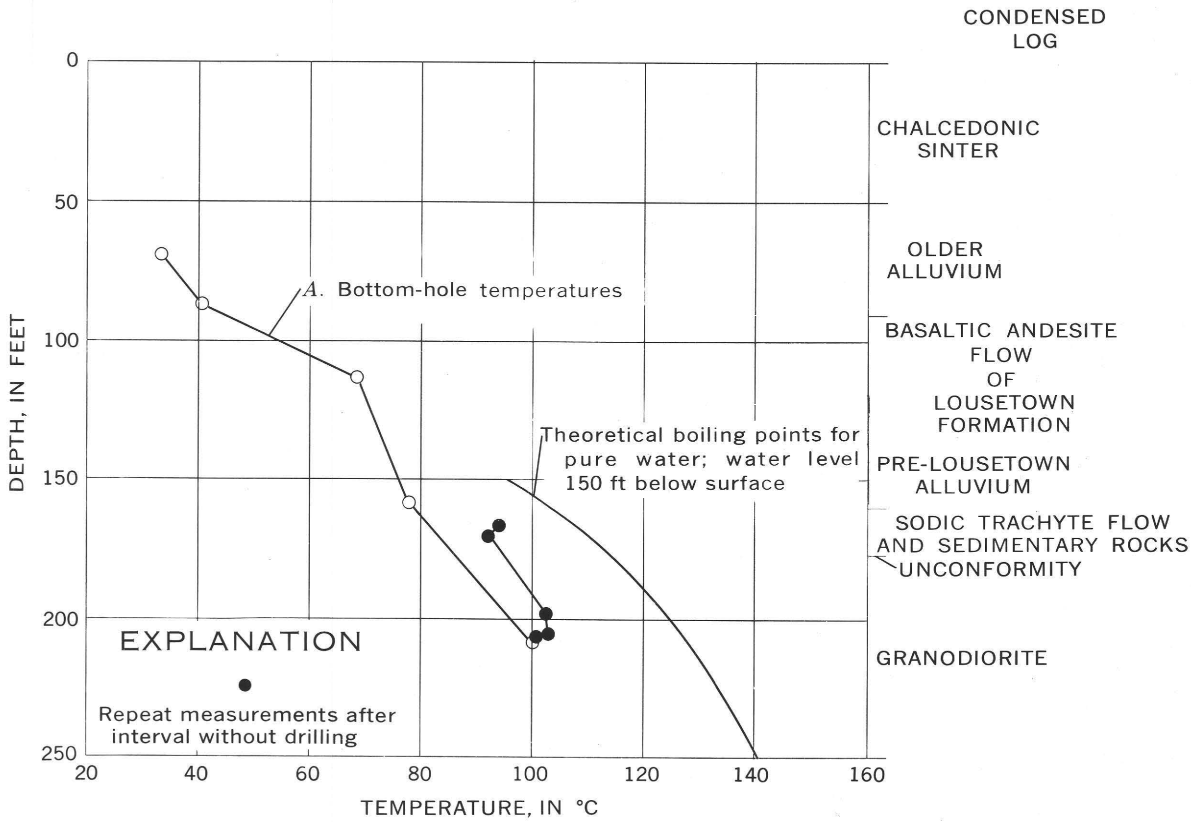

FIGURE 36.—Temperature and related data from drill hole GS-6 near crest of Sinter Hill.

small-scale fluctuations are characterized by brief periods of high discharge separated by 1½- to 2-hour intervals of slightly lower discharge. Until March 19 no other important fluctuation occurred, except for a gradual decrease in average rate of discharge related to calcite deposition in the pipes. Late in the afternoon of March 19, the rate of discharge abruptly increased about 30 percent at the gaging point.[4]

From March 19 to 23 the major superposed eruptions were about 40 hours apart, but by the middle of April the interval was less than a day, and the magnitude of each eruption had become much greater. The peak discharge at the recorder during the major eruption of April 14, for example, increased to 97 gpm (0.37 ft of head on V-notch wier) from a previous average of about 6 gpm (0.125 ft of head).

If a geyser is defined as a spring (or well) with intermittent rather than continuous discharge, the superposed eruptions described here are not true geyser eruptions, because discharge was continuous between the superposed eruptions. Occasionally, one of the Steamboat wells did cease to flow after one of these very strong eruptions. On such an occasion a true geyser eruption had occurred, except that the vent was manmade rather than natural. The phenomenon illustrated in figure 44 obviously has much in common with natural geysers, many of which do show one or more superposed and more vigorous phases during a single eruption. Other geysers have two or more very different intensities of eruption. This is illustrated on a small scale by the graphed behavior of geyser 23n, figure 22, from November 11 through November 15. During this interval the average duration, rate of discharge, and intensity of each eruption was much greater than the general average. The kind of behavior in a continuously producing well that is illustrated in figure 44 is here called a superposed eruption. The smaller fluctuations in this figure are caused by much smaller superposed eruptions; detailed relationships of the records suggest strongly that minor eruptions triggered each major eruption.

The water supply of many geothermal wells is derived from two or more different aquifers at different depths and temperatures. As calcite is deposited in a well and the total rate of discharge decreases, the flow

[4] Probably at least 100 percent at the well head. The water-stage recorder was located about 300 feet downstream from the wells and storage tanks. The actual records are therefore smoothed-out curves of larger fluctuations in discharge.

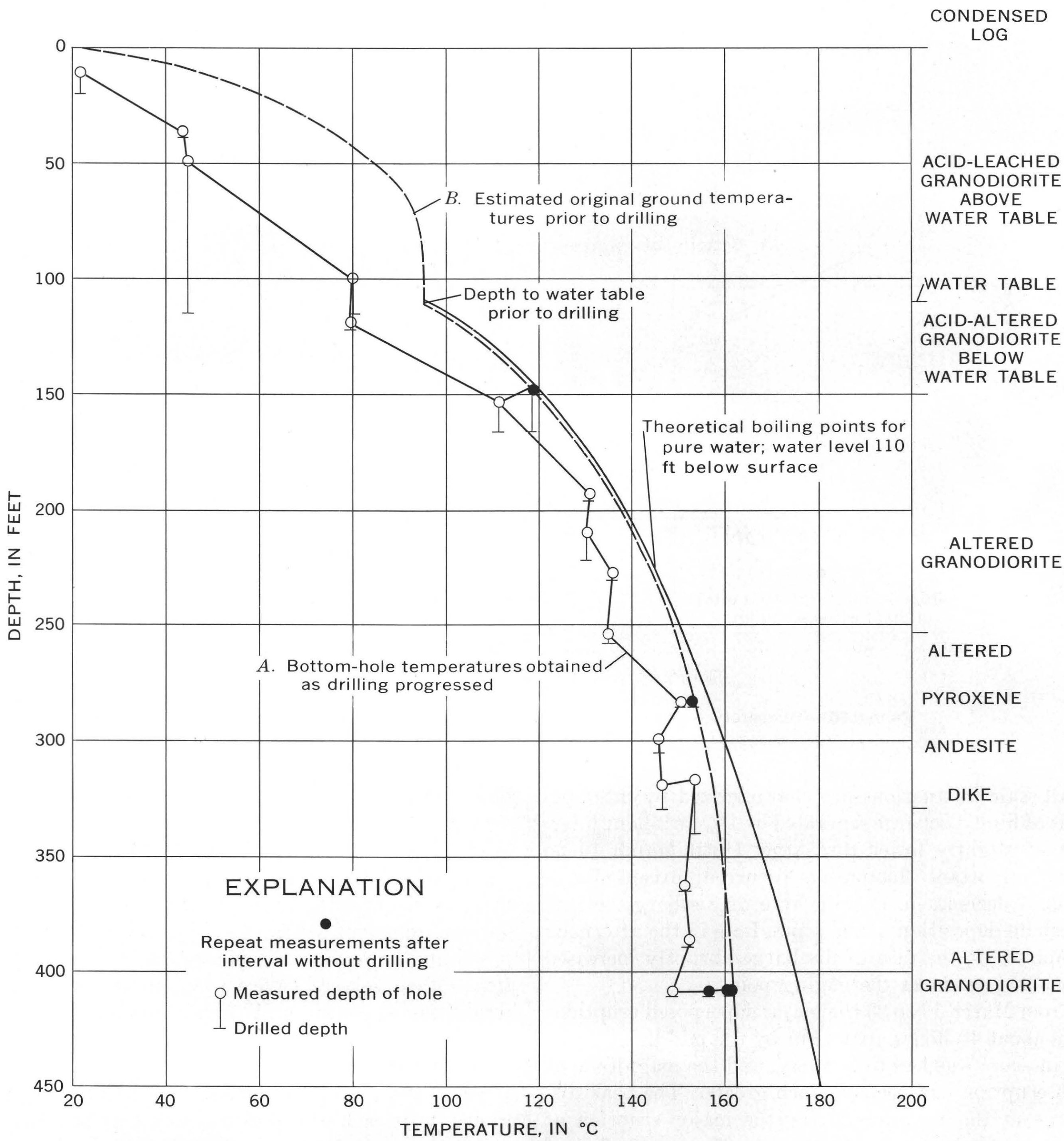

Figure 37.—Temperature and related data from drill hole GS-7, silica pit area.

from one or more of the aquifers may be inhibited relative to the flow from other aquifers. The discharge from an inhibited aquifer may then become intermittent rather than continuous. In relation to each contributing aquifer, a true geyser eruption actually occurs. In Steamboat well 4 the major eruptions probably occurred from a high-temperature aquifer at or near 180 feet in depth (fig. 25). Older wells that produced from the main aquifer near 107 feet were not observed to have vigorous superposed eruptions.

The above-described phenomenon is here called an aquifer eruption. The concept is applicable to continuously erupting geothermal wells that have superposed intermittent eruption from one or more different aquifers; it also helps in understanding the behavior of many natural geysers and pulsating springs.

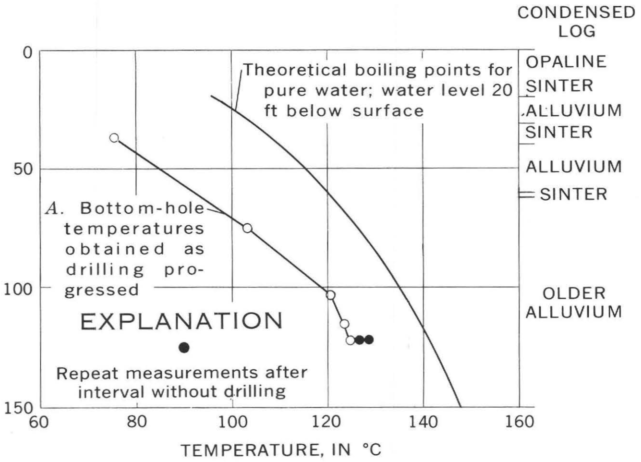

FIGURE 38.—Temperature and related data from drill hole GS–8 near east base of the Main Terrace.

OTHER CHANGES IN GEOTHERMAL WELLS WITH TIME

When a well is drilled, original temperature relationships and circulation patterns are almost certain to be upset to some extent and perhaps drastically. This is the reason why bottom-hole temperatures obtained as drillings progresses give far better data on original ground temperatures than a temperature log measured after completion of the hole. Data from the South Steamboat well (fig. 24), Rodeo well (fig. 26), GS–1 (fig. 31), and GS–2 (fig. 32) substantiate these conclusions. In each of these examples, previously existing circulation patterns were evidently changed by the drilling of the hole. A drill hole can be viewed as a new highly permeable channel that short-circuits tenuous interconnections of low permeability that existed in the structurally complex natural system.

TABLE 35.—*Chloride content of the most reliable water samples from drill holes of the Steamboat Springs thermal area, arranged by intervals of depth of drilled bottom when samples were collected*

Drill hole or well	Chloride content (ppm), by indicated depth intervals (feet) when sample was collected						
	0–100	100–200	200–300	300–400	400–500	500–600	>600
South Steamboat well	820	860	10	----------	----------	----------	----------
Harold Herz 1	240	430	----------	----------	----------	----------	----------
Harold Herz 2	----------	400	----------	----------	----------	----------	----------
Steamboat 4	870	620	----------	----------	----------	----------	----------
Rodeo well	890	930 (146 ′) 500± (166 [1])	[1] 840	----------	----------	----------	----------
Auger hole:							
4	[2] 1,300	----------	----------	----------	----------	----------	----------
5	[2] 1,160	----------	----------	----------	----------	----------	----------
6	120	----------	----------	----------	----------	----------	----------
12	90	----------	----------	----------	----------	----------	----------
13	350	----------	----------	----------	----------	----------	----------
No. 32 Geyser well	890	----------	----------	----------	----------	----------	----------
Murray	510	----------	----------	----------	----------	----------	----------
Mount Rose:							
1	----------	824	----------	----------	----------	----------	----------
2	[3] 1,170	848	----------	----------	----------	----------	----------
3	----------	840	----------	----------	----------	----------	----------
East Reno well	----------	870	----------	----------	----------	----------	----------
West Reno well	----------	900	----------	----------	----------	----------	----------
Senges well	----------	830	----------	----------	----------	----------	----------
Mercury well	----------	600	----------	----------	----------	----------	----------
Nevada Thermal Power Co.:							
1	----------	----------	----------	----------	----------	----------	870
2	----------	----------	----------	----------	----------	----------	[3] 955
3	----------	----------	----------	----------	----------	----------	[3] 1,080
4	----------	----------	----------	----------	----------	45	[3] 874
GS–1	650	830	600	840	----------	----------	----------
2	←——?	——560——	——	?——→			
	←——?	——800——	——	?——→			
3	----------	----------	780	----------	760	780	790
4	----------	----------	830	790	800	----------	----------
5	----------	820	810	810	820	825	----------
6	----------	12	----------	----------	----------	----------	----------
7	----------	----------	----------	----------	10	----------	----------
8	860	900	----------	----------	----------	----------	----------

[1] Mixture from higher levels?
[2] Probably influenced by evaporative concentration at the surface.
[3] Generally 10–14 percent too high because of concentration by eruptive boiling.

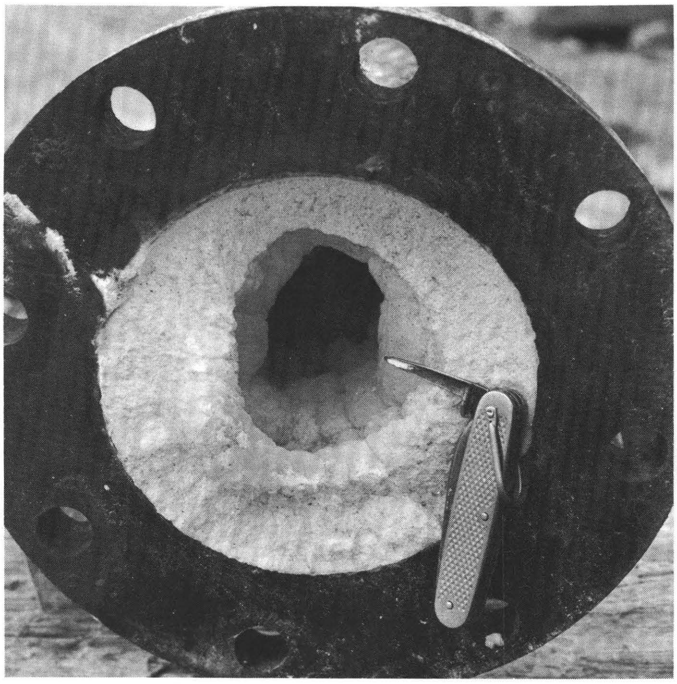

FIGURE 39.—Calcite deposited in Steamboat well 4 during 4 months of eruptive discharge, 1952. The deposit commonly emphasizes seams and other irregularities in pipe and fittings.

Even where little or no net discharge occurs from one aquifer to another, convection may be initiated that changes the original temperature distribution. Harold Herz well 1 (fig. 27) is a particularly clear example. After completion of this well, convection occurred between water level near 60 feet and mud level near 133 feet; the lower part of the well was cooled as convection heated the upper part of this interval. Permeabilities are evidently very low; the water table was penetrated at about 55 feet, but as drilling progressed, the depth to water increased to 60 feet. If permeabilities of all aquifers had been high, cooler water from near the water table would have flowed down the hole into an aquifer of lower potential. Such circulation evidently did occur to a very minor extent but was limited by low permeabilities in much, and perhaps all, of the well. Convection with similar influences evidently also occurred in the South Steamboat well (fig. 24), GS–2 (fig. 32), and GS–3 (fig. 33), although the evidence is somewhat less convincing.

In GS–3 (fig. 41) and the Rodeo well (fig. 26), progressive temperature changes continued to occur over a long period of time.

The Rodeo well (table 20; fig. 26) was completed April 15, 1950, at a drilled depth of 282.5 feet. On August 17, 1950, a water-stage recorder was installed on the uncapped well. Observations of the previous 4 months seemingly indicated that temperature relationships were stable and an eruption was unlikely. Table 20 provides evidence, however, that the temperature at water level had increased rapidly to 81.0°C on April 25 and continued to increase slowly to 86.4°C on August 17, when the recorder was installed. The initial small-scale fluctuations, perhaps related to rising gas bubbles, were about 0.02 foot. By late November 1950 when the records of figure 21 were obtained, the small-scale fluctuations still average less than 0.025 foot (only the median of these small-scale fluctuations shown in curve *A*); by early June the average had increased to about 0.03 foot, and by June 22, to 0.04 foot. Figure 45 shows the gradual changes in amplitude that occurred during the following 2 weeks, culminating in an eruption during the late afternoon of July 5, that destroyed the recording instrument and platform. It is evident now that temperatures continued to increase, at least in the upper part of the well, throughout the previous year. No water-level temperature was obtained after the instrument was installed, but the behavior of the well and the increase in amplitude of the small-scale fluctuations are evidence that water-level temperatures increased to about 95°C on July 5. Another change that was also occurring was a decrease in depth to the water table. At times corresponding to the fluctuation data generalized above, the general depth to water was 12.2 feet below the instrument platform (7.6 ft below ground level) on August 17, 1950; about 12.0 in November 1950 (fig. 21); 11.2 early in June 1951; and 10.9 just before the eruption on July 5. This decreasing depth to water level probably correlates directly with increasing temperatures and thermal expansion of water throughout the column. The net decrease in depth to water through the above time interval is 1.3 feet, which is very close to 0.5 percent of the total water column of about 276 feet. Graphically, from curves *A* and *C* of figure 30, we can determine how much the initial average temperature of the water column (~140°C) must be increased to decrease the density by 0.5 percent. With several approximations, we find that an increase in average temperature of 5°C to about 145°C can explain the observed results. It seems likely that such an increase did occur.

INFLUENCE OF PRODUCING WELLS ON NEARBY SPRINGS AND VENTS

Spring 33, also known as the "Chicken Soup" spring, is about 475 feet southwest of the Steamboat wells previously described. The rate of discharge of this spring is relatively high when discharge from the wells is low, and vice versa (fig. 42). The response to change was particularly clear on October 2, 1946, when Steamboat well 2 was shut down shortly before 8:30 a.m. Spring 33 was first measured at 9:30 a.m., when its rate of discharge

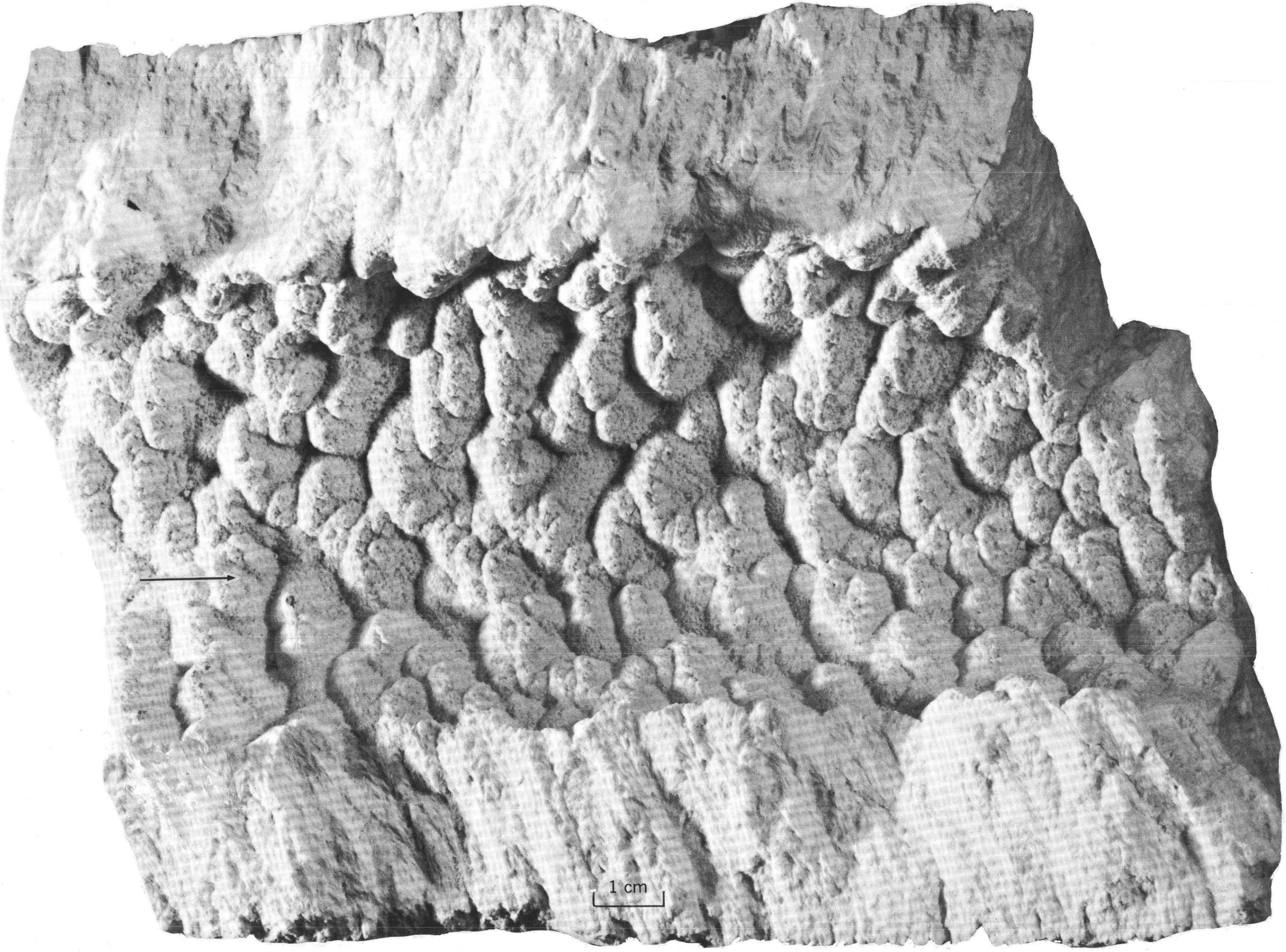

FIGURE 40.—Aragonite deposited in 5 days, West Reno well, March 28 to April 2, 1946. Plumose clusters of crystals grew inward from the well casing and inclined to the left; direction of flow of erupting water indicated by arrow.

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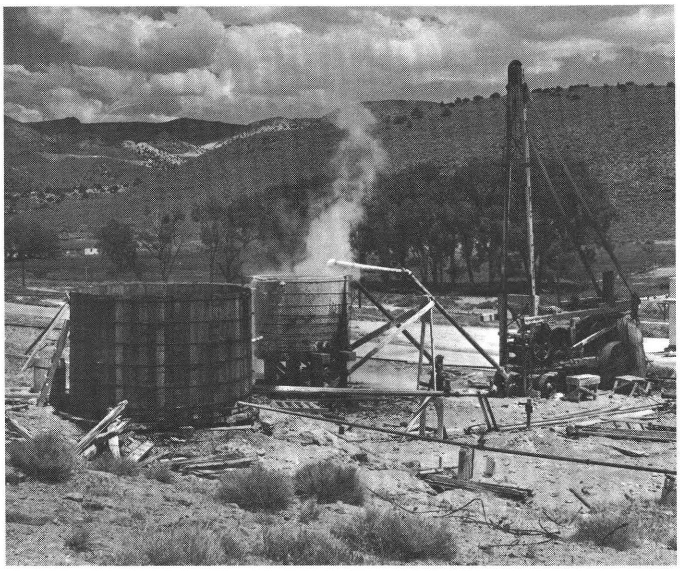

FIGURE 41.—Erupting Steamboat well 4 with inactive well 2 on the left, June 1952. A steam-driven cable-tool drilling rig is positioned to drill out calcite deposited in the casing. A part of the erupted water is discharged into the storage tank and part is diverted horizontally through a side valve.

had already increased to 0.64 gpm from 0.13 gpm of the previous day. This spring had attained a minimum rate of discharge 9 days after the well was first cleaned. Because the spring is so strongly influenced by changes in the Steamboat wells, its discharge is not included in the total measured discharge of plate 4.

A similar but less pronounced response is shown by No. 32 Geyser well 350 feet northeast of the Steamboat wells (pl. 3 and fig. 42). The water level is generally high when discharge from the Steamboat wells is low, and vice versa. Throughout the time interval shown in figure 50, the Geyser well was erupting at intervals of a few days to several weeks. The measurements shown in figure 42 are selected and are least influenced by eruptions and barometric pressure. Many additional measurements are not shown here.

The water level in No. 32 Geyser well attained a general minimum late in October 1945 after cleaning of Steamboat well 2; the water level rose fairly steadily through November and December. The general decline of 1½ feet through January and early February is not related to the Steamboat wells but is one of the random fluctuations of unknown origin discussed on page C82. If behavior near the Geyser well had been normal, its water level would have risen to a maximum when discharge from the Steamboat wells was near a minimum.

Response to the cycle of the Steamboat wells is shown by all other measured vents and springs within a radius of 300 feet of No. 32 Geyser well. None of these vents discharged continuously, but a few erupted as small geysers, generally in response to major eruptions of the Geyser well (table 4).

Weekly measurements were made of springs 25 and 26 and, after January 1946, of spring 50 also. Discharge data from these springs are shown in figure 42. Springs 25 and 26 commonly showed inverse reactions; when the discharge of 25 was high, that of 26 was generally low, and vice versa, providing a clear example of an exchange of function (Marler, 1951). The total discharge from both springs is shown in figure 42.

None of these more distant springs, including spring 50, shows convincing response to changes in the producing Steamboat wells. The general increase in discharge of springs 25 and 26 from October through December is probably either a random increase or the normal increase that many springs show from late summer to winter. The discharge from springs 25 and 26 and from spring 50 was nearly constant for 2 months after Steamboat well 2 was cleaned in March 1946. Any effects on these distant springs that may be due to changes in the Steamboat wells are very minor and are concealed by other fluctuations related to rainfall, barometric pressure, and random change.

The South Steamboat well, drilled during the winter of 1946–47, is 800 feet southwest of Steamboat Resort. Saline thermal water was encountered from depths of about 30–160 feet. Thermal water that is very low in chloride (6–8 ppm) was found at greater depths; chemical and isotope data provide strong evidence that this thermal water is entirely meteoric in origin. The upper saline water is excluded from the completed well by casing that is 188 feet deep. All fluctuations of water level thus are related to the deep meteoric water that is being heated as it flows into the thermal system.

Figure 46 shows the marked changes in water level of the South Steamboat well that occurred during and immediately after the cleaning of Steamboat well 2 on August 12, 1947. The effects are comparable to those of spring 33 shown in figure 46.

Figure 46 contains clear evidence that the South Steamboat well has a permeable connection with Steamboat well 2. The hydrodynamic gradient or direction of flow of the deep dilute water is normally from the southern well toward Steamboat well 2 (see water level altitudes, table 5), which erupts very hot saline water. The water level of the South Steamboat well had been rising during the summer of 1947 as the rate of discharge from well 2 was decreasing because of calcite deposition.

The water level of the South Steamboat well decreased 2 feet in the 3 days following the cleaning of well 2 on August 12, 1947. In contrast, the comparable response of No. 32 Geyser well only 350 feet northeast of the erupting well, was no more than a few tenths of a foot.

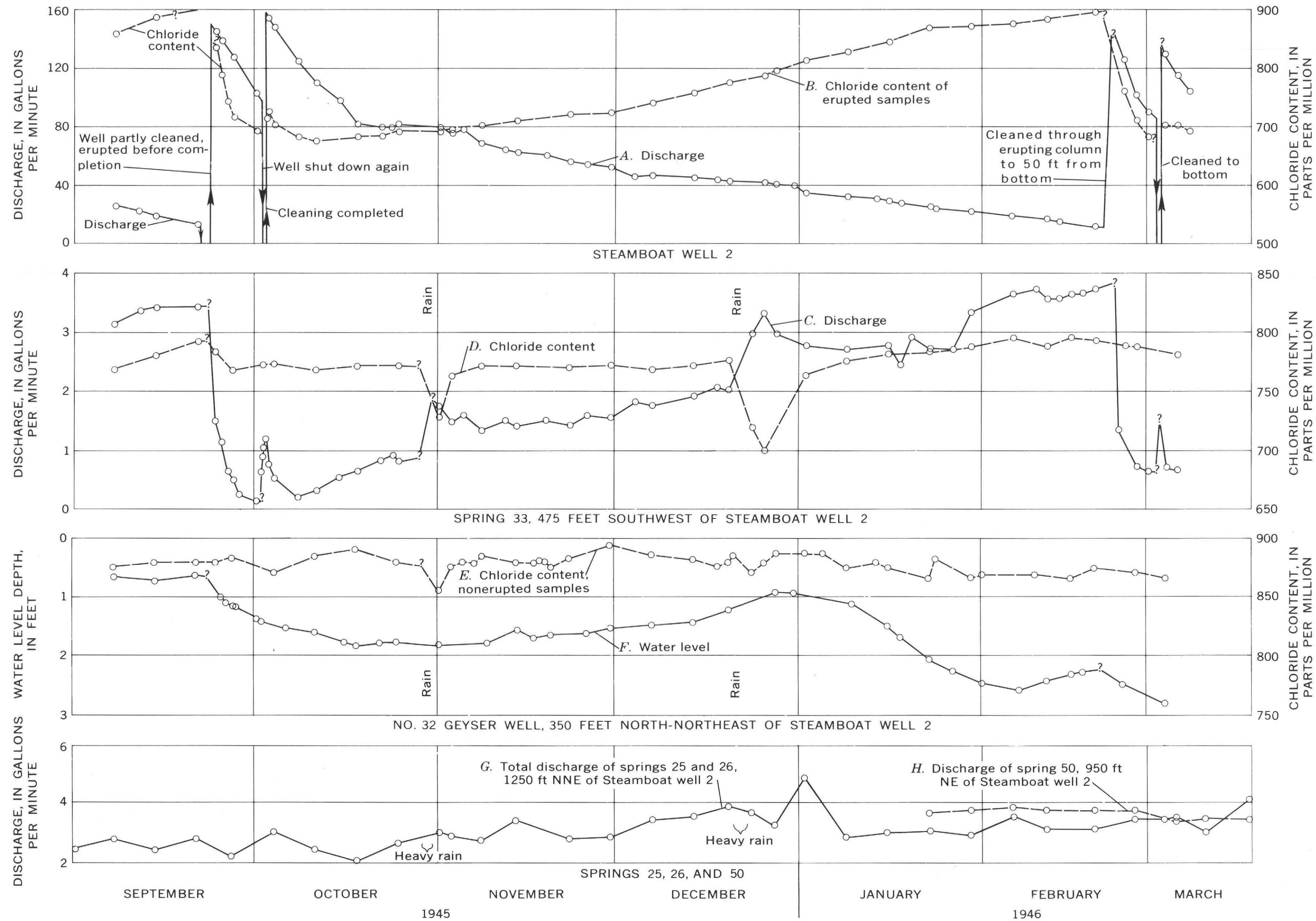

Figure 42.—Production cycle of Steamboat well 2 as related to deposition of calcite in the casing. Shows influence on nearby springs and No. 32 Geyser well, Low Terrace.

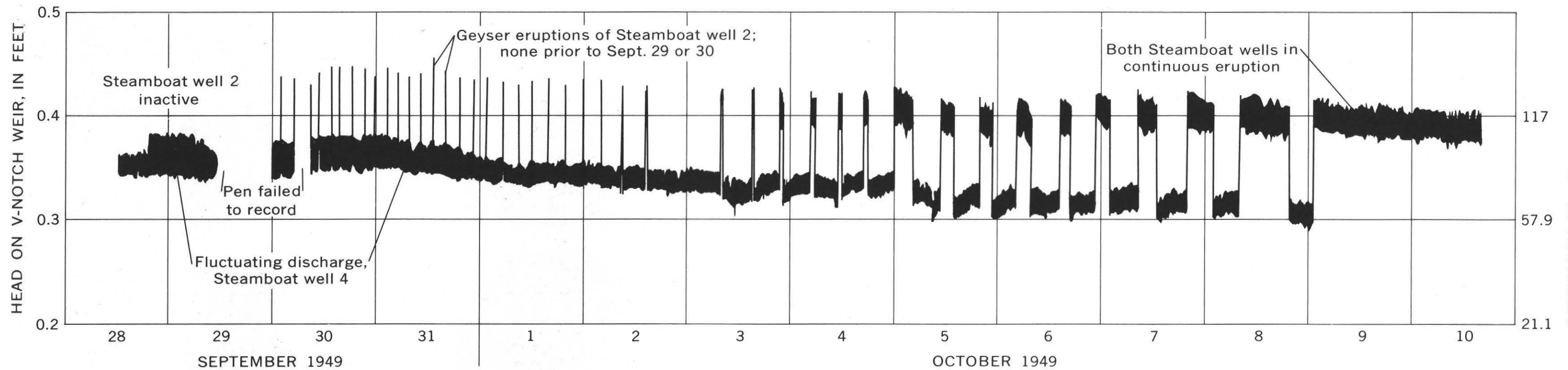

Figure 43.—Combined discharge record of Steamboat wells showing initiation of and changes in geyser eruptions of well 2 as discharge of well 4 decreased because of calcite depositing in its pipes.

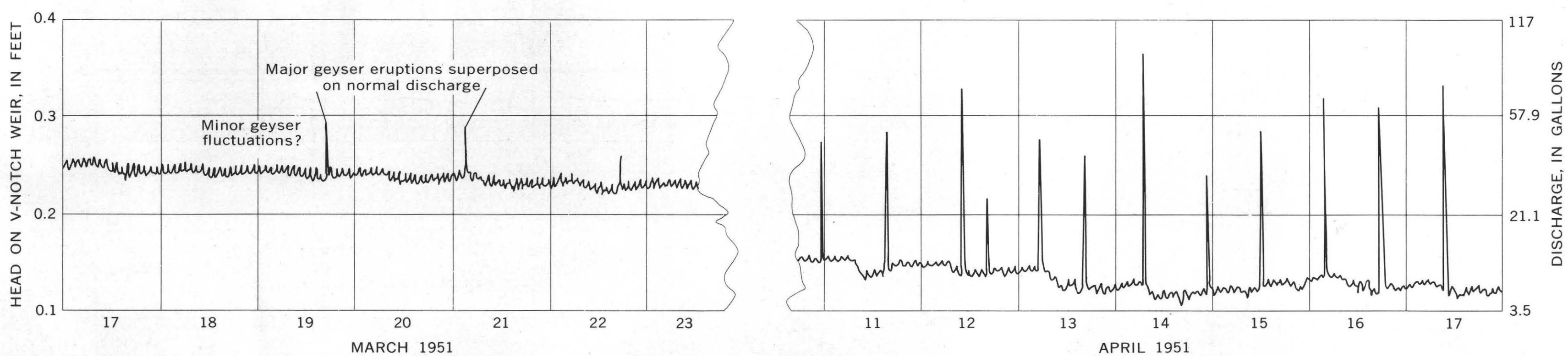

Figure 44.—Discharge record of Steamboat well 4 showing geyser eruptions superposed on normal discharge. Major eruptions increased in frequency and intensity as normal discharge decreased because of calcite deposition.

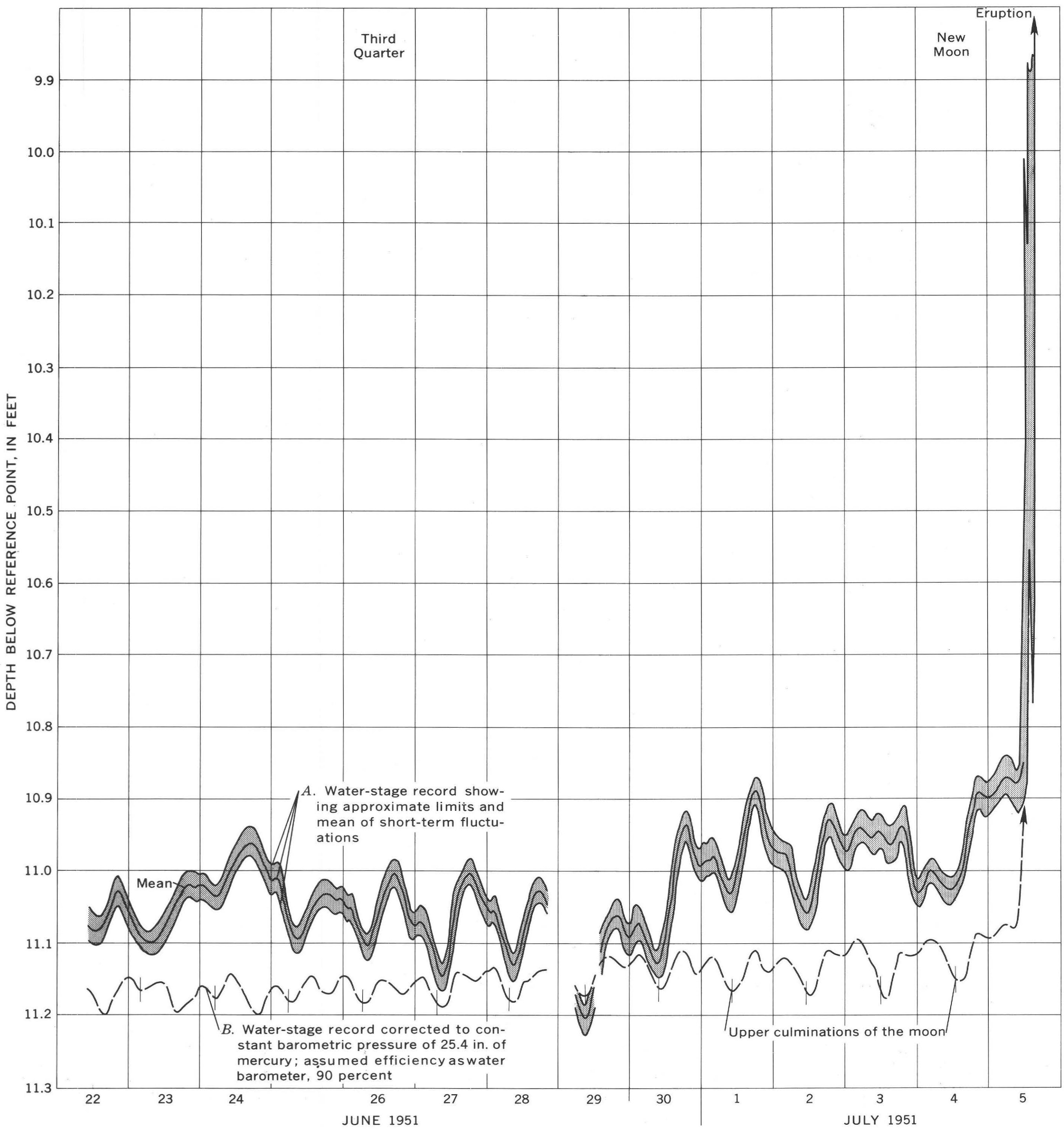

FIGURE 45.—Fluctuations of water level, Rodeo Well, Main Terrace, preceding the eruption of July 5, 1951. (Reference point for water levels about 4.65 ft above ground level.)

Major changes in chloride content occurred during a cycle of activity of Steamboat well 2 (fig. 42). A smaller but similar change is evident in spring 33, complicated by an influx of precipitation. The chloride content of the spring attained a maximum value just before the Steamboat well was cleaned; it then fell to a minimum within a few days and gradually rose again. No similar change is evident in No. 32 Geyser well, and none could be expected in the dilute water of South Steamboat well.

When Steamboat well 2 was cleaned in September 1945 and also the following February, its chloride content was very high (about 900 ppm, which is near the

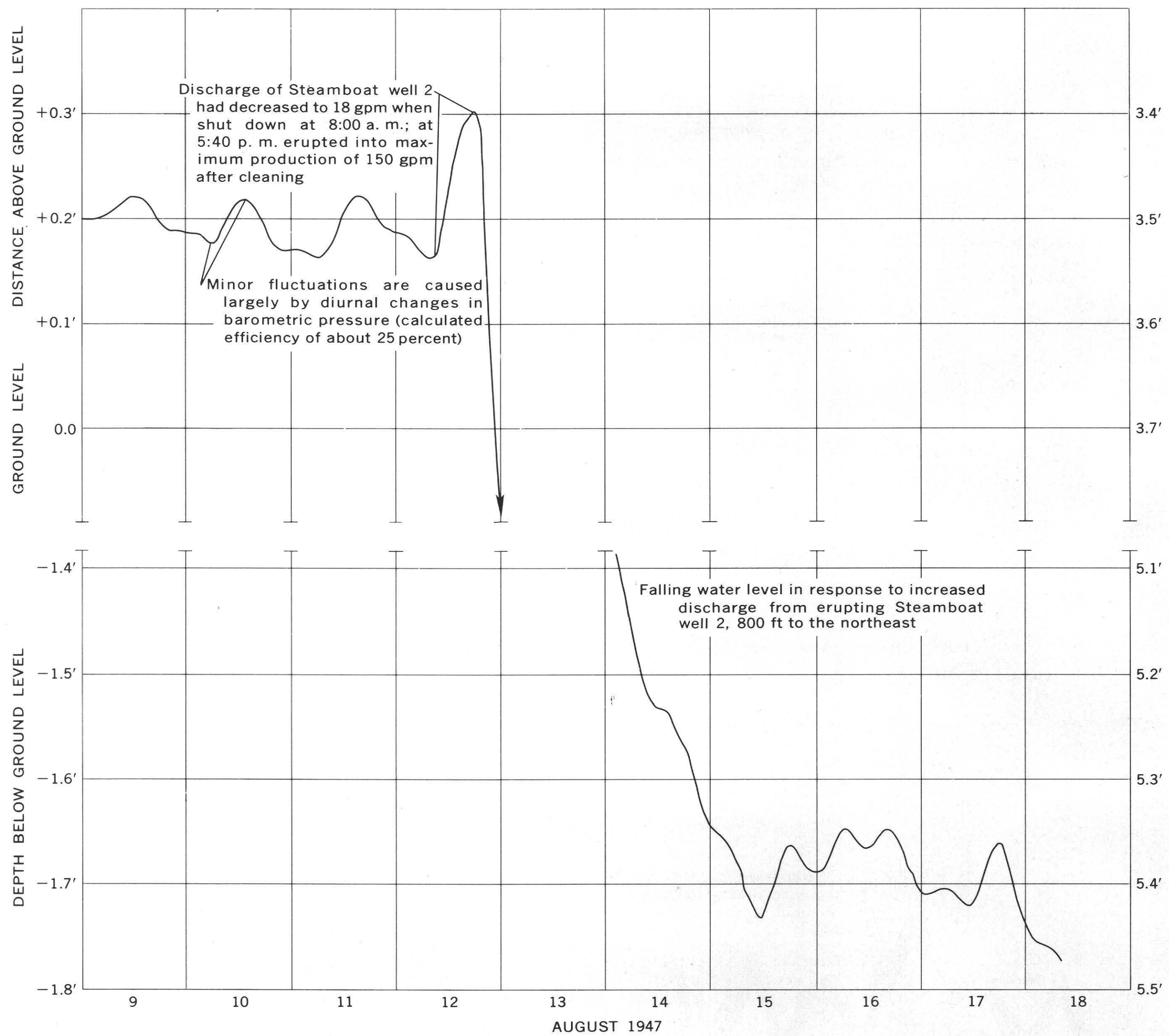

FIGURE 46.—Level of meteoric water in South Steamboat well showing influence of Steamboat well 2 cleaned of calcite on August 12, 1947.

content of most springs). At such times, minimum dilution by low-chloride water is indicated. In contrast, when the chloride content of erupted water was 800 ppm or less, extensive dilution by low-chloride water is indicated. Evidently, when total rate of discharge is low, pressures build up and near-surface inflow of meteoric water is largely excluded. When water levels in the surrounding area rose as eruptive discharge from the well decreased, water relatively high in chloride was stored in pore spaces and this stored high-chloride water then became available when the well was next cleaned. This explains why water discharged from the newly cleaned erupting well and from spring 33 was first high in chloride, and then decreased to a minimum as decreasing pressures permitted entry of low-chloride water from the deep aquifer of South Steamboat well.

INFLUENCE OF PRODUCING WELLS ON TOTAL MEASURED DISCHARGE

Most of the previous discussion concerns the producing wells of the Steamboat Resort, because these wells were easily measured and were close to flowing springs.

The behavior and influences of the wells north of the Main Terrace could not be determined in detail, but they are summarized here because they probably account for major changes that occurred in springs of the Main

Terrace between 1916 and 1945. The present West Reno well was drilled about 1939 to a reported depth of 205 feet and is cased to 50 feet. The measured depth on March 21, 1946, was only 185.8 feet.

The East Reno well was drilled in March and April 1940 to a depth of 172 feet and is cased to 100 feet. The measured depth in October 1946 was only 156.5 feet.

Other wells have been drilled from time to time at the Mount Rose Resort, 3,000 feet north of the Main Terrace. This resort has had difficulty in obtaining a reliable supply of hot water by the self-eruption mechanism. A common practice was to induce eruption by compressed air introduced through a small pipe extended close to the bottom of the hole. These induced eruptions generally ceased after an hour or so. The average discharge over a day or week was never determined but probably was only several tens of gallons per minute, because the requirements of the establishment were low.

In contrast, hot water from the Reno Resort wells supplied 10 small pools and one very large outdoor swimming pool. The hot water was erupted into large tanks near the wells and, after cooling, was piped to the pools. The total discharge was not accessible for measurement at any place, except after a large part had gone through the various pools. Even then, the discharge was distributed between two and sometimes three separate channels leading to Steamboat Creek.

From July 1949 to December 1951 a water-stage recorder was maintained on the largest of these discharge streams, but little useful information was obtained. The ratio of discharge between the two streams was not constant; even more important, the large pool was completely or partly drained at irregular intervals and then filled. At such times the large pool was drained as rapidly as possible to avoid unnecessary delay in refilling; the maximum rate of discharge exceeded 2,800 gpm at the gaging station. Even after efforts were made to line the channel near the gaging station with concrete, the wier and recording setup were periodically washed out.

The best measurement of combined discharge from the Reno wells is based on the time required to fill the large outdoor swimming pool. This pool is 240 feet long and 100 feet wide, and the average depth of filling is 5.2 feet; its capacity is therefore about 125,000 cubic feet, or 935,000 gallons. The time required for filling depended largely upon how much carbonate the wells contained but generally ranged from 3 to 4 days. If an average time of 3½ days is assumed, the rate of discharge into the pool was 190 gpm. Even when the pool was filling, about 10 gpm was used for other purposes in the resort. The total average discharge from both erupting wells is therefore close to 200 gpm.

The West Reno well has a discharge of about 400 gpm immediately after its casing is completely freed of carbonate. The discharge then decreases rapidly as aragonite is deposited in the pipes (fig. 40) until removed every 4–5 days. Aragonite as much as 2 inches thick is deposited near the well head, and the thickness may be locally greater from about 60–95 feet in depth. Most of the water supply is erupted from depths of 95–110 feet and perhaps from near the bottom of the well.

The East Reno well generally yields one-third or less of the total discharge of both wells; the East well is used principally during the summer months when the large swimming pool has maximum use. Hot water from the East well is then run through a cooling system to attain the desired temperature.

A water-stage recorder was installed on the East Reno well during a nonerupting interval from the middle of March to the middle of April 1946. The record showed strong fluctuations that were mostly related to activity of the West well. Water levels during the month ranged from 6.2 to 9.3 feet below the surface; the highest levels coincided with times when the West well was nearly choked with aragonite. After cleaning, the level in the East well fell rapidly for 3 days and then rose rapidly during the fourth and fifth day until the West well as next cleaned. The highest level was 6.2 feet below the surface attained on March 28, 1946, when the West well was shut down for thorough cleaning by cable-tool rig. When the West well was erupted into production, the water level of the East well started to respond almost immediately, and dropped 0.1 foot within 1½ minutes after initiation of the eruption despite the fact that the two wells are 560 feet apart. Both wells evidently share at least one common permeable aquifer.

Calcite is deposited slowly in the East well, whereas aragonite is deposited rapidly in the West well. During the winter months no effort was made to keep the East well free of carbonate. In the fall of 1946, this well was shut down successfully by closing the valve, but in the following spring it was induced into eruption only with great difficulty. During each of the following winters this well was permitted to erupt without cleaning, and its activity changed from continuous eruption in the late fall to intermittent geyser eruptions during the winter and late spring. When cleaned again each spring, the well would again erupt continuously.

The East Reno well influences spring 53 and other nearby vents 200–400 feet south and southeast of the well (pl. 1; table 4). Spring 53, which is 300 feet southeast of the East well, does not discharge when the well is producing continuously; the spring started to discharge in November 1945 after the East well was shut down, and it ceased flowing within 3 or 4 days after the well was cleaned and brought into production late in April 1946. When the spring was active, its discharge ranged from 0.1 to 2 gpm; when inactive, the water level was normally not more than 1.8 feet below the surface.

There is no direct evidence that the Main Terrace springs are influenced by the Reno wells, 2,000 feet to the north. The activity of the springs is shown graphically on plate 4. The West Reno well erupted continuously throughout the period of measurement except when shut down for major cleaning by cable-tool rig. The average rate of discharge from the well changes somewhat in relation to the major cleanings, but weekly and monthly averages are much more nearly constant than for the Steamboat wells. The East well, on the other hand, is always at or near its maximum during the summer months, when water is most needed, but production decreases to minimum levels during the winters when the well is either shut down or erupts intermittently as a geyser.

No detailed correlation was found between the activity of the wells and the rates of discharge of the springs. Detailed measurements of spring 21, the northernmost of the Main Terrace springs, showed no evident correlation with changes in the Reno wells. If the wells do influence the springs, the relationship is a greatly dampened one, obscured by other influences.

The best evidence for a major but dampened influence of the Reno wells on Main Terrace activity is provided by L. H. Taylor's observations in 1916 (fig. 8; table 3). His map shows many points of discharge or visible water in the northern half of the Main Terrace in places where water was not visible from 1945 to 1965. According to Taylor, the total was about 180 gpm in October 1916; in contrast, it did not exceed 90 gpm at any time during the interval of systematic observation from 1945 to 1952; the average discharge for the months of October of these years was about 45 gpm, or 135 gpm less than in 1916. This difference is close enough to the average of 200 gpm estimated for Reno well production to support the view that the wells discharge a large part of the supply that formerly flowed from natural springs.

RANDOM INFLUENCES OF UNDETERMINED ORIGIN

We have considered a variety of influences that affect the springs and geothermal wells in different ways. Some of these relate to specific causes that have been identified and, for barometric and earth tidal responses, are predictible. The effects of earthquakes on the other hand seem largely unpredictible.

We have also seen evidence for many changes of undetermined origin that seem to occur frequently and unpredictably in the system. For example, on plate 4, note the unsystematic intervals of discharge of many individual springs. A comparable behavior is evident in curves *E* and *F*, which show depths to water level in vents 35 and 36. Initially, measurements were made in these vents in the hope that depths to water level could be keyed to total discharge from the Main Terrace. These and other measurements soon proved that activity fluctuates everywhere, and that no single point provides an index to the general level of activity.

We have seen that significant changes occurred in the system during the precipitation years 1949–50 and 1950–51 that resulted in a net decrease in discharge of at least 20 percent and a decrease in chloride content of about 3 percent from the values predictable from general trends of the data.

The major random changes may possibly be caused by changes deep in the thermal system, such as seismically induced changes in permeability of fissures that lead from a magma chamber. Other similar changes may occur at higher levels in the thermal system, permitting variations in proportion of different waters with different energy and salinity contents. Local shifts that frequently result in increased activity at one vent, perhaps offset by an equivalent decrease in other nearby vents, are probably all determined within the upper few hundred feet of the system. One possible explanation for such changes is deposition and flushing of silica gel, precipitated near the surface. Similar changes occur very frequently at Mammoth Hot Springs of Yellowstone Park, where abundant calcium carbonate is deposited.

In contrast, springs that lack rapid chemical precipitation at the surface commonly have very stable vents that do not readily change position. Examples that can be cited from the author's personal experience include: Wilbur Springs, Calif.; Utah Springs, Utah; Amedee Springs, Calif.; and Bowers Spring, Nev. (White and others, 1963, tables 15, 16, 23, 26).

TOTAL DISCHARGE OF THE SPRING SYSTEM

The total surface discharge of thermal water from the area included on plate 1 is summarized in table 36. Much additional discharge of thermal saline water from the same system escapes unseen into Steamboat Creek between this area and Huffaker Hills, 4½ miles to the north.

In table 37 we see that the cold springs of the region are all low in chloride. All spring waters with chloride content exceeding 100 ppm are thermal, and all of these are compositionally similar to Steamboat Springs and are located in the southern part of Truckee Meadows between Steamboat Springs and Huffaker Hills (table 38).

TABLE 36.—*Summary of surface discharge of thermal water from the immediate area of plate 1, Steamboat Springs*

	Discharge (gpm)	Cl (ppm)	Discharge [1] (gpm, calculated to 820 ppm Cl)
Average discharge from visible measured springs	47	899	52
Estimated discharge, visible unmeasured springs	13	~850	13
Unseen discharge into Steamboat Creek—determined by measurements of discharge and chlorinity of the creek [2]	~250	~850	260
Average discharge thermal wells Steamboat Resort	~60	~750	55
Average discharge thermal wells Reno Resort	~200	~980	240
Estimated discharge from wells, Mount Rose Resort	~20	~920	22
Total	590	--------	642

[1] Water samples from deep drill holes, summarized by depth intervals in table 35, indicate chloride content of 800–820 ppm in deep rising water before any boiling, evaporation, or near-surface dilution. This column therefore includes an estimate for discharge of steam, originally in the liquid water at depth.

[2] The northern limit of the immediate thermal area was taken for convenience to be the Virginia City highway bridge over Steamboat Creek near the north edge of sec. 28, T. 18 N., R. 20 E. (Thompson and White, 1964, pl. 2).

TABLE 37.—*Springs in the basin areas of the Mount Rose and Virginia City quadrangles, Washoe County*

Spring	Location and No.[1]	Estimated discharge (gpm)	Temperature (°C)	Cl (ppm)	pH	Specific conductivity, $K \times 10^6$
Bowers	16/19–3 BA–1	50–75	47	6	9.4	240
Heidenrich	16/19–21 AD–1	25	Cold	--------	------	--------
Greil	16/20–17 AA–1	8	15½	7	7.5	343
Sutro tunnel drainage	16/21–2 DA–1	100	--------	8	8.28	--------
Reeves	17/20–3 AD–1	20	19	6	7.74	398
Unnamed	17/20–3 DD–1	1	15½	5	7.68	348
Lone Tree [2]	18/19–10 AA–1	--------	15½	3	7.6	237
Page	18/19–11 AB–1	30	15½	--------	------	--------
Unnamed	18/19–11 CA–1	--------	Cold	3	7.46	324
Brookline	18/19–12 CA–1	70	16	--------	------	--------
Caffrey	18/19–12 DA–1	20	--------	--------	------	--------
Christman	18/19–12 DB–1	200	16	11	------	--------
Unnamed	18/19–14 BC–1	5	--------	6	6.86	341
Forest Service	18/19–22 DB–1	¼	12½	6	7.81	328
Huffaker	18/20–3 AC–1	5	27	420	7.1	--------
Double Diamond	18/20–9 BD–1	20	29	250	7.2	1,500
Dimonte drainage ditch	18/20–15 C	50	20	110	------	--------
Dimonte	18/20–16 DC–1	70	53	560	6.8	2,400
Zolezzi	18/20–17 DC–1	125	39	130	7.4	789
Steamboat Springs	18/20–28, 33	[3] 590	--------	~900	------	--------
Mesa	18/20–30 CB–1	Seep	Cold	--------	------	--------
Steamboat cold spring	18/20–33 DB–1	2	13	7	7.1	435
Unnamed	18/20–33 DB–2	5	Cold	7	7.3	520
Unnamed acid spring	18/20–35 AB–1	2	Cold	28	2.86	3,800
Lawton [2, 3, 4]	19/18–13 AC–1	--------	49	57	9.0	625

[1] Location designation system of Water Resources Division, U.S. Geol. Survey in Nevada (Cohen and Loeltz, 1964). The first three numbers as in 16/19–3 are, respectively, township, range, and section number. The first letter denotes quarter section in counterclockwise sequence starting with A as NE quarter; the second letter denotes quarter of quarter section determined by similar sequence; the final number is the specific spring or well recorded, in sequence for each quarter-quarter section.

[2] Data from Cohen and Loeltz (1964, table 5).

[3] Area of pl. 1.

[4] Located 5 miles west of center of Reno on Truckee River.

The actual composition of the unseen water no doubt differs from place to place, but it is probably within the range of visible springs such as Dimonte, Zolezzi, and Huffaker (tables 37, 38), with chloride contents of 110–560 ppm, averaging 250–300 ppm. The B:Cl ratios of table 38 are particularly significant because these two elements are so characteristic of thermal waters of volcanic association (White, 1957a, b) and because they are least likely to be affected by chemical precipitation, base exchange reactions and other factors. The B:Cl ratio in Steamboat Creek at Huffaker Hills is almost identical to that of Steamboat Springs and of the other analyzed high-chloride hot springs. Ratios of other constituents included in table 38 show somewhat greater differences, but all are within the ranges to be expected from slightly different degrees of base exchange and other reactions with the basin sediments.

The total thermal water inflow to Steamboat Creek can be determined, using chloride as a tracer, by measuring discharge and chloride content of Steamboat Creek above the springs and comparing these with similar measurements at the narrows through Huffaker Hills (SW1/4 sec. 34, T. 19 N., R. 20 E.; Thompson and White, 1964, pl. 2; Cohen and Loeltz, 1964, pl. 2).

Measurements of this type, to be reliable, must be made at a time that is as free as possible from complications related to temporary storage of salts from evaporation, removal of stored salts by precipitation or very high runoff, and filling or emptying of the large swimming pool at the Reno Resort. For all of these possible complications, measurements from late fall through early spring are likely to be much the best. One particularly good series of measurements was made on April 12, 1955. The preceding 20 days were without trace of precipitation at the Reno weather station and daily average temperature had been about 50°F, which was too low for significant evaporation, temporary storage, or unusual runoff. The discharge of Huffaker Narrows was measured by current meter by Don Clendenon, Water Resources Division of the Geological Survey, and was 7.43 cfs, or 3,330 gpm. The chloride content of a water sample collected at the same time was 282 ppm. In view of the fact that all cold surface waters and springs of the area are low in Cl (table 37), and all major streams unaffected by the Streamboat thermal system average less than 4 ppm, about 278–282 ppm of the Cl of Steamboat Creek at Huffaker is assumed to be contributed by saline thermal inflow. Furthermore, we will assume that all this thermal inflow is from the Steamboat Springs system, and has an average chloride content of 820 ppm deep in the system (from table 29). With these assumptions, the inflow (X) is calculated from the formula (X) (820 ppm) = (3,330 gpm) (282 ppm–4 ppm). The rate of inflow, then, is equivalent to 1,130 gpm of water containing 820 ppm of chloride.

Another series of measurements was made on April 29, 1964, with the help of J. E. Parkes, Water Resources Division. A light snowstorm had occurred on April 22 and 23, resulting in 0.32 inch of precipitation at Reno. The calculated discharge of thermal water from the

TABLE 38.—*Summary of surface discharge of chloride in water from Truckee Meadows south of Huffaker Hills*

Site	Location	Concentration Cl (ppm)	K/Na	Li/Na	B/Cl	HCO_3/Cl	SO_4/Cl	Approximate discharge (gpm)	Discharge (gpm, calculated to 820 ppm Cl)
Known springs:									
Steamboat Springs	Area of pl. 1.	600–1,000	0.10	0.010	0.056	0.4	0.15	590	640
Dimonte	(1)	560	.09	.011	.054	.65	.14	70	48
Zolezzi	(2)	130	.14	.011	.049	1.82	.13	125	20
Huffaker	(3)	420	.09	.007	.052	.64	.37	5	3
Dimonte drainage ditch.	(4)	110						50	7
Double Diamond	(5)	130–310 (assumed avg 250).						20	7
Total discharge of known high-chloride spring water.								860	725
Stream discharge:									
Dilute creeks	Surrounding area.	<1–9	~.29	~.01	~.017	~50	~2	~2,500 avg	~20
Steamboat Creek, Apr. 12, 1955.	Huffaker Hills.	282			.053	1.184	.336	(3,330)	1,130
Net unseen spring discharge.	Outside area of pl. 1.	Varied dilutions largely to 100–600.						Unknown	~400

[1] On Steamboat Creek 1¼ miles north of area of pl. 1; SE¼ sec. 16, T. 18 N., R. 20 E., Virginia City quadrangle.
[2] 1¼ miles north-northwest of area of pl. 1; near midpoint of south edge sec. 17, T. 18 N., R. 20 E.
[3] On banks of Steamboat Creek east and southeast of Huffaker Hills 4 miles north of area of pl. 1.
[4] Springs and seeps south half sec. 15, T. 18 N., R. 20 E.
[5] 2¾ miles north of area of pl. 1; NW¼ sec. 9, T. 18 N., R. 20 E.

system (with assumed Cl content of 820 ppm) was equivalent to 1,385 gpm. Any anomaly from the storm of April 22 and 23 would probably indicate too high a rate of discharge of Cl from previously stored salts, and consequently, too high an estimate for total discharge of thermal water. The 1955 series of measurements is selected as the more conservative in later calculations, but it is not necessarily the more accurate.

Of the total in April 1955, about 640 gpm was discharged directly into Steamboat Creek from the immediate Steamboat system (area of pl. 1; table 36), and the equivalent of an additional 85 gpm is assumed from known visible springs in the southern part of Truckee Meadows south of Huffaker Hills (table 38). About 400 gpm of additional saline water (calculated to 820 ppm of Cl) flows unseen into the bed of Steamboat Creek between the area of plate 1 and Huffaker Hills.

Actual total rate of discharge of the unseen springs is probably about 3×400 gpm, or 1,200 gpm, but in later heat-flow calculations (p. C98–C103) the discharge of 400 gpm, calculated to 820 ppm Cl, prior to near-surface changes is much more significant.

All this saline thermal water is presumed to rise within the Steamboat system, escaping northward and diluted below the surface, eventually discharging as visible warm springs and as unseen discharge directly into the bed of Steamboat Creek.

No positive evidence was found in Truckee Meadows south of Huffaker Hills for any hot-spring system that is entirely independent of Steamboat Springs. Two small geothermal systems without natural visible discharge contain dilute water that may be entirely of meteoric origin, but their supply of heat is likely to be closely related to the Steamboat system. One of these systems is about 1½ miles northwest of the Main Terrace (table 39, Nos. 68–70); the second system is on the south border of Steamboat Valley, 1¼ miles south of the Steamboat Resort (table 39, Nos. 7–11).

Just southwest of Reno the Moana geothermal system also lacks natural flowing springs and is the source of hot artesian water for the Moana Springs resort and for local farm houses (table 39, Nos. 110–132). The hot water is localized in the NE¼ sec. 26 and NW¼ sec. 25, T. 19 N., R. 19 E., about 7 miles northwest of the Main Terrace. The Moana system seems to be entirely independent of Steamboat Springs, except for the probability that the ultimate source of heat is related to that of the Steamboat system.

TABLE 39.—*Depths and temperatures of some wells in the Mount Rose and Virginia City quadrangles and adjacent areas, and other data*

[Compiled from data prepared by Water Resources Division and measurements by D. E. White. Water samples from top of column]

Well				Reported depth (feet)	Bottom temperature (°C)	Other temperature (°C); remarks	Cl (ppm)	pH	Specific conductance (micromhos, 25°C)
No.	Location No.[1]	Location	Name						
1	16/19-21AA	Southwest of Washoe Lake	Miller	235	17				
2	17/19-9AB1	Head of Galena Creek	Sinai	115	10				
3	17/20-3CC1	East side Steamboat Valley	Taylor	[2] 32.4	16		7.2	7.68	452
4	4AB1	Steamboat Valley	Neilson	35		11; flowing			
5	4BA1	do	Johnson	[2] 67.15	19.7	19.8; surface	4.0	7.77	633
6	4BD1	do	Peterson	45	12		4.4	6.66	390
7	4DC1	do	Tachino	[2] 52.3	36.8	34.7; surface	3.9	7.63	256
8	4DC2	do	do	[2] 24.7	36.1	22.2; surface			
9	4DC3	do	do	[2] 21.6	27.4				
10	4DC4	do	do	[2] 14.8	22.4		5.6	7.25	256
11	9BA2	do	Neilson	56		32; flowing	5.6	8.18	251
12	17/21-6AC1	Virginia Range	Young drill hole	662		17.2 at 364 ft	9.4	7.08	2,650
13	18/19-12CA1	Truckee Meadows	Kuser	[2] 150	14	14.2 at 145 ft 6 months after completion.			
14	12DA3	do	Caffrey	[2] 165	18	17 at 30 ft; 16.5 at 50 ft			
15	12DB1	do	Christman	[2] 301	20	17; surface when pumping.	8		
16	12DB2	do	do	[2] 376.5	21	14.9 at 47 ft; 15.2 at 73 ft	8.0	7.75	256
17	13AD1	do	Ramsey	127	11		9		
18	18/20-6AC2	do	Jaksich	[2] 130	14.5				
19	7CB1	do	Herz	109	11				
20	8DC1	South Truckee Meadows	Keithly	100+		17 at 100 ft	7.0	7.08	191
21	9BD1	do	Double Diamond ?	(?)		21; flowing	133	7.18	751
22	9BD3	do	do	[2] 29.3	29	29; flowing	310	7.24	1,500
23	14BB1	East side Truckee Meadows	LeCroix	[2] 163	[2] 30.7		680	6.55	2,680
24	14BC1	do	Davidson	[2] 60	29.5	24.5; from tap	509	7.38	2,325
25	16BD1	South Truckee Meadows	Humphrey	[2] 38.9	16.5		6.0	7.47	313
26	17CA1	do	(?)	[2] 137	13.5				
27	17DC1	do	Eccles	80	20		6.2	7.67	387
28	17DC2	Zolezzi Springs	do	[2] 99	45.5		86	7.57	623
29	17DC3	South of Truckee Meadows	Berns	54	14		3.0	7.37	242
30	18AB1	do	Jenkins	[2] 105	16.5				
31	18AB2	do	Griffin	[2] 130	16.5				
32	19AC1	do	Field	[2] 191.2	11				
33	20AA2	do	Snook	71	15+	15 at tap	30	6.80	293
34	20AA3	do	do	100	19	19; pumping	26	7.13	333
35	20AA4	do	Watts	70	11.5	11.5 at tap	4.6	6.91	184
36	20AB1	do	Hall	[2] 73.3	21+	21 at tank	11.6	8.01	308
37	20AD1	do	Cicchese	[2] 86	17		2.6	7.38	201
38	20AD2	do	Damonte	[2] 22.8			6	7.03	211
39	20AD3	do	Dawson	[2] 84.7	19.5		3.8	7.13	213
40	20BA2	do	Mulert	35+	10.5		3.2	7.48	283
41	20BA5	do	Stevens	[2] 96	19.5		9.6	7.34	395
42	20BB1	do	Zolezzi	107	12				
43	20BB4	do	Enholder	107		10 at tap	4.0	7.43	379
44	20DB1	do	Pincolini	[2] 87	19.5		12.0	7.10	197
45	20DD1	do	Herz	[2] 66	23.5		6.0	7.84	282
46	21CA1	do	Damonte	10			3.0	6.60	128
47	21CA2	do	Sutherland	[2] 93	21.5				
48	21CA3	do	McComas	[2] 44	21		10.0	6.96	214
49	21CA4	do	Sutherland	[2] 55	48	36.5 at 40 ft; 44.5 at 45 ft			
50	21CA5	do	do	[2] 55	30	29 at 42 ft			
51	21CA6	do	do	50	25.5	24.5 at 40 ft			
52	27CA1	Southeast of Truckee Meadows	McKnight	[2] 192.2	28.5	24.3 at 124 ft			
53	27CC1	do	Isbell	[2] 134.7	28.1				
54	28AB1	Near Steamboat	Warren	[2] 63.1	37				
55	28AB2	do	Murray	680		37.2 at 63.3 ft	508	6.15	
56	28AD1	do	Damonte	77	19	19 at 62 ft			
57	28BA2	Steamboat	Mount Rose 1	[2] 110	[2] 121.3		[3] 824		
58	28BA3	do	do	[2] 159.6	[2] 133	86.4 at 53.6 ft	[3] 844 [4] 896	[3] 7.38 [4] 8.55	[3] 3,400 [4] 3,315
59	28BA4	do	do	[2] 133.2	[1] 119.2	76.5 at 53 ft	[3] 844	7.38	3,400
60	28BB1	do	Harold Herz: 1	[2] 154.5	93.2	47.8 at 56 ft	[3] 416	7.1	1,820
61	28BB2	do	2	[2] 153	89.6		400	7.0	1,745
62	28BB3	do	3	[2] 23.3	25.8				
63	28BD1	do	East Reno	[2] 156.5	137.7	76 at 8 ft	[3] 868	7.01	
64	28BD2	do	West Reno	[2] 186	137.5	96.5 at 30 ft	[3] 898	7.46	
65	28BD3	do	Reno	[2] 34.3	93.8				
66	28CB1	do	Senges	[2] 177	145.5	96.3 at 31 ft	[3] 826 [4] 882	6.40 6.67?	3,440 3,810
67	28CD1	do	Nevada Thermal Power Co. 2.	964	(5)		[4] 955	8.5	3,770
68	29AA1	Near Steamboat	Peigh	[2] 57	26.1				
69	29BA1	do	do	[2] 73.7	27.1	24 at 44.7 ft	6		
70	29BC1	do	do	[2] 82.2	33.9	32.0 at 47 ft	2.6	7.65	194
71	29DA3	Steamboat	GS-2	[2] 398	154.4	161.4 at 303 ft	[3] 564	5.98	2,730
72	29DD1	do	GS-6	[2] 212	102.5	85.5 at 143 ft	12	6.72	372
73	30CA1	Southwest of Truckee Meadows	Anderson	[2] 68.5	<14				
74	30CA2	do	do	[2] 243.1	14.2	13.1 at 235 ft	2.2	6.99	485
75	32AB1	Steamboat	Mercury well	[2] 129.4	102.8	96.1 at 106 ft	600	7.8	
76	32AB2	do	Cox	[2] 93	94.2				
77	32AB3	do	Nevada Thermal Power Co.: 3	1,263	(5)		[4] 1,080	8.4	4,250 (erupting)
78	32BA4	do	6	716	179				
79	32BA1	do	4	[2] 520 [2] 726	~163 186		[3] 45 [4] 874	7.2 8.2	1,950 3,430
80	32BB1	do	Reno Press Brick	[2] 58.0	70.0				
81	32BB2	do	Nevada Thermal Power Co. 5.	[2] 826	163	175 at 719 ft			

See footnotes at end of table.

TABLE 39.—*Depths and temperatures of some wells in the Mount Rose and Virginia City quadrangles and adjacent areas, and other data*—Continued

Well				Reported depth (feet)	Bottom temperature (°C)	Other temperature (°C); remarks	Cl (ppm)	pH	Specific conductance (micromhos, 25°C)
No.	Location No.[1]	Location	Name						
82	32DB1	Steamboat	Cox	[2] 14.5	57.8				
83	32DB3	do	GS-7	[2] 407.8	161.0		0.5	6.66	158
84	33AB1	do	GS-8	[2] 121.8	128.7	80 at 25.6 ft	896	6.55	3,300
85	18/22 33AB2	do	Nevada Thermal Power Co.	1,830	([5])		870	8.4	3,560
86	18/20-33BA1	do	Rodeo	[2] 276.9	168.8	86 [3] at 7.6 ft	836	7.02	3,440
87	33BA2	do	GS-3	[2] 683.6	164.0	169.1 at 530 ft	791	6.60	3,280
88	33BA3	do	GS-4	[2] 503.0	171.1		816	6.55	3,180
89	33BA4	do	GS-5	[2] 572.2	169.6	172.2 at 543 ft	826	6.00	3,270
90	33DA1	do	Knox	[2] 6	13.5				
91	33DA2	do	do	[2] 56	29.9				
92	33DB	do	Carver	[2] 28.3	94.8	92 at 14.6 ft			
93	33DB2	do	Steamboat 2						
94	33DB5	do	Steamboat 4	[2] 184	154.8	96 at 17 ft	702		2,700
95	33DB6	do	GS-1	[2] 398.4	156.9	158.1 at 193 ft	817	6.05	3,250
96	33DB7	do	No. 32 Geyser well	[2] 43.2	114.6		[3] 885	7.4	3,335
							[4] 986	8.7	3,810
97	33DC1	do	South Steamboat well	[2] 258	75.5		10	7.43	348
98	33DC2	Steamboat Valley	Carver	[2] 77	19.5		6	7.08	298
99	34AC1	Southeast of Truckee Meadows	Johnson	[2] 136.5	21.8	17.9 at 62 ft	12	7.20	1,947
100	34CD1	do	Frazier	[2] 160	22.0		8.4	8.03	367
101	34DB1	do	Schindler	[2] 125.4	19.1		5.6	7.23	583
102	10DB1	West of Reno	(?)	365	13		35		
103	11DB4	Reno	Magnin	140	12				
104	11DB5	do	Johnson and Davidson	190	13		14		
105	13BB1	do	Crystal Springs Ice Co.	118	14				
106	13BC1	do	Sierra Pacific	785	29+	29 in pumped water	30		
107	14AD1	do	Diamond Springs Water Co.	210	18+	18 at tap	12		
108	21CA1	Southwest of Reno	Caughlin oil test	1,865	(?)	"hot water reported at 1,200 ft (Anderson, 1910).			
109	23DC1	South of Reno	Country Club	140	(?)	Reported "hot"			
110	25AB1	do	Moffat	662	41.5	41.5 at tap			
111	25AB2	do	do	500	40.5	40.5 at hose	26	7.85	836
112	25BA2	do	Moana Springs	571.45	53.0	40; flowing			
113	25BA3	do	do	(?)	51+	50.8; flowing	28	8.13	832
114	25BB7	do	Smith	60	31.5+	31.5 at hose	29	7.26	944
115	25BD1	do	Van Slyck	77	33.5	22 at 36 ft; 29 at 72 ft			
116	25BD2	do	Randall	260	45.5				
117	25BD3	do	do	111	45.5				
118	25BD4	do	Johnson	95	44.5				
119	25DA1	do	Pecknam	700	24.5	24.5 when pumped			
120	25DC1	do	University farm	500		Reported "hot"			
121	26AA2	do	Kimberly	155	41				
122	26AA3	do	Martic	600	76.5				
123	26AC1	do	Erskine	155	25.5				
124	26AD1	do	Moana	179	88.5		55		
125	26AD2	do	do		91+	90.8 at leak in discharge pipe.	52	7.94	1,320
126	26AD3	do	Yates	[2] 184.2	71				
127	26AD4	do	do	[2] 168.3	71				
128	26AD5	do	do	[2] 197.5	75.5				
129	26AD7	do	Frey	464	82+	Discharge, 82			
130	26DA2	do	Kelty	200	91				
131	26DB1	do	do	104	95.5	5 gpm discharge, 86.5	48	8.33	1,384
132	26DC1	do	Campbell	750	82+	82 at leak in discharge pipe.	48	7.89	1,327
133	36BB2	do	Allison	197	15+	15; pumped			
134	36BB3	do	Morgan	175	14+	14; pumped			
135	36DA2	do	Del Monte	400		13.5; pumped			
136	36DA3	do	Giroux	536		11.5; pumped largely from 21 ft.			
137	36DC2	do	Eddy	129	14.5				
138	36DC3	do	Ghiglieri	142	13.5		17		
139	19/20-19DA1	Southeast of Reno	United Airlines	116	14.5+	14.5; pumped			
140	19DCA1	do	Preston	29.5	13.5+	13.5; flowing			
141	19DCB3	do	Frazier	45	14+	14; flowing			
142	19DCC2	do	Holmes	53	12+	12; flowing			
143	19DCC3	do	Summers	20.4	13.5+	13.5; flowing			
144	19DCD1	do	Breaker	27.1	13.5+	13.5; flowing			
145	27CA1	East side Truckee Meadows	Birbeck	80	16		284	7.65	1,660
146	29AB1	Southeast of Reno	Niturnio?	68	14+	14; flowing	6.0	7.26	458
147	30BC1	do	Newton	360	29+	29; flowing			
148	30BD9	do	Allard	43	14+	14; flowing			
149	30CC1	do	Fife	400	26.5+	26.5; flowing			
150	30DA1	do	Model Dairy	220	15.5+	15.5; flowing			
151	30DB2	do	Sanderson	70			3.5	6.83	296
152	31AC1	do	Lehnert	143	16+	16; flowing			
153	31AC2	do	DeLucchi	167	16.5				
154	31BC2	do	Pollard	115	12				
155	31DA1	do	Capuro	139	14.5+	14.5; flowing			

[1] See footnote 1, table 37; denotes, in order, township, range, section, quarter section, quarter-quarter section, and sequence in quarter-quarter section.
[2] Measured depth.
[3] Nonerupting.
[4] Erupted sample.
[5] Not measured.

Thus, nearly all chloride and boron supplied to Steamboat Creek south of Huffaker Hills is closely related to the Steamboat system. The average discharge from this system is at least 1,100 gpm of water, calculated to a chloride content of 820 ppm. On this basis, the average visible discharge of the flowing springs (65 gpm) is only 6 percent of the total discharge from the system (table 40).

TABLE 40.—*Inventory of total discharge of Steamboat Springs thermal system in southern Truckee Meadows, calculated to chloride content of 820 ppm*

	Discharge (gpm)	Percent of total
Steamboat, measured springs (table 36) including steam calculated as water.	52	4.6
Steamboat, visible unmeasured springs (table 36).	13	1.2
Erupting thermal wells, including net excess condensed steam (table 36).	~315	28.0
Unseen discharge, area of pl. 1 (table 36) [1]	260	23.1
Visible springs north of pl. 1 (table 38)	85	7.6
Unseen discharge into Steamboat Creek north of pl. 1 (table 38; data of Apr. 12, 1955).[2]	~400	35.5
Total hot spring discharge, calculated to 820 ppm of chloride.	~1,125	100.0

[1] Northern limit taken to be at the Virginia City Highway bridge over Steamboat Creek, near north edge sec. 28, T. 18 N., R. 20 E. (Thompson and White, 1964, pl. 2).
[2] Northern limit at narrows of Steamboat Creek through Huffaker Hills, SW¼ sec. 34, T. 19 N., R. 20 E.

Present production from geothermal wells accounts for nearly 30 percent of the total, or an average of about 300 gpm. Evidently, much or all of this supply was diverted from formerly more active natural visible springs, and the unseen springs that discharge directly into Steamboat Creek.

TEMPERATURE RELATIONSHIPS AND CIRCULATION PATTERNS OF INDIVIDUAL AREAS

LOW TERRACE

SPRINGS

The locations, temperatures, and discharges of the natural springs are considered elsewhere. No active springs now exist near the crest of the southern half of the terrace (pl. 3), probably because of diversion of water supply to the producing Steamboat wells. Near the central and northern axial crest of the terrace, spring temperatures are near boiling and most of the active vents discharge as geysers or boiling springs (table 4); the only consistent discharge from this axial area has occurred in the northern cluster containing springs 25 and 26, but their total discharge is only a few gallons per minute. In contrast, most of the natural discharge from the Low Terrace is from springs of moderate temperature on the low eastern flank of the terrace and unseen vents in the bed of Steamboat Creek. Cooling by dilution is only a minor factor. Because these springs are marginal and horizontal temperature gradients are high, more heat is lost by conduction than from the centrally located channel of upflow that are surrounded entirely by hot ground.

DRILL HOLES

Wells drilled on the Low Terrace show pronounced changes in temperature and salinity with depth (figs. 24, 25, and 31). The degree of detail that can be deciphered depends very much on method and rate of drilling; at best, only a part of the original differences existing in the ground before drilling is shown in these figures. The circulation pattern is exceedingly complex in detail. The upflowing and outflowing waters tend to be hottest and most saline, and the downflowing and inflowing waters tend to be relatively cooler and less saline, but some exceptions exist.

In drill hole GS–1 (fig. 31), water relatively high in temperature and chloride content characterizes a middle zone at depths of 140–220 feet and a bottom zone near 400 feet. A minor zone of moderate salinity exists near 100 feet, overlain and underlain by less saline zones that have slightly depressive effects on temperatures. A zone of water of only moderate salinity occurs near 240 feet in depth, but associated original ground temperatures were not well defined. A temperature reversal almost certainly occurred here, but satisfactory equipment for measuring temperatures below a reversal was not at first available. When an uninsulated maximum thermometer is lowered in a well, the thermometer registers the maximum temperature attained regardless of depth (see data, table 27). If such a thermometer is placed in a very well insulated receptacle, it can be lowered rapidly through a hot zone, but held in a lower cooler zone long enough to attain equilibrium, and then withdrawn rapidly. The measured temperature of 154°C at 365 feet in depth was determined by an insulated thermometer, but lower ground temperatures probably existed near 240 feet and 325 feet.

Steamboat well 4 (fig. 25; table 19) penetrated aquifers characterized by temperature minima near 58 and 105 feet in depth. Minima also occurred in chloride content of water from the same depths. Zones of high-chloride water at relatively high temperature occur near 50 feet and 70–80 feet; the very high temperatures near the bottom of the well are probably associated with high-chloride water that is diluted greatly by water from the major aquifer near 105 feet.

The South Steamboat well (fig. 24; table 18) is 400 feet southwest of drill hole GS–1 and about 200 feet south of the southern limit of hot spring sinter. Temperatures are far below boiling—about 75°C below at

25 feet in depth, 80°C at 120 feet, and 95°C at 240 feet. The low temperatures are obviously related to the marginal position of this well relative to the Low Terrace subsystem.

As in other wells of the Low Terrace, there is a close but no precise relation between chloride content of the water and relative temperature in the upper part of the hole. The high-chloride zones are generally relatively high in temperature and the low-chloride zones, low.

Outstanding features of this well are: 1. Temperatures are low compared to those of the Steamboat Resort wells and GS–1. 2. A major zone of saline water occurs between the surface and 160 feet in depth. 3. Zones of less saline water occur within this major zone. 4. Water below about 160 feet in depth is very low in chloride and is interpreted to be entirely meteoric in origin; as the well is drilled, some time is required to flush out high-chloride water already introduced into the well from a higher zone. The well is cased to 188 feet; samples from greater depth decreased in chloride to only 8 ppm, which is very little more than normal Steamboat Creek water (average about 4 ppm).

GEOPHYSICAL DATA

Geophysical resistivity measurements were very useful in determining the limits of the thermal system south of the Low Terrace because hot saline water is a very good electrical conductor compared to cold dilute water (White and others, 1964, p. 1359, 1369). Specific conductance of normal saline spring water at 25°C is about 10 times higher than that of water from Steamboat Creek and South Steamboat well. This contrast in conductance is magnified much more by the natural temperatures of the environment.

Resistivity measurements near the south end of the Low Terrace (White and others, 1964, pl. 4) show a well-defined resistivity low that coincides with the axis of the Low Terrace, extending southwest from the hot springs and surface sinter. The contrast in resistivity is particularly evident in the curves of 100-foot electrode spacing, commonly called 100-foot depth, but is less pronounced in the curves for 200-foot spacings (200-ft depth). The contrast in salinity and perhaps temperature is evidently less at the greater effective depth. This is consistent with the evidence from the South Steamboat well, which proves the deeper water to be lower in salinity; temperatures are also much lower than in drill hole GS–1 to the north. The minimum resistivity values increase southwest from about 600 ohm-cm on the traverse through drill hole GS–1 to 1,300 ohm-cm on the traverse through South Steamboat well (White and others, 1964, pl. 4), and 1,900 ohm-cm on the traverse 200 feet farther to the south, indicating that temperatures and salinity continue to decrease southward, as expected from the distribution of hot springs and sinter. The subsurface limits of the thermal area can be determined much better from the resistivity data than from surface evidence, and the strike of the controlling structure is evidently to the southwest rather than to the south, which seemed equally probable from the regional mapping (Thompson and White, 1964, pl. 2; p. A–38). In all these traverses the minimum resistivity occurs in the 100-foot spacing curves, thereby indicating the presence of near-surface matter of high conductivity. The South Steamboat well has shown this to be water of high salinity, underlain by other water of much lower salinity.

CIRCULATION WITHIN THE TERRACE

Hot saline water rises toward the surface along or near the principal fault zone of the Low Terrace. The andesitic tuff-breccia dike penetrated by drill hole GS–1 is one of the chief hot-water aquifers of the system, particularly in the middle and lower parts of the hole. The highest temperature measured in the terrace is 158°C in tuff-breccia at a depth of 193 feet. A temperature nearly as high, 154.9°C, was found at 184 feet in Steamboat well 4. Both temperatures were very close to the boiling points for prevailing pressures.

Very little water is discharged at the surface directly from faults or fissures of the Low Terrace. As the hot water approaches the surface, it is diverted horizontally into permeable sinter and sediments along the walls of the fissures. This water, as well as other saline water rising into the margins of the terrace, is cooled considerably by conduction before being discharged at only moderate temperatures. Saline water in the upper part of the South Steamboat well is evidently flowing nearly horizontally to the south along and near the Steamboat fault zone, eventually discharging into Steamboat Creek. Its temperature decreases to the south because heat is being conducted to the surface and also to the cooler meteoric water flowing northeastward into the spring system at higher and lower levels. The isotope data summarized on pages C13–C15 show that this meteoric water is not directly from Steamboat Creek, except in small part, but probably is from streams on the flanks of the Carson and Virginia Ranges, as suggested in figure 4.

MAIN TERRACE

SPRINGS

The hot springs of this terrace have been discussed in previous sections but are reviewed briefly here. Most of the springs of highest temperature and discharge issue from open fissures near the crest of the terrace (pl. 3; table 4) at altitudes above 4,660 feet. This includes most but not all vents that have discharged as geysers. Springs on the margins of the terrace generally discharge at temperatures below 90°C, and a few are below 80°C. Of the latter, 19w, 20n, and 21s are in the northwestern part of the terrace, and 12sw, 3, 10, 52, 4, and 51 are on the eastern slope. The centrally located springs are usually close to boiling even if discharge is small, because the channels of rising water are surrounded by very hot ground and little heat can be lost by conduction except very close to the surface; as the water rises and pressure decreases, steam is formed, so that all temperatures must be close to the boiling-point curve. In contrast, water in marginal channels can lose much heat by conduction. In marginal springs, temperatures are generally very low when discharge is low. When discharge increases, a smaller proportion of the total transported heat can be lost by conduction, so that temperature of discharge then rises. A good example of these principles in operation is the record of spring 2 (fig. 19), which shows an abrupt 150 percent increase in discharge following the earthquakes of September 7, 1947. Why, on the other hand, did the corresponding decrease in discharge of nearby spring 5 result in no significant decrease in temperature? Spring 5 had been a boiling spring, and its deep channels were already at boiling temperatures for prevailing pressures. The abrupt decrease by nearly three-fourths of its former discharge probably resulted in some decrease in the proportion of water converted to steam, not accompanied by any significant change in discharge temperature, because temperatures remained on the boiling-point curve.

DRILL HOLES

Most of the available information on the subsurface of the Main Terrace was obtained from the Rodeo well and GS–3, GS–4, and GS–5 (pl. 3). A private company drilled a well just east of GS–3 to a depth of 130 feet in June 1954; Nevada Thermal Power Co. 1 was drilled to 1,830 feet just east of the Main Terrace, and Nevada Thermal Power Co. 2 was drilled near the center of the terrace to 964 feet in 1959. These three wells were drilled when the author was engaged in other work, and no other data were obtained.

Data from the Rodeo well are shown in figure 26 and table 20; a detailed log was published by White, Thompson, and Sandberg (1964, table 3). The well was collared only 12 feet from a fissure containing boiling water, but measured temperatures were much below the boiling-point curve for the first 100 feet of depth. This is related to heat loss by conduction from the high-chloride zone at 23–46 feet deep; heat is being conducted upward to the surface, where the temperature is less than 20°C, and downward to an underlying low-chloride zone that is relatively cool for its depth of 50–82 feet. The latter indicates that some meteoric water is migrating into the system, presumably from the small drainage area west of the terrace. Another minor zone of less saline water is near 120 feet in depth; oxygen dissolved in this water had oxidized pyrite to such an extent that water and mud erupted from this zone was red brown. The relatively high bottom-hole temperatures measured near this zone indicate that rate of inflow of low-chloride water is very low, at least in recent times. A major aquifer from about 130–150 feet contains high-chloride water at temperatures that are very close to the boiling-point curve. At still greater depth from about 154 feet to the unconformity at 166 feet, an impressive zone of cooler less saline water must be migrating inward into the system. The chloride content of the water column (as sampled at water level) was only 10 percent less than at higher levels, but the actual chloride content of the deep aquifer is probably considerably less than the 520 ppm of the sample of water diverted into a side pipe during the induced eruption of May 20 (table 20). The rate of movement of this water of probable mixed origin was sufficiently great to depress the temperature about 30°C below the boiling-point curve at this depth. The relatively low temperature and low chloride content is best explained if a considerable proportion is local cold meteoric water. The most reasonable source is precipitation in the small drainage area west of the well.

The granodiorite is weathered and perhaps slightly permeable just below the unconformity at 166 feet in depth. The rock was sufficiently soft to be drilled by rotary rock bits from 166–180 feet, but drilling then became so slow that the well was completed with diamond coring bits. No faults, veins, or permeable zones were identified from the drill core; this is evidence that the well erupts nearly all its water supply from depths of less than 170 feet. The extremely slow rate of attainment of temperature equilibrium at the bottom of the drill hole after the induced eruption of May 20, 1950 (table 20), is evidence for absence of a high-temperature aquifer near the bottom of the hole.

Drill hole GS–3 (table 29; fig. 33), just west of the fissures of the Main Terrace, was designed to search for a west-dipping thrust fault as the possible principal structure of the area (White and others, 1964, p. B48–B50), but no evidence for thrusting was found. Temperatures and intensity of rock alteration decreased somewhat near the bottom of the hole, and the rocks were entirely granodiorite below 50 feet.

Numerous faults and breccia zones provide structural control for migrating thermal water and for rock alteration (Schoen and White, 1965), but veins are scarce compared to their frequency in drill holes GS–4 and GS–5 east of the crest of the terrace (see structure section, pl. 2, and logs of holes in White and others, 1964, table 3).

The chloride content of all water samples from GS–3 are below 800 ppm and are slightly below most other drill holes of the Main Terrace, including the Rodeo well and GS–4, GS–5, and GS–8 (table 35). This fact, along with data on temperature, structure, and alteration, indicates that the fissures cut by GS–3 are not so closely related to the main channels of upwelling water as the fissures cut in GS–4 and GS–5.

Data from drill hole GS–4 are shown in table 30 and figure 34; the detailed log of the hole was published by White, Thompson, and Sandberg (1964, table 3). Temperatures increase steadily to 170°C at a depth of 338 feet. From 100–280 feet, all temperatures are within 10°C of the boiling-point curve despite the fact that most measurements were obtained only 15 hours after previous drilling. Repeat measurements at a depth of 166.7 feet (table 30) at intervals of ½, 1½, and 2½ days after the last previous drilling showed differences of only 3.5°C, which indicated a close approach of temperature to equilibrium nearly on the boiling-point curve (fig. 34). Presumably, original ground temperatures at other depths from 100 to about 280 feet were also very close to, but slightly below, the boiling-point curve for pure water, but the data do not support a similar relationship for shallower and greater depths.

In contrast to the Rodeo well and GS–3, there is no evidence that GS–4 intersected low-temperature or low-salinity aquifers. The deep water (table 35) has a chloride content of about 800 ppm, or slightly higher than the average of about 780 ppm indicated for GS–3.

Chalcedony and calcite veins and moderate to intense rock alteration are much more abundant (Schoen and White, 1965) than in drill hole GS–3 and the Rodeo well. This fact, in addition to the temperature and salinity relations just discussed, indicates that the fissures controlling upwelling of thermal water dip eastward from the crest of the terrace and are intersected by drill hole GS–4, as shown in one structure section of plate 2.

Drill hole GS–5 (table 31; fig. 35) is one of the most significant of the area. The detailed log was published by White, Thompson, and Sandberg (1964, table 3). Temperature relations are very similar to those of GS–4. From 100 to about 280 feet in depth, temperatures are very close to the boiling-point curve, but repeat measurements at 175 and 236 feet indicate that equilibrium temperatures are slightly below the theoretical curve, as suggested in GS–4. The inflection at 157 feet, unchecked by repeat measurements, could result from greater-than-normal cooling from drilling of the previous day. It should be noted here that the amount of cooling during drilling and the rate of attainment of temperature equilibrium during a rest period are strongly dependent upon local bottom-hole conditions. For example, equilibrium is attained rapidly when the hole is bottomed in or very near an aquifer controlling a vigorous through-circulation of water. Attainment of equilibrium is obviously much slower when a large volume of adjacent rock has been cooled by drill water and must then be reheated entirely by rock conduction. For this reason, when the depth accessible to the thermometer is considerably less than the drilled depth, equilibrium is generally attained slowly.

Temperatures in GS–5 decreased slightly below 380 feet in depth. Most of the measurements of table 31 were made with an uninsulated maximum thermometer, but a decrease of 2½° C from the highest reading is indicated on October 23 when the thermometer was insulated. This may indicate, as shown in the structure sections (pl. 2), that GS–5 has penetrated almost to the footwall of the main system of east-dipping fissures controlling the upwelling water.

Drill hole GS–8 (fig. 38) is shown on the same structure section as GS–3 and GS–4 (pl. 2) but is farther to the east; its detailed log was published by White, Thompson and Sandberg (1964, table 3). Water was first found at a depth of 27 feet, which is 24 feet farther below the ground surface than in GS–4. The water table therefore slopes more steeply to the east than the topographic surface (pl. 2). The hydraulic gradient at the water table is about 80 feet vertically in the horizontal distance of 300 feet between GS–4 and GS–8, and is 55 feet vertically in the horizontal distance of 150 feet between spring 43 and GS–8. Permeability adjacent to the fissure that supplies springs 6, 7, and 43 is evidently very low, probably because pore spaces have been filled by silica minerals. In contrast, the water level in GS–8 is only 15 feet above Steamboat Creek, which is 450 feet farther to the east; permeability between GS–8 and the creek is evidently relatively high. Temperatures are significantly lower in GS–8 (max, 129°C at 122 ft) than in GS–4 (141°C at 130 ft) but relative to the water table, temperatures in GS–8 are only slightly lower.

The fact that temperatures in GS–8 are similar to those of the upper part of GS–4, which cuts the principal channels of upflow, is evidence that the rocks of GS–8 are not heated entirely by rock conduction. The temperature relationships and hydraulic gradients are best explained if heat is being transported in water that escapes from fissures into the permeable wallrocks to the east and then migrates down the hydraulic slope, eventually escaping into Steamboat Creek. Heat is lost by conduction to the surface as the water migrates eastward.

Auger holes 4 and 5 on the apron of the Main Terrace (pl. 2 and fig. 47), were dug to the water table. Temperatures are near 25° and 35°C, respectively, near the water table, which is here about 7 feet below the ground surface. The water is about 30 percent higher in salinity (table 35) than normal spring water, which indicates extensive concentration of salts by surface and near-surface evaporation. Some heat is probably being transferred in circulating water as well as by rock conduction.

Auger holes 11–13 were dug west of the Main Terrace and just north of the structure section line through GS–3 (pl. 2). A–11 and A–12 about 800 feet west of the Main Terrace fissure system are relatively low in temperature (fig. 47) and are similar to near-surface temperatures in the South Steamboat and Harold Herz wells on the south and north margins of the thermal area.

A–13 is only 400 feet west of the Main Terrace fissure system, and this close proximity probably explains why it is higher in temperature than the more distant A–11 and A–12 holes. Above the water table the thermal gradient in A–13 is about 3°C per foot of depth; the two

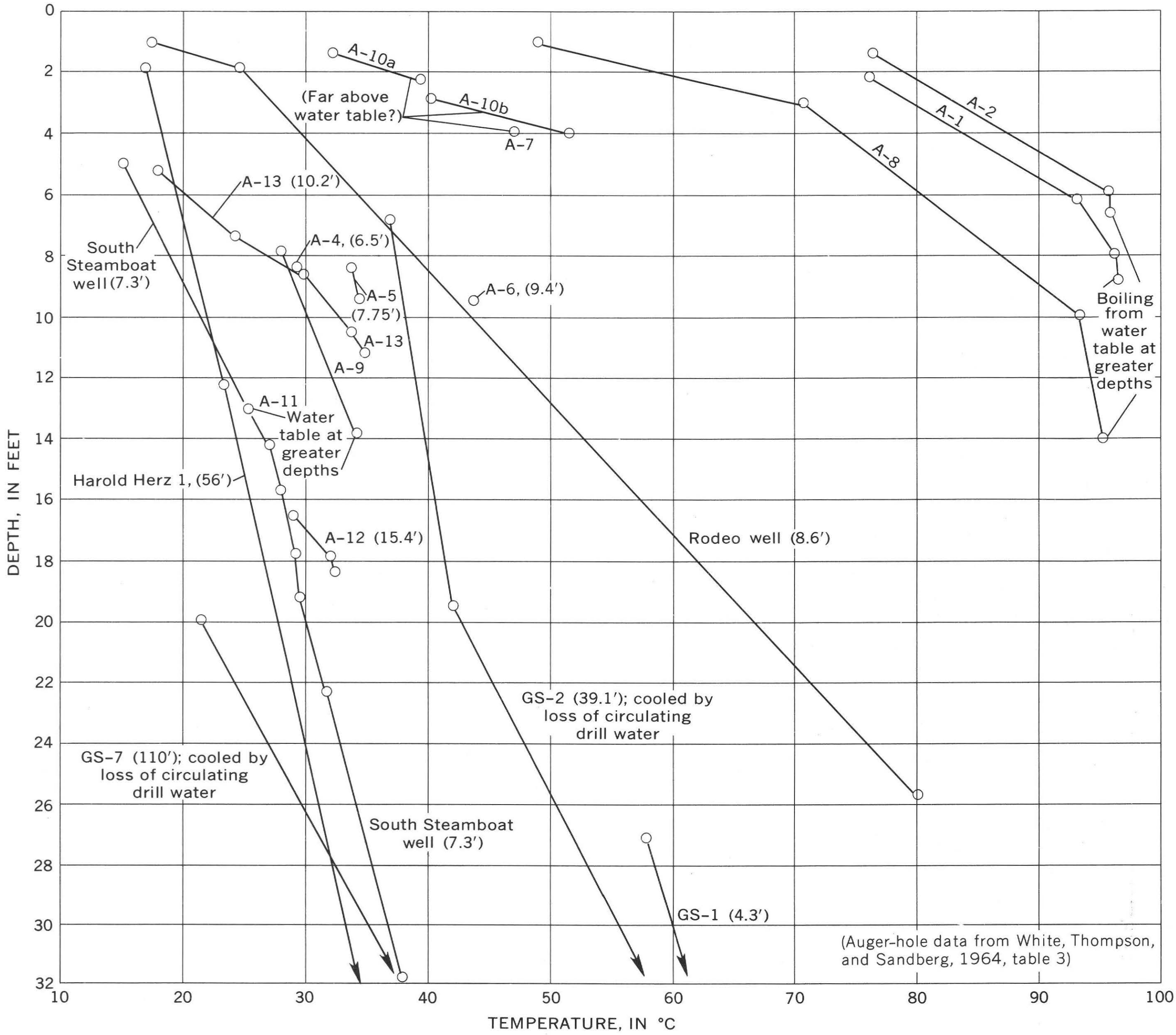

FIGURE 47.—Temperature-depth relationships in auger holes, compared with shallow measurements in some other drill holes. Number in parenthesis following the well name is depth to the water table, in feet.

rather closely spaced measurements below the water table suggest a gradient of only 2°C per foot of depth. This difference, although perhaps not significant, may be caused by greater thermal conductivity of water-saturated sediments relative to similar unsaturated sediments; a water-saturated capillary fringe may extend upward 1 or 2 feet above the water table at 10.2 feet.

The East Reno well (pl. 2; tables 5, 39) had a temperature of about 138°C at a depth of 157 feet, which is about 150 feet below the probable local water table. This temperature is about 8°C below the corresponding temperature in drill hole GS-4 and 14°C below the boiling point of pure water for the depth.

The West Reno well is located 600 feet west of the East Reno well (pl. 1). Its water level when measurable was about 26 feet below the surface (table 5). The altitude of the water table intersected by the two Reno wells is probably almost identical, sloping slightly toward the West well as part of a cone of depression during and immediately after intervals of high production; natural slope unaffected by production is probably toward the East well and Steamboat Creek.

The temperature of the West well was about 138°C at a depth of 186 feet below the surface or 160 feet below the water table; its measured temperatures are slightly lower than those of the East well.

At least four or five wells have been drilled near the Mount Rose Resort on the northern extension of the Main Terrace, about 1,200 feet north of the Reno Resort (pl. 1); three are listed in table 39 (Nos. 57–59). Temperatures as high as 133°C were measured in well 58, only 8°C below the theoretical boiling point for the existing depth and water table. The few measured temperatures in these wells are erratic with depth and time, and relationships are not clear because of the absence of data as drilling progressed. The Mount Rose wells are commonly induced into eruption, at times with great difficulty.

The Murray well (table 39, No. 55) is 1,500 feet east of the Mount Rose Resort and just east of Steamboat Creek. The well reportedly was drilled (Otto Herz, oral commun., 1948) many years ago to a depth of 680 feet. In 1946 the well was not accessible below 63 feet, where the relatively low temperature of 37°C was measured.

Three wells were drilled by Harold Herz (table 39, Nos. 60–62) in the northernmost part of the main thermal area, about 2,000 feet northwest of the Reno Resort, (pl. 1). Temperature and other data of well 1 are shown in table 21 and figure 27. The data are of particular interest because—

1. Bottom-hole temperatures measured as drilling progressed are unusually close to original ground temperatures, because 5 days normally elapsed between weekend periods of renewed drilling; the well was drilled by cable-tool rig, which is the optimum drilling method for reliable bottom-hole temperatures. For these reasons, figure 27 is probably the best representation of original ground temperatures yet obtained in a thermal area.
2. Original ground temperatures were in general about 45°–50°C below the boiling-point curve for the prevailing environment.
3. The temperature gradient in the upper 50 feet of the hole is 1°C per 1.8 feet of depth, and the slope projects to a surface temperature of 16°C (6°C above the mean annual temperature). The shallowest point at 1.8 feet is of little significance, however, because of probable strong influence by daily and seasonal changes.
4. A change in slope occurs at or near 51 feet (5 ft above the best figure for the water table as measured on Oct. 25 in the shallow open hole), and this slope extends down to 122 feet, with an average of 1°C per 2.0 feet of depth. The change at or near 51 feet may be related to slightly higher thermal conductivity of ground saturated with water, including capillary water, relative to shallow water-unsaturated ground. An interpretation of this nature would not be warranted by ordinary data, but it is suggested here as a possible application of very good data obtained because of the unusual circumstances of drilling. The interpretation of effects by a capillary fringe depends in part on an unusually reliable figure of 47.7°C at 56.9 feet and an equilibrium temperature that is unlikely to be less than 44.9°C at 51.1 feet (only 16 hr had elapsed after the previous drilling).
5. The temperature gradient from about 122 feet to the bottom of the hole is 1°C per 2.6 feet of depth, or only two-thirds of the near-surface gradient. The explanation for this lower gradient is not clear, but it could be related to presence of an aquifer with slight horizontal fluid movement near 120 feet in depth.
6. Convection is clearly demonstrated in this uncased 6-inch hole by temperature measurements made 6 and 54 days after drilling ceased (fig. 27). There is no evidence for water flowing into or from the well at different levels. The top of the water column was heated and the bottom cooled by convection.
7. The sediments are all relatively impermeable, and original ground temperatures are probably ex-

plained almost entirely by heat conduction, at least to a depth of 120 feet.

GEOPHYSICAL DATA

Geophysical traverses across the Main Terrace were far less revealing than similar traverses on and south of the Low Terrace (White and others, 1964, p. B58–B61). A gravity traverse through GS–3, GS–4, and GS–8 provided evidence against major Pleistocene faulting along the fissures and breccia zones. Most of the decipherable structural offsets are evidently east of this fissure system, as indicated in the structure sections of plate 2.

CIRCULATION WITHIN THE TERRACE

Drill holes GS–4 and GS–5 provide evidence that the hottest thermal water is rising up the main east-dipping fissure system, shown diagrammatically in the structure sections of plate 2. A little cold meteoric water does flow into the system from local sources west of the terrace, but this seems to be restricted largely to shallow sedimentary layers at and above the unconformable contact with granodiorite. The chloride content of about 780 ppm in deep GS–3 water as compared to 790–825 ppm in samples from GS–4 and GS–5 may point to dilutions of 1–5 percent within accessible depths below the unconformity.

In the absence of more complete data from very deep drilling or from rates of mixing and conductive heat loss, the temperature gradient in the main channels of upflow cannot be known with certainty. All temperature data from drill holes consistently demonstrate, however, that little increase in temperature occurs in the Steamboat Springs system at depths below 350 feet. The water temperature deep in the system is evidently about 175°C or slightly higher, and all gases are contained in solution in a single liquid phase because of the high pressures. The upwelling single-phase liquid loses only a little heat into the walls of the fissures because the horizontal thermal gradients must be relatively low in an old well-heated system and because the thermal conductivities of fresh and hydrothermally altered rocks are so low (table 41) that these rocks serve as good insulators.

When this hot water has risen within about 350 feet of the surface, however, hydrostatic pressure has decreased so much that vapor bubbles first start to form at temperatures that are still about 10°C below the boiling-point curve for pure water; note in particular GS–4 and GS–5 (figs. 34, 35) at depths between 250 and 350 feet. This is evidence that the vapor pressure of dissolved gases, such as CO_2 and H_2S are high, and that the first vapor bubbles must be strongly enriched in these gases.

TABLE 41.—*Thermal conductivities of fresh and hydrothermally altered rocks from the Steamboat Springs area, Nevada*

Sample	Density[1]	Conductivity millical per cm sec deg[1]	Description
A–17	2.69	6.23	Fresh granodiorite, Winters Creek.
W419	2.69	6.66	Fresh granodiorite boulder, crest of Low Terrace.
W420–a	2.70	6.58	Fresh granodiorite, west of Low Terrace.
W421	2.73	6.27	Fresh metatuff, southwest end of Steamboat Valley.
W128–0	2.62	3.59	Fresh Steamboat basalt, northeast of silica pit.
W422–a	2.58	3.66	Fresh soda trachyte, Steamboat Hills.
W355	2.54	4.55	Slightly altered soda trachyte, east of Steamboat Resort.
W226	1.44	1.4	Porous opaline sinter, Main Terrace.
W366–195	2.62	6.5	Adularized soda trachyte, Senges well.
W387–656	2.70	5.45	Slightly altered hypersthene-augite andesite, Young drill hole, Virginia Range.
586	2.66	5.50	Moderately propylitized hypersthene-augite andesite, Young hole.
225	2.32	5.10	Hornblende-pyroxene andesite altered to zeolites, argillic minerals, Young hole.
GS–2–82	2.24	6.0	Opaline sinter, partly reconstituted to chalcedony.
GS–3–29	2.02	3.5	Opal-cemented arkosic sandstone, partly opalized.
51	2.30	4.2	Arkosic sandstone, partly argillized, cemented.
74	2.62	6.9	Granodiorite, slightly altered, some fresh biotite.
139	2.62	7.0	Granodiorite, moderately altered, little fresh biotite.
169	2.67	7.0	Granodiorite, slightly altered, much fresh biotite.
205	2.61	6.9	Granodiorite, moderately altered, little fresh biotite.
243	2.53	7.9	Calcite-chalcedony vein.
256	2.61	7.4	Granodiorite, moderately altered, biotite to chlorite, chalcedony veinlets.
360	2.58	6.5	Granodiorite, moderately altered, biotite to chlorite.
460	2.61	6.7	Granodiorite, similar to GS–3–360.
513	2.52	6.7	Granodiorite, strongly altered, biotite and plagioclase argillized.
632	2.66	6.7	Granodiorite, slightly altered, some fresh biotite.
685	2.66	6.8	Granodiorite, slightly altered, much fresh biotite.
GS–4–16	1.77	2.8	Opaline sinter, in part geyserite.
62	1.94	4.4	Arkosic sediments, partly altered to opal, kaolinite.
77	2.08	3.8	Arkosic sediments, fine grained, clay minerals common.
144	2.58	8.3	Arkosic sediments, cemented by adularia, chalcedony.
162	2.62	7.0	Granodiorite, moderately altered, all biotite gone.
218	2.64	10.1	Granodiorite, sheared, strongly altered, replaced by quartz, hydromica, stongly altered.

See footnote at end of table

TABLE 41.—*Thermal conductivities of fresh and hydrothermally altered rocks from the Steamboat Springs area, Nevada*—Con.

Sample	Density[1]	Conductivity millical per cm sec deg [1]	Description
GS-4-261	2.09	4.9	Granodiorite, sheared, strongly altered, replaced by hydromica, calcite.
320	2.42	6.3	Granodiorite, strongly altered, plagioclase argillized.
356	2.29	5.7	Do.
404	2.27	5.9	Do.
462	2.53	7.8	Granodiorite, sheared, replaced by quartz, sericite.
496	2.36	6.5	Granodiorite, strongly altered, plagioclase argillized.
GS-5-535	2.44	6.0	Soda trachyte, highly altered, hydromica.
GS-6-16	2.51	8.1	Chalcedonic sinter, reconstituted from opaline sinter.
160	2.61	5.55	Soda trachyte, altered largely to adularia.
GS-7-20	1.58	2.6	Granodiorite, acid-leached to opal, relict quartz.
68	1.32	1.6	Granodiorite, acid-leached to opal, relict quartz; pyrite.
111	1.37	2.3	Similar to GS-7-68 but more pyrite.
127	2.42	7.9	Granodiorite, acid-altered to alunite, kaolinite.
217	2.43	6.3	Granodiorite, silicates largely altered to montmorillonite.

[1] Determined by Prof. Francis Birch, Dunbar Laboratory, Harvard University. Temp 30°C, pressure 2,000–6,000 psi except for sample W226, where pressure was 800 psi. Effect of pressure in this range stated to be "only a few percent." Samples were dry, and pore spaces presumably occupied by air.

As the water rises further and pressure decreases, more vapor with increasing steam content forms and substances of low volatility, such as NaCl, are concentrated in the remaining liquid water. Curve *B* of figure 38 indicates that if no heat is lost by conduction, 14.3 percent of water at 172°C is converted to steam when pressure is lowered to the average atmospheric pressure at Steamboat Springs. The best value for chloride content of the deepest and hottest waters of the Main Terrace is 810–820 ppm (table 35, GS-4 and GS-5). For comparison, the average chloride content of water discharged from the hottest and most centrally located springs, such as springs 18, 23, 23n, 24, 27, and 28, is close to 925 ppm. This difference could be explained by residual concentration from the vaporization of about 13 percent of the deepest and hottest water; this figure is in very close agreement with the calculated value of 14.3 percent from figure 30. Other factors involved in controlling the actual chloride contents of individual waters are conductive loss of heat, dilution by cooler low-chloride water and near-surface cooling, and evaporation in open fissures at temperatures below boiling. These later factors tend to offset each other in the average spring water, thus explaining in part the close agreement between the actual and calculated residual concentrations cited above; but these other factors help to explain existing differences in chloride content of individual springs, which has ranged from as low as 600 ppm in spring 33 to as much as 980 ppm in springs 18 and 34.

Springs on the borders of the terrace, such as springs 2, 3, 19, 19n, 20, 20n, and 21, are generally lower in temperature and in chloride content; typical chloride concentrations of these springs are 860–900 ppm. Perhaps 5–10 percent of the deep-water supply of these springs was converted to steam; the lower temperatures of these springs also indicate cooling by evaporation at the surface, or by conducted heat flow.

Table 35 and other tables provide evidence that the chloride content of some water at relatively deep levels is higher than can be explained by simple upflow and boiling; for example, 930 ppm in Rodeo well at 146 feet as much as 828 ppm at depth in GS-5 (table 31); and at least 896 ppm at 130 feet in GS-8 (table 34). The best explanation for these anomalously high chloride contents is that convection is occurring within the upper part of the system, where permeabilities of interconnected channels are presumably highest. Evaporative concentration occurs in the upflowing parts, but some of this water circulates downward again in a complex network of interconnecting fissures. Independent evidence for shallow convection within the Main Terrace is presented elsewhere (White, 1967a) from SO_4:Cl ratios in the flowing springs and drill holes. H_2S is concentrated in the vapor phase as boiling occurs; the vapor rises above the water table and SO_4 is produced by near-surface oxidation of H_2S; SO_4 in sulfates and H_2SO_4 is carried downward in precipitation that falls on the area or in water condensing from vapor in the rising gases; SO_4 is then incorporated in the water body at the water table. The higher SO_4:Cl ratios of certain spring and well waters are best explained by circulation and convection of this SO_4-enriched water.

HIGH TERRACE

Natural springs have not discharged from the High Terrace since the late Pleistocene. Very slight evidence for activity is observable at the ground surface, especially in the winter when snow melts faster near the crest of the terrace and condensation of water vapor is visible in warm feeble gas seeps when atmospheric conditions are particularly favorable.

Temperature relationships and the circulation pattern within the High Terrace must be interpreted almost entirely from drill hole GS-2 (table 28; fig. 40; structure section, on pl. 2). The detailed log of the hole was published by White, Thompson, and Sandberg (1964, table 3).

The water table was penetrated at about 40 feet in depth. The temperature measured nearest the water table (71°C) was probably much below the original ground temperature, because of cooling by an uncommonly large amount of introduced drill water; the water loss was high because the local rocks were brittle and brecciated, and because pressure of drill water was 12–18 psi greater near the water table of GS–2 (~40 ft below surface) than in drill holes previously discussed.

Below the water table, temperatures were about 10°C below the theoretical boiling curve down to a depth of 300 feet; the maximum temperature of 161°C was measured at 303 feet. Temperatures then decreased slighthly at greater depths, probably because the hole penetrated the footwall of the principal channels of upflow of hot water, as indicated in a structure section of plate 2.

Because much drill water was lost in drill hole GS–2, water samples showed no significant increase in chloride over that of the drill water until 5 days after the hole was completed (table 28). Several zones of differing chloride content could have been cut by the hole without detecting differences. This possibility is suggested by the fact that the erupted sample on September 20, 1950, had nearly 50 percent more Cl than the nonerupted sample of September 21, although eruption of water from 160° to atmospheric pressure can account for only 11.8 percent evaporative concentration (fig. 30, curve *B*). Either a low-chloride aquifer occurs below the cased interval or anomalous low-chloride water forms in the upper part of the hole, possibly by rising steam that condenses and concentrates near the water level. A low-chloride aquifer, perhaps from 150–190 feet below the surface, is the preferred explanation.

GS–2 was cemented to 129 feet (table 28) and cased but not cemented to 185 feet. The drilling of the hole must have changed previously existing circulation patterns in some unknown way, because in the winter of 1950–51 steam and other gases escaped from nearby natural fissures at rates far greater than predrilling rates (fig. 48). The well was never erupted for more than a few minutes at a time, but a blocking of the casing at 86 feet in depth had occurred by May 1952, probably from deposition of $CaCO_3$ in circulating water.

The Nevada Ore Minerals well shown on plate 1 was drilled on the east slope of the High Terrace some time in 1955, but no satisfactory data were obtained. The well reportedly erupted continuously during its early life, but from about 1956 to 1965, it erupts periodically as a geyser.

Senges well (table 5; pl. 1) is near the east base of the High Terrace. A temperature of 145°C was meas-

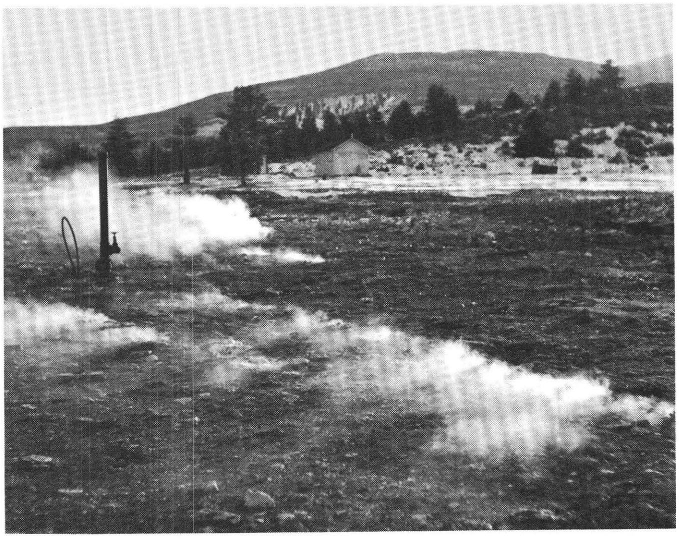

FIGURE 48.—Condensing steam and other gases escaping vigorously near drill hole GS–2, High Terrace, from formerly inactive natural fissures. Water table about 40 feet below surface. Photograph by Dr. P. K. Ghosh, Geological Survey of India, 1951.

ured at a depth of 177 feet, which is about 147 feet below water level; the theoretical boiling point for this depth is 152°C. The well normally boiled vigorously when the valve was open but did not erupt spontaneously. When an eruption was induced, a normal water-steam mixture was first ejected, followed within a minute or two by a steam phase with sparse water droplets. Temperatures clearly were high enough to sustain continuous eruption, but permeabilities of rocks in this well are evidently too low to sustain an adequate flow of water.

The water table in the High Terrace was penetrated at an altitude of about 4,680 feet in GS–2; northward, the water table slopes to about 4,545 feet in the Harold Herz wells (table 5) and eastward to about 4,640 feet in Senges well and 4,580 feet in the Reno wells. No water-table data are available for the area just west of the High Terrace, but a mounding under the terrace with lower levels to the west as well as to east and north is probable. The hypothesized mounding should be caused by upflow of hot water in the fissure system, as in figure 6. The close approach of temperatures in GS–2 to the boiling-point curve is evidence that upwelling does occur in the High Terrace. The temperatures cannot possibly be explained by heat conduction alone or even by conduction as a major factor. Some water undoubtedly escapes into the most permeable wallrocks and flows to the east and north below the surface; some subsurface flow to the west may also occur.

SINTER HILL

The temperature relationships and circulation patterns of Sinter Hill, an old faulted and extensively eroded hot-spring terrace, are imperfectly known (White and others, 1964, p. 51 and table 3). No springs have discharged from the hill since middle or late Pleistocene time. Drill hole GS–6 is located in the northeastern part of the sinter deposits (pl. 1; fig. 36; table 32).

Auger holes 1 and 2 are near the west end of the hill, and auger hole 7 is on the north flank.

Despite the fact that hot water has not discharged at the surface for many thousands of years, temperatures are rather high in GS–6. The measurement of May 1952 (table 32; fig. 36) was only 18°C below the theoretical boiling-point curve for the environment. Heat is flowing from the water table largely by conduction, with relatively little transport by rising gases.

The very low chloride content of 12 ppm in GS–6, even 1½ years after completion (table 32), is evidence that the deep chloride water characteristic of other terraces does not flow up to the explored depths. The sampled water must be precipitation that falls on the area and percolates downward, heated because of the high geothermal gradient.

Auger holes A–1 and A–2 are in a small basin just west of the crest of Sinter Hill. This basin is surrounded and underlain by acid-altered rocks and has formed in part by subsidence caused by removal of matter by acid leaching. An old steaming well a few feet deep is on the southeast rim of the basin.

The auger holes A–1 and A–2 shown in figure 47 indicate a very rapid rise in temperature to depths of 6 and 8 feet, where nearly saturated steam occurred. The water table was not penetrated and is likely to be at least 100 feet below the surface, judging from relationships in the nearby drill hole GS–6 and the Mercury well (table 5).

Rather vigorous subsurface convection is indicated in the western part of the Sinter Hill area. Hot water probably rises along the northwest fault that is one of the structural controls of this basin (pl. 1). Temperature at the water table must be at the boiling point, and considerable steam and other gases are evolved. The upward flow of steam is so high that little is condensed, except very near the surface where the thermal gradient is very steep and much heat can be removed by conduction. In A–2, for example, the temperature rose about 65°C in the uppermost 1.4 feet of depth.

The shapes of the temperature curves of A–1 and A–2 (fig. 47) are characteristic of an environment above the water table where considerable steam is being evolved, but not at rates to permit escape at the surface (see Banwell, 1957a, p. 31). The hypothesized relationships are shown diagrammatically in the dashed-line curve for GS–7 (fig. 37). Similar relationships probably also existed above the water table near drill hole GS–2 on the High Terrace before drilling.

A–7 was driven in a small collapse depression on the north flank of Sinter Hill. A temperature of 48°C was found at a depth of only 4 feet (fig. 47). Hard sinter prevented further deepening of the hole.

SILICA PIT

Springs have not discharged from the silica pit area (pl. 1), at least not since early Pleistocene time. The area is characterized by intense alteration and leaching by sulfuric acid, the chemistry of which is summarized by White, Thompson, and Sandberg (1964, p. B45, B46). There is little surface evidence for abnormal heat flow in the area at present. Warm gases rise at one place near the base of the southeast wall of the main pit, where temperatures as high as 65°C were measured about 1 foot below the surface.

GS–7 was drilled 300 feet southwest of the main quarry (pl. 1; fig. 37; table 33). The water table was penetrated at about 110 feet in depth. Original ground temperatures at greater depths were probably very close to the theoretical boiling-point curve to about 250 feet. A maximum temperature of 161°C was found at the bottom of the hole after a steady state was attained. In contrast to all other holes drilled below the water table in the Steamboat area, GS–7 after completion did not fill with water when its valve was closed. Steam without water evidently rises from the bottom of the hole and is diverted into wallrocks at or below the bottom of the casing (325 ft). Some water evidently condenses from steam in the cooler upper part of the casing and drips downward; it could be collected slowly in a sampler lowered near the bottom under pressure, but no standing water level was ever identified after the hole was completed. All samples of water collected from GS–7 were very different in composition from the chloride waters of the flowing springs. An analysis of water from this hole is included in the sodium bicarbonate type by White, Hem, and Waring (1963, p. F47). The origin of such water was discussed by White (1957a).

The measured temperatures in the upper part of GS–7 (fig. 37, curve *A*) suggest that original ground temperatures before drilling may have been close to a straight-line gradient from the surface to the water table, presumably dominated by rock condition. The thermal conductivities of three specimens of near-surface porous acid-leached rocks are all nearly identical and are only one-third as high as the conductivities of the altered rocks below the water table (table 41). A geothermal gradient controlled entirely by conductivity of the rocks could then be relatively high. The near-boiling temperatures measured at the water table, however, provide evidence for an active convection system; because the rocks above the water table are so porous, the vapor pressure of hot water at the water table must be nearly that of the normal air pressure of the area. Some steam and other gases must then separate at the water table and rise into higher ground, thereby transferring some heat. The shape of the temperature curve above the water table before drilling was probably much more similar to the curves of A–1, A–2, and A–8 shown in figure 47 than to a straight-line gradient. The actual temperatures measured in the upper part of the hole (fig. 37, curve *A*) that suggested a straight-line gradient controlled by rock conduction are probably much too low because of excessive loss of cold drill water in these porous rocks. The estimated dashed-line curve *B* of figure 37 is qualitatively much more probable.

MUD VOLCANO BASIN

Mud Volcano Basin, in the northwestern part of the thermal area (pl. 1), was probably the site of a major mud-volcano eruption in middle or late Pleistocene time (White, 1955b). The rim of a small recent mud volcano is distinguishable near the south end of the basin (White and others, 1964, p. B43–B44). Little other surface evidence for present thermal activity exists, but temperature relationships in auger hole 8 near the center of this small mud volcano prove continued upflow of steam, gases, and heat (fig. 47). Depth to the water table is not known but is probably at least 50 feet below the surface. Judging from the temperature curves in figure 47, the rate of upflow of gases is slightly less than in auger holes A–1 and A–2, but an active convection system is indicated in Mud Volcano Basin.

CLAY QUARRY-PINE BASIN AREA

The clay quarry which is about 1,000 feet northwest of Nevada Thermal Power Co. well 4 (pl. 1; White and others, 1964, p. 52 and fig. 18) is an area of extensive alteration and leaching by sulfuric acid. There is little evidence for present activity, and most of the ground is cold. Condensation of water vapor in the southeast corner of the quarry was seen occasionally on cold days. An abnormal rise in temperature with depth is also evident in auger holes A–10a and A–10b (fig. 47). Depth to the water table is not known but is probably at least 50 feet below the surface, judging from the fact that an old well 1,000 feet farther to the west is 58 feet deep but does not reach water (table 39, No. 80). The bottom-hole temperature of this well was 70°C.

Four geothermal exploration wells were drilled from 1959 to 1961 in the hydrothermally altered area extending east from the clay quarry to Pine Basin. Temperatures were measured in three of these four wells as drilling progressed and are shown in tables 22–24 and in figure 28. These three wells show marked similarities. Temperature in each increases rapidly to depths of 100–140 feet and then almost on a straight-line gradient to depths that range from 470 feet in Nevada Thermal Power Co. well 6 to 550 feet or a little more in their well 4. Temperatures at greater depths increase only slightly, and in well 5, clearly decreases below 700 feet.

Data defining the water table are scanty, as is usual in geothermal drilling unless special attention is given to this important parameter. They are best for Nevada Thermal Power Co. well 6 (table 24). The driller's comment indicates clearly that the water table was intersected at a depth no greater than 95 feet. The shape of the temperature curve in figure 28 suggests that the water table in well 6 was close to 86 feet. The driller's observations when the well was 300–400 feet deep suggest a water table some 30 feet deeper. Confirmed changes of as much as 12 feet with deeper drilling were found in drill hole GS–2 (table 28), and lesser changes were found elsewhere. Unless permeable channels are intersected, the water level cannot be closely defined in tight rocks if only short time intervals separate disturbances caused by drilling. Cable-tool drilling that removes cuttings and water by bailing depresses the water level, and indicated depths are likely to be greater than equilibrium levels unless bailing is more than offset by addition of water.

The water table was probably intersected in Nevada Thermal Power Co. well 4 at about 95 feet below the surface. The depth of 217 feet suggested by the driller (table 22) had a measured temperature close to 112°C when first penetrated by the drill; this is obviously much too high for a water table in direct contact with the atmosphere through fractures and porous leached rock. The water table in Nevada Thermal Power Co. well 5, just 900 feet northwest of well 4, is probably not more than 130 feet below the surface and is likely to be about 120 feet, judging from the temperature curve.

The temperature record of Nevada Thermal Power Co. well 4 suggests zones of circulating thermal water at depths of about 90 to 140 feet, 540 to perhaps 570 feet, and near the bottom of the hole. A cooler aquifer is suggested near 575 to 600 feet. The low chloride content (45 ppm) of the water sample collected when the depth was 520 feet suggests, however, that this water was largely from shallow meteoric circulation, heated in the high geothermal gradient. Water erupted from the well after completion had a nearly normal chloride content of 874 ppm.

Nevada Thermal Power Co. well 5 may intersect zones of actively migrating thermal water near 130, 340, 530, and 700 feet in depth. Migrating cooler water may exist near 325 feet and near or below the bottom of the hole.

Nevada Thermal Power Co. well 6 may have thermal zones from 90–140 feet and at 310, 470, and 625 feet. Cooler water is suggested near 575 feet and perhaps locally elsewhere.

The maximum temperature of the deep wells in this area are close to or slightly above 180°C, which is a little higher than the maximum of 172°C measured in the most active spring terraces. The explanation for the higher temperatures in the western area, where there is no direct discharge, is not clear.

HEAT FLOW

STEAMBOAT SPRINGS SYSTEM

Much interest has been shown in recent years in the total natural heat flow of individual hot-spring areas (see summary by White, 1967a). The total heat flow is a very useful parameter in exploration of an area for geothermal energy, providing a first approximation of the minimum rate at which heat can be withdrawn for power development of water or steam (Benseman, 1959a; Bodvarsson, 1964). The total heat flow before drilling is also the most significant reference base for evaluating effects of accelerated withdrawal from a producing geothermal field (White, 1964, 1967a).

In low-temperature spring systems from which all circulating water is discharged at the surface and none escapes unseen below the surface, the approximate total heat flow is determined easily from rate of discharge of water and discharge temperature. Fukutomi (1962) found that conducted heat flow from such low-temperature areas accounts for only about 10 percent of the total and that discharging water accounts for about 90 percent. The actual proportion will vary with the magnitude of discharge of water and the size of the area involved in discharge and conduction. Where discharge is very low and the area involved in conduction is high, the rising water loses most of its heat by conduction. In extreme cases, where discharge is less than 1 gpm, a mineral water of deep origin can be near the mean annual temperature, with nearly 100 percent of its original heat content lost by conduction.

The total heat flow from a spring system is much more difficult to measure where water escapes unseen below the surface (Benseman, 1959b), or where a large part of the total escapes in steam or by rock conduction under the influence of extreme near-surface geothermal gradients (Benseman, 1959a; Dawson, 1964).

White (1967a) has summarized almost all estimates and measurements of total heat flow published through 1962 for hot-spring areas of the world. In near-boiling systems that lost much heat in steam, temperatures are likely to rise rapidly just below the surface, closely controlled by the boiling-point curve. At some depth that depends on the temperature of the upwelling single-phase liquid without vapor, the temperatures tend to level off, showing little further increase within explored depths. The most reasonable explanation for this commonly observed phenomenon is that water is circulating in a huge convection system below explored depths and attains a base or levelling-off temperature that is characteristic for each system, depending on the rate of flow of water and heat in the system, similar to the idealized models of Donaldson (1962, figs. 2, 6).

After heating to the base temperature, as suggested in figure 3 the water rises by convection within the core of each system because of density differences related to thermal expansion. Little change in temperature occurs in the rising insulated mass until decreasing pressure near the surface permits a vapor phase to form.

This picture is no doubt much oversimplified for systems with two or more convection cells or systems with convection rates so low that temperatures differ little between downflowing and upwelling parts, as in the curves of least contrast considered by Bredehoeft and Papadopulos (1965, fig. 2). Nevertheless, the method does permit an easy means for estimating the heat flow of the system, provided that (1) the rising water is relatively high in a very soluble constituent such as Cl or B that is present only in low concentration in the normal surface and ground waters of the area; (2) all the tracer constituent in the thermal water is discharged by one means or another into a river or stream that can be monitored, as Steamboat Creek was monitored, to provide the data for tables 36, 38, and 40; and (3) subsurface drilling has provided a reasonable base or "leveling off" temperature.

We have seen that the best figure for the Cl content of deep upwelling water of the Steamboat Springs system is near 820 ppm, and this figure, together with discharge and chlorinity data from Steamboat Creek at Huffaker Hills, is used to calculate a total discharge of 1,130 gpm from the system.

Using this discharge figure, a base temperature of 175° C, and a mean annual temperature of 10° C, the total heat flow of the Steamboat Springs thermal system is calculated as 11.8×10^{6} cal per sec, using the formula—total heat flow $=\left(1130\ \frac{\text{gal}}{\text{min}}\right)\left(63.1\ \frac{\text{cm}^3\text{ per sec}}{\text{gpm}}\right)$ $\left(1\ \frac{\text{g}}{\text{cm}^3}\right)\left(1\ \frac{\text{cal}}{\text{g}^\circ\text{C}}\right)(175^\circ-10^\circ\text{C})$.

A more conservative estimate of 7×10^6 cal per sec for the heat flow, based on estimated total discharge of only 700 gpm from the system, was published by White (1957a). Other measurements have indicated at least as much as 1,300 gpm, equivalent to 13.5×10^6 cal per sec of heat flow.

The mean heat flow of the earth (Lee and Uyeda, 1965) is about 1.5μ cal cm^2 sec, or 1.5×10^4 cal per km^2 sec. The Steamboat Springs system, with an area of 5 km^2 pl. 1, has a total upflow of heat equivalent to 780 km^2, or an anomaly at least 150 times 'normal.' Does some mechanism exist for deep circulation of water that collects the heat of a large surrounding area and funnels it through a small hot-spring area? The following sections examine this possibility, first assuming "normal" heat flow for the region and then some higher regional heat flow that is significantly above the world average.

TRUCKEE MEADOWS AND OTHER NEARBY BASINS

If the heat flow of the region were near the crustal average, with circulating water of the hot-spring system absorbing most of this heat and funneling it into the hot-spring system, the surrounding region should actually be subnormal. This would be necessary to offset the strikingly abnormal flow of heat that is discharged in hot-spring water and steam.

No heat-flow data are available from the surrounding region. The only data we have to assess the possibilities consist of temperatures measured in wells, nearly all of which were drilled for domestic or industrial water supply. Moreover, nearly all these wells were drilled in the basin areas, constituting roughly 120 square miles of the 350 square miles of total area that drains into Steamboat Creek and Truckee Meadows and could be involved in the hypothesized convection system.

Table 39 includes 155 wells, most of which have some temperature data. For many of these wells, only temperatures of flowing tap water or artesian flow are available, providing probably minimum bottom-hole temperatures. Some additional data from the Truckee Meadows basin are available from Cohen and Loeltz (1964, table 5).

Figure 49 shows the best available data from wells of the region, excluding those within the immediate thermal area and within a broad belt that extends northward from the springs to Huffaker Hills. These excluded wells are likely to be influenced thermally by subsurface flow from the spring system (Cohen and Loeltz, 1964, pls. 1–3).

In figure 49, reported temperature is plotted against the measured or reported depth of each well. Most of the wells are in basin sediments unsaturated with water to depths that are generally only a few feet but are rarely as much as 100 feet or more. In the absence of measured thermal conductivities for the rocks and sediments actually involved, a reference line of a "normal" geothermal gradient is shown, calculated by assuming a "normal" heat flow of 1.5μ cal per cm^2 sec and a conductivity of 3×10^{-3} cal per sec cm °C. We see that only 13 points fall on or below this line and more than half of the points indicate geothermal gradients of at least two times the reference "normal" gradient.

A closer examination of some individual points shown in figure 49 is of interest. Points 102 and 103 are in Truckee River sediments in and west of Reno; their low temperatures are probably related to relatively rapid inflow of water from the Truckee River. Point 2 is from a shallow well in the upper part of the Galena Creek fan near the base of the Carson Range. Since Galena Creek is fed by melt water and the average temperature of inflow into the fan sediments is no doubt significantly less than 10°C, the local geothermal gradient in this well may not be anomalously low.

Points 17, 32, 40, and 74 are all near the lower part of Whites Creek, and point 19 is near the lower part of Thomas Creek. Their low temperatures may be due to relatively rapid recharge from these creeks, which drain the Carson Range just north of Galena Creek. The wells in the lower Whites Creek drainage are in or near a conspicuous structural graben cutting pre-Lake Lahontan sediments 2 miles northwest of the thermal area. These graben faults may be significant structural channels for recharge for the Steamboat Springs system, as suggested in figure 4. The relatively low temperatures in wells in and near the graben are consistent with this hypothesis. Cohen and Loeltz (1964, p. S21) note that total seepage losses from Whites Creek in the 3 miles of streambed just west of the graben are about 300 gpm, but streamflow was not measured within or east of the graben.

Unfortunately, a good net of deep-well temperatures is not available for most of the Galena Creek fan west of Steamboat Hills. If a major part of the recharge of the system occurs west of the hills along the east front of the Carson Range, as favored by the stable isotope data summarized in an introductory section and as suggested in the model shown in figure 4, subsurface temperatures in this western area should be anomalously low. The study by Cohen and Loeltz (1964) provides no additional data bearing on this problem.

The isotope data and the sulfate contents of the thermal waters are easier to explain if some recharge occurs

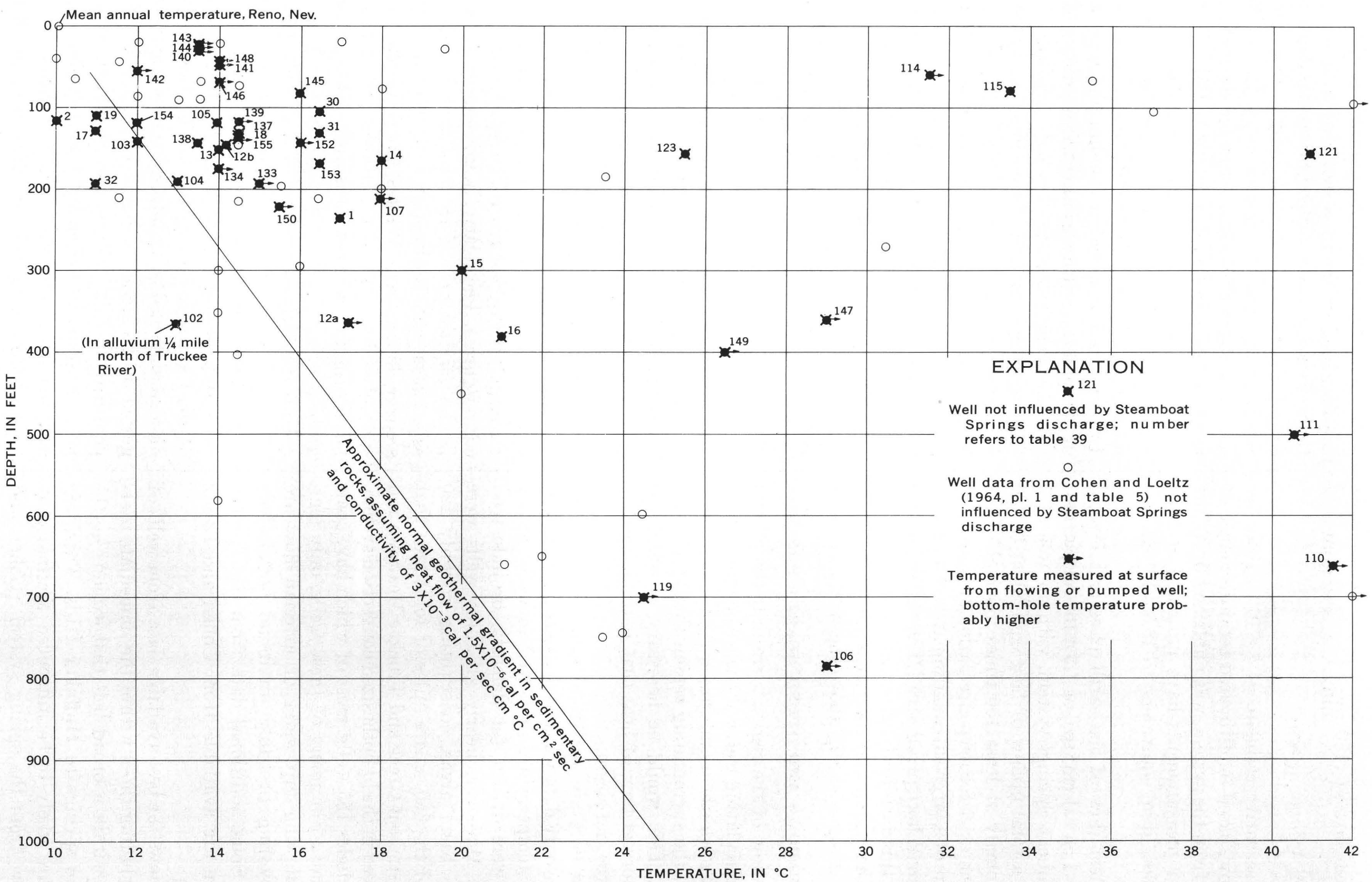

FIGURE 49.—Temperatures and depths of wells in Truckee Meadows and surrounding region, plotted with depth of well.

from high-sulfate streams draining the Virginia Range east of the springs (see Cohen and Loeltz, 1964, pl. 3). This possibility is not supported by the limited temperature data available. Points 3, 23, 24, 52, 53, and 99–101 are all near the western base of the Virginia Range east of Steamboat Springs. If a high rate of recharge exists in this area, subsurface temperatures should be low. The average geothermal gradient in these 8 wells is at least 10°C per 100 feet, or 10 times "normal." Their relatively high temperatures may be influenced by close proximity to the hot upwelling part of the Steamboat system, more than offsets the cooling effects of downflow.

CARSON AND VIRGINIA RANGES

The parts of the Carson and Virginia Ranges that drain into Steamboat Creek and that might be involved in the recharge of the Steamboat system constitute about 230 square miles of the total of 350 square miles that could be involved. Unfortunately, almost no subsurface temperature data are available from this large area, because of its small population and its lack of need for water wells.

The relatively high subsurface temperatures in the Comstock Lode just east of the crest of the Virginia Range only 7 miles southeast of the Steamboat Springs are well known (Thompson, 1956, p. 72; Becker, 1882, p. 386). The upflow of hot water in the mines has been regarded by some as an end stage of the ore-forming processes, and by others, including the present author, as more probably a subsurface hydrothermal system related to high regional heat flow but not closely related to ore deposition. With almost no data from elsewhere in the ranges, we cannot distinguish between these two possibilities.

A single drill hole with two observation points shown as 12a and 12b in table 39 and figure 49 provides our only other clues. This hole was drilled 4 miles north of Virginia City in an area of acid-bleached andesitic rocks, propylitized and unoxidized at depths below about 70 feet (for summary of log and alteration, see Thompson and White, 1964, p. A28). The hole was drilled 656 feet deep by a small rotary rig in an unsuccessful search for gold and silver ore. Viscous mud was left in the hole at the time of completion. The day after circulation of mud ceased, the thermometer could be lowered only to 364 feet, where a temperature of 17.2°C (point 12a) was measured; the original ground temperature unaffected by circulating mud must have been considerably higher. More than 6 months after completion of the hole, the accessible depth was only 144.7 feet where the measured temperature was 14.2°C (point 12b). This is probably much closer to an equilibrium temperature than point 12a, but many complicating factors prevent reliable analysis:

1. The mean annual temperature at 6,300 feet altitude is considerably less than the usual reference of 10°C and is probably close to 5°C;
2. Oxidation of sulfides in the upper 50 feet may counterbalance the above, at least in part;
3. The water table is from 42–45 feet deep, and ground water is probably migrating westward to the steep front of the Virginia range;
4. Conductivities of propylitized unoxidized andesite in the hole are all close to 5.5×10^{-3} 1 per cm sec °C (samples numbered 387, table 41), which is close to average igneous rocks; the oxidized rock above the water table was not cored, and its thermal conductivity is not known;
5. In spite of all these uncertainties, a geothermal gradient of two or three times "normal" seem likely for this hole.

Our knowledge of heat flow from the mountainous 75 percent of the area is therefore highly unsatisfactory. There is some indication that heat flow is at least twice "normal," but for lack of better data, it will be considered only 1.5 times "normal" in the following section.

TOTAL HEAT FLOW AND IMPLICATIONS

We have seen that the upwelling water of the Steamboat system transports about 12×10^6 cal per sec of heat. A part of this total is discharged from the immediate thermal area ($\sim$5 km^2) in steam and hot water and by conduction to the surface through the shallow parts of the system that have very high thermal gradients. About a third of the total heat included in the above estimate, however, is contained in very hot water that flows northward below the surface. This water is cooled extensively by mixing with cooler meteoric water and by thermal conduction to the surface. The area of direct influence by Steamboat upflow and northward outflow of heat is considered to be about 3 miles wide and 6½ miles long; it includes Steamboat Springs near its southern end and extends northward to Huffaker Hills. This area of possible direct influence contains about 20 square miles, or 50 km^2.

Other basin areas probably not influenced directly by upflow of heat in fluids of the Steamboat system include other parts of the Truckee Meadows basin (70 sq mi, or $\sim$175 km^2) and Washoe basin (30 sq mi, or $\sim$75 km^2). The heat flow from these other basin areas is not known, but temperature data from water wells suggest an average of two or three times the "normal" crustal heat flow. We shall assume the more consecutive figure of two times "normal," or 3μ cal per cm^2 sec for the 250 km^2 in these other basin areas, or about 8×10^6 cal per sec.

The 230 square miles ($\sim$600 km^2) of mountainous area that drains into Steamboat Creek and Truckee Meadows and that might be involved on a huge scale in recharging the Steamboat Springs system is even less well known. The scanty data yield no evidence that heat flow from this area is subnormal or even as low as "normal." If we assume 1.5 times "normal" or 2.2μ cal per cm^2 sec for 600 km^2 of area, its heat flow is 13×10^6 cal per sec. This estimate admittedly may be very inaccurate, but it is as likely to be too low as too high.

The flow of heat from the total drainage area of 350 square miles ($\sim$900 km^2) thus is probably at least 20×10^6 cal per sec, and a more likely figure is 35×10^6 cal per sec; these estimates are equivalent to the heat flows from 1,300–2,300 km^2 of "normal" area. If the lower value is assumed, the average heat flow from this large area could be as little as 1.5 times "normal" if no heat flows directly to the surface within the mountainous areas; an average of at least 2.5 times "normal" seems much more likely.

Up to this point a crustal average heat flow of 1.5μ cal per cm^2 sec has been used as the standard for comparison. Scanty data from the Basin and Range province (Lee and Uyeda, 1965) suggest that the whole province may be characterized by abnormally high heat flow, perhaps as much as 1.5 times the crustal average. If this is so, the Steamboat drainage area could have an average heat flow that is no higher than the provincial average, if the whole drainage area is actually involved in recharging the hydrothermal system, and if rate of downflow of water throughout most of this area is so high that geothermal gradients are effectively zero.

The total calculated discharge of 1,130 gpm of deep water from the system is equivalent to 4,300 liters per minute, or about 7×10^4 cm^3 per sec. If recharge of this water were distributed uniformly over the total drainage area of 900 km^2 (9×10^{12}cm^2), 1 cm^3 of recharging liquid water must pass through each square centimeter of surface area in 1.3×10^8 sec, or about 4 years. At such a low rate of inflow (a little less than 1 percent of the average precipitation), the geothermal gradient would be decreased slightly but not nearly to the zero gradient demanded by the model.

A much more realistic model assumes that recharge occurs only in certain parts of Truckee Meadows basin and that Washoe basin and the mountainous parts of the whole drainage area are not involved. Furthermore, we shall assume from previous calculations that total heat flow from this basin area consists of 12×10^6 cal per sec from upflow in the Steamboat system, plus about 8×10^6 cal per sec from the parts of Truckee Meadows basin not directly affected by the Steamboat upflow. With these assumptions, 20×10^6 cal per sec of heat is flowing from about 90 square miles (235 km^2) of area. Since "normal" heat flow from an area of this size is about 3.5×10^6 cal per sec, the indicated heat flow for the whole basin is nearly six times "normal" and the anomaly above "normal" that is to be accounted for is about 16×10^6 cal per sec.

If the anomaly is on a broad regional heat-flow high that may characterize the whole Basin and Range province (Lee and Uyeda, 1965, p. 100) and the average in this province is assumed to be 1.5 times the crustal average, then 20×10^6 cal per sec of heat is flowing from an area that should have only 5×10^6 cal per sec of heat flow. On this basis, the local anomaly is four times greater than the regional average, but its excess of 15×10^6 cal per sec is not much different from the previous calculation of 16×10^6 cal per sec.

White (1957a, p. 1642) has suggested that abnormal heat flows of this type can be computed to equivalent yearly requirements of cooling and crystallizing magma. Assumptions are that the magma is granitic and initially molten at 900°C, and that it crystallizes completely as it cools to 500°C; the mean heat of crystallization of granite is 75 cal per g, and its heat capacity through the stated temperature range is ¼ cal per g°C. With these assumptions, the total available heat is 175 cal per g, or about 4.7×10^{17} cal per km^3 (density assumed 2.7). The thermal anomaly of 12×10^6 cal per sec directly related to upflow in the Steamboat Springs thermal system is equivalent to an annual requirement of 0.0008 km^3 of magma. If the indicated anomaly from the whole Truckee Meadows basin is 15×10^6 cal per sec above the regional heat flow assumed for such an area, the magma requirement is 0.001 km^3 per yr.

The age of the Steamboat Springs system is at least 100,000 years, probably more nearly 1 million years (White and others, 1964). A magma supply of batholithic proportions on the order of 100–1,000 km^3 is required to supply heat in such quantities and over such a long period of time.

A batholith intruded into the shallow crust and then remaining static as it cools and crystallizes is not a satisfactory answer to the heat-flow problem unless the fissure system for the circulating water can extend itself by some means deeper into the batholith as stored heat is removed at higher levels. One alternative that can maintain very high thermal gradients between conduits of circulating water and heat source is contingent upon convection within the magma chamber. This possibility is most attractive in providing a model that can account for a thermal anomaly over tens and hundreds of thousands of years (White, 1957a; Shaw, 1965, p. 147), but the viscosities of granitic magma are so high that convection may be precluded. Very slow convection may still occur in large masses in spite of high viscosities; actual viscosities in the crust may be lower, for some season, than indicated by laboratory-determined viscosities; or the actual magma chamber may be andesitic or basaltic, with viscosities lower than in a granitic magma.

A second alternative involves progressive downward extension of the fracture system controlling convective circulation of meteoric water. This alternative is favored if convection in the magma chamber is proved to be improbable.

Other alternatives that might also be invoked to explain an unusually high steady-state flow of heat involve radioactive decay of one or more of the elements potassium, uranium, and thorium. All available data on the rocks of the region and their contents of radioactive elements indicate no marked abnormalities on the scale adequate to account for the huge thermal anomaly.

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