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Vapor-Dominated Hydrothermal Systems Compared with Hot-Water Systems¹

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Abstract

Vapor-dominated ("dry-steam") geothermal systems are uncommon and poorly understood compared with hot-water systems. Critical physical data on both types were obtained from U. S. Geological Survey research in Yellowstone Park. Vapor-dominated systems require relatively potent heat supplies and low initial permeability. After an early hot-water stage, a system becomes vapor dominated when net discharge starts to exceed recharge. Steam then boils from a declining water table; some steam escapes to the atmosphere, but most condenses below the surface, where its heat of vaporization can be conducted upward. The main vapor-dominated reservoir actually is a two-phase heat-transfer system. Vapor boiled from the deep (brine?) water table flows upward; most liquid condensate flows down to the water table, but some may be swept out with steam in channels of principal upflow. Liquid water favors small pores and channels because of its high surface tension relative to that of steam. Steam is largely excluded from smaller spaces but greatly dominates the larger channels and discharge from wells. With time, permeability of water-recharge channels, initially low, becomes still lower because of deposition of carbonates and CaSO₄, which decrease in solubility with temperature. The "lid" on the system consists in part of argillized rocks and CO₂-saturated condensate.

increased

Our model of vapor-dominated systems and the thermodynamic properties of steam provide the keys for understanding why the major reservoirs of The Geysers, California, and Larderello, Italy, have rather uniform reservoir temperatures near 240° C and pressures near 34 kg/cm² (absolute; gases other than H₂O increase the pressures). Local supply of pore liquid and great stored heat of solid phases account for the physical characteristics and the high productivity of steam wells.

We suggest that vapor-dominated systems provide a good mechanism for separating volatile mercury from all other metals of lower volatility. Mercury is likely to be enriched in the vapor of these systems; the zone of condensation that surrounds the uniform reservoir is attractive for precipitating HgS.

A more speculative suggestion is that porphyry copper deposits form below the deep water tables hypothesized for the vapor-dominated systems. Some enigmatic characteristics of these copper deposits are consistent with such a relationship, and warrant consideration and testing.

Introduction

ALTHOUGH hot springs throughout the world have been studied for centuries, direct knowledge of their subsurface relationships was lacking until commercial and research drilling was initiated in the 20th Century. With a few notable exceptions (Allen and Day, 1927; Fenner, 1936) little significant scientific data were available prior to 1950.

Efforts to produce electricity from natural steam were first successful in the Larderello region of Italy, starting about 1904. Drilling from 1920 to 1925 showed that large quantities of natural steam could also be obtained at The Geysers in California, but economic development was not feasible until 1955. At both The Geysers and Larderello, wells deeper

than a hundred meters or so² and near centers of surface activity were found to yield slightly superheated steam (Burgassi, 1964). Some wells on the borders of the active systems³ produced hot water

² The metric system is used throughout this paper. Some readers may find useful the following conversion factors:

- Length: 1 m = 3.281 ft; 1 km = 3,281 ft = 0.6214 mi.
- Temperature: (°C × 9/5) + 32 = °F.
- Pressure: 1 kg/cm² = 0.9678 atm = 0.9807 bars = 14.22 psi. All pressures absolute, with 0.78 kg/cm² added to gage pressure for Yellowstone Park, and 1.03 kg/cm² added to gage pressure at sea level and geothermal areas at low altitudes.
- Heat: 1 cal = 3.9685 × 10⁻³ BTU; 1 cal/gm = 1.80 BTU/lb.

³ A geothermal system includes a source of heat within the earth's crust (regional heat flow or local igneous intrusion) and the rocks and water affected by that heat. When geothermal systems involve circulating waters, they are also

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and steam in noncommercial quantities and pressures (Allen and Day, 1927, p. 82); the characteristics of such wells have not yet been adequately described.

From 1946 to 1970 approximately 100 geothermal systems throughout the world were explored at depth by drilling. Initially, the objective of this search was to discover areas yielding dry steam, as at Larderello and The Geysers. This effort, however, soon revealed that most hot-spring systems yield fluids that are dominated by hot water rather than by steam.

New Zealand first demonstrated that a source of dry steam was not essential for the generation of geothermal power. At Wairakei, subsurface hot water at temperatures up to 260° C is erupted through wells to the surface; some of the water flashes to steam as temperature and pressure decrease to the operating pressure, commonly from 3 to 6 kg/cm². This steam, generally 10 to 20 percent of the total mass flow, is separated from the residual water and directed through turbines to generate electricity. The high energy potential of subsurface water has also been demonstrated in Mexico, Iceland, Japan, USSR, El Salvador, the Philippines, and the United States.

A few systems, other than Larderello and The Geysers, yield vapor with little or no associated liquid water. These include the Bagnore and Piancastagnaio fields near Monte Amiata southeast of Larderello (Burgassi, 1964; Cataldi, 1967), and probably the Matsukawa area of northern Honshu, Japan, (Saito, 1964; Hayakawa, 1969; Baba, 1968), the Silica Pit area of Steamboat Springs, Nevada (White, 1968b), and the Mud Volcano area of Yellowstone National Park, considered in this report.

Hot-water systems have attracted nearly all of the research drilling in natural hydrothermal areas. The first two research holes in the world were drilled by the Geophysical Laboratory of the Carnegie Institution of Washington in the hot water systems of Yellowstone Park in 1929-30 (Fenner, 1936), and seven of the eight research holes drilled at Steamboat Springs, Nevada, in 1950-51 (White, 1968b) were in a hot-water system. The eighth was in the small vapor-dominated Silica Pit system, subsidiary to the larger water-dominated area.

Although research drilling by the U. S. Geological Survey in Yellowstone National Park during 1967 and 1968 was aimed mainly at a better understanding of the hot-water systems of the major geyser basins, a specific effort was made to find and drill a vapor-

called hydrothermal systems. The hot part of each hydrothermal system is commonly emphasized, but in its broader meaning the marginal parts involve convective downflow of cold water, and are also included. A hot spring area is the surface expression of a geothermal system and contains hot springs, fumaroles, and other obvious hydrothermal phenomena.

dominated system. The Mud Volcano area was found to be such a system and is described here.

In spite of long and extensive commercial development at Larderello and The Geysers, the origin and nature of the systems that yield dry or superheated steam, and why they differ from the abundant hot-water systems, are not nearly so well understood. Facca and Tonani (1964), for example, seem to deny that Larderello and The Geysers differ significantly from Wairakei, New Zealand, and the other water-dominated areas. Marinelli (1969) states that Larderello is a hot-water area. James (1968) and in less detail Elder (1965) and Craig (1966) have instead proposed that the reservoirs are filled with steam maintained by boiling from a deep water table.

We submit, in agreement with James (1968), that fundamental differences do exist between two main types of natural hydrothermal systems; each type is recognizable by geologic, physical, and geochemical criteria. However, in contrast with James (1968) and others, we consider that steam and water must coexist in the reservoirs of these systems that yield dry steam at the surface.

Acknowledgments

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Summary of Characteristics of Hot-water Systems

Hot-water systems are usually found in permeable sedimentary or volcanic rocks and in competent rocks such as granite that can maintain open channels along faults or fractures. Total discharge from typical systems ranges from several hundred to several thousand liters per minute (lpm), with individual springs commonly discharging a few lpm to several hundreds of lpm. Where near-surface rocks are permeable and the surrounding water table is relatively low, much or all of the circulating hot water escapes below the ground surface, and little or none is discharged from local surface springs. For example, nearly 95 percent of the water at Steamboat Springs, Nevada, escapes in such a way (White, 1968b). On the other hand, where spring outlets are at or below the level of the surrounding water table, all hot water of the system is likely to be discharged in local visible springs.

The spring systems that discharge at low to moderate temperatures are commonly similar chemically

Why is this not a vapor down sys above water table?

to nearby ground waters, but the near-boiling hot waters of moderate to high discharge are nearly always characterized by relatively high contents of alkali chlorides, SiO₂, B, and As (table 1, anal. 4, 8, and 10; White and others, 1963, Tables 17 and 18). In confusing contrast, some gassy springs of low discharge may differ greatly from these chloride-rich waters in physical and chemical characteristics. Surrounding ground is commonly bleached and hydrothermally altered to a porous siliceous residue that may be mistaken for hot-spring sinter. The bleached ground may contain native sulfur, white, yellow, and orange sulfate minerals, and clay minerals, especially kaolinite; vegetation is generally sparse or absent. Chemical analyses of such springs (table 1, anal. 9; White and others, 1963, table 20) contrast strikingly with those of higher discharge; chloride is generally less than 20 ppm, sulfate is the dominant anion, pH is usually between 2.5 and 5, and Fe, Al, Ca, and Mg are abundant relative to Na and K.

Where these two contrasting types of springs coexist in the same general area, topographic relationships and results of shallow drilling and augering indicate that the nearly neutral to alkaline chloride springs are from the main water body, occurring where the water table intersects the ground surface. Where the water table is low, acid springs may result from boiling at this water table. Some steam condenses in cooler ground and in pools of rain water, perched ground, and previously condensed steam.

H₂S that evolves with the steam reacts near the surface with atmospheric oxygen to form sulfuric acid, thus accounting for the high sulfate contents and the low pH's characteristic of these waters. Bacterial oxidation of intermediate forms of sulfur may be involved (Schoen and Ehrlich, 1968). The acid dissolves available cations from the surrounding rocks, which are adequate sources for the reported constituents (White and others, 1963, table 20).

The geochemistry of chloride is critical in understanding the differences between the coexisting neutral-chloride and acid-sulfate waters, as well as the differences between vapor-dominated systems and hot-water systems. Most metal chlorides are highly soluble in liquid water, and the low content of Cl in most rocks can be selectively dissolved in water at high temperatures (Ellis and Mahon, 1964, 1967). The common metal chlorides, however, have negligible volatility and solubility in low-pressure steam (Sourirajan and Kennedy, 1962; Krauskopf, 1964). The only chlorides with sufficient volatility to account for significant transfer of Cl in steam at low temperatures and pressures are HCl and NH₄Cl, both of which are minor constituents of most hot-spring systems. The very low Cl content of the perched acid

springs associated with some hot-water systems is thus consistent with near-surface attainment of acidity from oxidation of H₂S, rather than by vapor transfer of HCl from initially acid sources.

The temperatures of many explored hot-water systems increase with depth to a "base" temperature (Bodvarsson, 1964a, 1970) that differs with each system that has been drilled deep enough. Temperatures at Wairakei, New Zealand, rise to a maximum of 260° C near 450 m of depth but increase little if any more at further explored depths (Banwell and others, 1957, p. 52-56), and at Steamboat Springs, Nevada, the temperatures in six drill holes were near 170° C at depths close to 100 m, but deeper drilling found no higher temperatures even though major channels were intersected below 150 m (White, 1968b). In such an area, meteoric water (Craig, 1963; White, 1968b) evidently penetrates to considerable depths along permeable channels of a huge convection system; the water is heated to its base temperature by rock conduction, perhaps augmented slightly by magmatic steam. It then rises in the core of the spring system, losing only a little heat because of its relatively high rate of upflow through wallrocks of low thermal conductivity. As the hot water rises the hydrostatic pressure decreases, and eventually a level is attained where pressure is low enough for boiling to begin.

Of about one hundred hot-water systems throughout the world that have now been explored by drilling, fewer than 30 are known to exceed 200° C in temperature and only about 10 demonstrably exceed 250° C. The liquid of the two reservoirs known to exceed 300° C is brine rather than relatively dilute water. The Salton Sea system has about 250,000 ppm of dissolved salts and a maximum temperature of about 360° C (Helgeson, 1968). The Cerro Prieto system, about 90 km to the south in Baja California, Mexico, has a salinity of about 17,000 ppm and temperatures as high as 388° C (Mercado, 1969). → 25 wt. %
→ 1.7 wt. %

Hot-water systems have a high potential for self-sealing (Bodvarsson, 1964b; Facca and Tonani, 1967) by means of deposition of minerals in outlet channels. SiO₂ is the most important constituent for the self-sealing of high-temperature systems because quartz is so abundant and its solubility increases so much with temperature (Fournier and Rowe, 1966). Quartz dissolves rather rapidly at high temperatures; when quartz-saturated waters are cooled, quartz precipitates rather readily down to about 180° C but with increasing sluggishness at lower temperatures. The SiO₂ content of many waters, after cooling, greatly exceeds the solubility of quartz and may even exceed the solubility of amorphous SiO₂. Near the surface where temperatures are near or below

Table 1.--Chemical analyses of waters associated with vapor-dominated and hot-water geothermal systems

Name	1/ The Geysers	2/ The Geysers	3/ GS-7	4/ Spring 8	5/ Mud Volcano	6/ Mud Volcano	7/ Y-11, Mud Volcano
Location	Calif.	Calif.	Steamboat, Nev.	Steamboat, Nev.	Yellowstone, Wyo.	Yellowstone, Wyo.	Yellowstone, Wyo.
Water type	HCO ₃ -SO ₄	Acid-sulfate	HCO ₃ -SO ₄	Cl-HCO ₃	Acid-sulfate	HCO ₃ -SO ₄	HCO ₃ -SO ₄
System type	Vapor-dom.	Vapor-dom.	Vapor-dom.	Hot water	Vapor-dom.	Vapor-dom.	Vapor-dom.
SiO ₂	66	225	14	293	540	215	
Al		14		0.5	146		
Fe		63		0.05	17		
Mn		1.4		0.05			
As				2.7			
Ca	58	47	6.3	5.0	14	28.7	28
Mg	108	281	0	0.8	11	16.4	0.47
Na	18	12	9.3	653	16	74.3	105
K	6	5	4.5	71	17	47.5	12.6
Li			0	7.6		.20	.18
NH ₄	111	1,400		<1	26	.18	3.2
H		9.5			43		
HCO ₃	176	0	21	305		298	258
CO ₂	--	--	--	--	--	--	
SO ₄	766	5,710	24	100	3,149	65.3	74
Cl	1.5	0.5	0.5	865	Tr.	13.5	9.6
F			0	1.8	1	2.0	
Br				0.2			
NO ₃			Tr.	--			0.2
B	15	3.1	1.3	49		.6	0.1
H ₂ S	0	--	2.4	4.7	0		
Total reported	1,330	7,770	83	2,360	3,980	761.7	491.4
pH	neutral	1.8±	6.5	7.9	Strong acid	7	8.5(?)
Temperature °C	100	Boiling?	161	89.2	65	58.5	131.7

^{1/}Witches Cauldron, White and others, 1963, p. F47, modified from Allen and Day, 1927.

^{2/}Devils Kitchen, White and others, 1963, p. F46, modified from Allen and Day, 1927.

^{3/}White and others, 1963, p. F47. Condensate inapor-filled hole.

^{4/}. . . Do p. F40.

^{5/}Allen and Day, 1935, p. 427; described as "Big Sulphur Pool" 0.3 km N of Mud Volcano; location indicates Old Sulphur Cauldron of fig. 4, 60 m SSW of Y-11 drill hole.

^{6/}Spring discharging from sinter, E. bank of Yellowstone River 0.5 km SE of Y-11 drill hole; has deposited sinter in recent past, if not now. Analyzed by Mrs. Roberta Barnes.

^{7/}Erupted from Y-11 drill hole Sept. 22, 1969 after hole had caved to 28 m depth (table 1); collected by R. O. Fournier, analysis by Mrs. Roberta Barnes. pH not representative of in-hole environment because of CO₂ loss, storage in plastic with clays.

Note: The wordapor should read vapor in footnote 3.
Reference to Table 1 in footnote 7 should read Table 4.

Also form pyrophyllite?

100° C, the excess silica in such waters may precipitate as chalcedony, opal, and cristobalite (White and others, 1956). Self-sealing by silica minerals is likely to be slight in hot-water systems that do not

exceed 150° C, but as maximum temperatures increase above this value, the potential for self-sealing increases greatly.

Calcite, zeolites, and some other hydrothermal

Table 1.--Chemical analyses of waters associated with vapor-dominated and hot-water geothermal systems (continued)

Name	8/ Norris Basin	9/ Norris Basin	10/ Well 4	11/ Well 5	12/ Carboli A,	13/ Well MR-1
Location	Yellowstone, Wyo.	Yellowstone, Wyo.	Wairakei, N.Z.	Wairakei, N.Z.	Italy	Matsukawa, Japan
Water type	Cl(HCO ₃)	Acid sulfate	Cl	HCO ₃ SO ₄	SO ₄ HCO ₃ (Cl)	SO ₄ (HCO ₃)
System type	Hot water	Hot water	Hot water	Vapor-dom.(?)	Hot water	Vapor-dom.(?)
SiO ₂	529	109	386	191		635
Al		2.4			Trace	29
Fe		0.8			Trace	508
Mn						
As	3.1					
Ca	5.8	2.2	26	12		
Mg	0.2	0	<0.1	1.7	5.0	8.7
Na	439	2.0	1,130	230	56.6	264
K	74	3.0	146	17	32.0	144
Li	8.4		12	1.2		
NH ₄	0.1	30	0.9	0.2	19.0	
H		14				
HCO ₃	27	--	35	670	89.7	37
CO ₃	--	--	0(?)			
SO ₄	38	758	35	11	137.4	1,780
Cl	744	15	1,930	2.7	42.6	.12
F	4.9		6.2	3.7		
Br	0.1					
NO ₃	--					
B	12	6.9	26	0.5	13.9	61.2
H ₂ S	.0		1.1	0		Trace
Total reported	1,890	943	3,750	1,140	396.2	3,478.9
pH	7.5	1.97	8.6	6.7		4.9
Temperature °C	84.5	90	228+	High	-300	-240

^{8/}Dr. Morey's Porkchop, 60 m southwest of Pearl Geyser (White and others, 1963, p. F40).

^{9/}Locomotive Spring, 55 m WSW of Norris Basin drill hole of Fenner (1936); seeping discharge (White and others, 1963, p. F46).

^{10/}Typical of shallow Wairakei system; 375 m deep with maximum temperature of 245°C (Banwell and others, 1957). Analysis by Wilson; also contains 11 ppm free CO₂ (Wilson, 1955; quoted in White and others, 1963, p. F40).

^{11/}Western part of Wairakei field (Wilson, 1955, quoted in White and others, 1963, p. F47). Similar to some waters of vapor-dominated systems; 467 m deep, maximum 217°C at 271 m.

^{12/}Deepest well of hot-water field on So. border Larderello steam fields (Cataldi and others, 1969). Orig. anal. in ppm, supplied by R. Cataldi, 1970.

^{13/}Well 945 m deep, produced steam, some water for 1 year before drying; this anal. while still wet: condensate of steam 50 ppm H₂S and 6.2 ppm S (Nakamura and Sumi, 1967; Hayakawa, 1969).

minerals are also effective in producing self-sealed margins of some hot-water systems, but generally less so than the silica minerals. Self-sealing is likely to

be most extensive where temperatures decrease most rapidly. These marginal parts are of secondary interest for production drilling, and they have not been

Table 2.--Analyses of gases associated with vapor-dominated and hot-water geothermal systems, in volume percent

	Total vapor, including H ₂ O		Gases, excluding H ₂ O					
	1/ The Geysers, California	2/ Larderello, Italy	3/ The Geysers, California (1), recalc.	4/ Larderello Italy (2), recalc.	5/ Y-11 Mud Volcano, Yellowstone	6/ Mud Volcano Yellowstone	7/ Y-9, Norris, Yellowstone	8/ Spring Norris Yellowstone
H ₂ O	98.045	98.08						
CO ₂	1.242	1.786	63.5	93.02	98.4	98.90	91.5	97.40
H ₂	0.287	-0.037	14.7	} 1.92	<0.01	0.00	0.9	0.00
CH ₄	0.299		15.3		Tr.	0.10	0.1	0.20
C ₂ H ₆					0.0		0.0	
N ₂	} 0.069	0.0105	} 3.5	0.54	0.8	} 1.00	5.1	} 1.60
A					0.013		0.08	
H ₂ S	0.033	0.049	1.7	2.55	0	0.10	1.4	0.75
NH ₃	0.025	0.033	1.3	1.72				
H ₃ BO ₃	0.0018	0.0075	0.09	0.39				
O ₂					0.2	0.00	1.0	0.05?
Total	100.002	100.003	100.09	100.14	99.42	100.10	100.08	100.00

1/ Well 1, The Geysers (Allen and Day, 1927, p. 76).

2/ Average vapor from producing wells (Burgassi, 1964), recalculated from analysis in gm per kgm; 2,850,000 kg produced per hour; also contains 1 cm³ total rare gases per kg.

3/ Recalculated from 1/, without H₂O.

4/ Recalculated from 2/, without H₂O.

5/ Collected July 10, 1968, by R. O. Fournier, when hole was still open to 316 ft (table 3).

6/ Gas from same spring as anal. 5 of table 1 (Allen and Day, 1935, p. 86).

7/ Collected by R. O. Fournier, Sept. 18, 1969, and analyzed by D. Byrd, U.S. Geol. Survey; gas separated from water; nearest drill hole to springs of anal. 8 and 9, table 1.

8/ Gas from unnamed acid-sulfate spring "near Congress Pool," perhaps Locomotive (table 1, anal. 9). Allen and Day, 1935, p. 86, 469.

Note: Reference to Table 3 in footnote 5 should read Table 4.

cored and studied in much detail except in research drilling in Yellowstone Park (unpublished data).

For similar geochemical reasons, most hot-water systems with subsurface temperatures of 180° C or higher (White, 1967a) have hot springs or geysers that deposit sinter (amorphous silica precipitated on the ground surface by flowing hot water). Waters that deposit sinter nearly always have SiO₂ contents of at least 240 ppm, equivalent to a quartz-equilibration temperature of 180° C. Because the solubility of amorphous SiO₂ is so much higher than that of quartz, a quartz-saturated water at 180° C must cool to about 70° C in order to precipitate amorphous silica. If the water becomes sufficiently concentrated in SiO₂ by evaporation, as on the borders of pools and in erupted geyser water, precipitation can occur at somewhat higher temperatures.

The existence of sinter, as distinct from travertine

(CaCO₃) and siliceous residues from acid leaching, is evidence for a hot-water system with present or past subsurface temperatures of more than 180° C.

Summary of Characteristics of Vapor-dominated ("Dry-steam") Systems

The near-surface rocks of Larderello, Italy, and The Geysers, California, are relatively tight and incompetent, and evidently do not permit large quantities of meteoric water to penetrate deep into their systems (White, 1964). Even in these areas, however, isotopic data indicate that most of the water is of surface origin (Craig and others, 1956; Craig 1963).

Surface springs at The Geysers* typically have

* "The Geysers" is an unfortunate misnomer. The area has never had true geysers, which are restricted to the hot-water systems (White, 1967a).

43 bars/500m

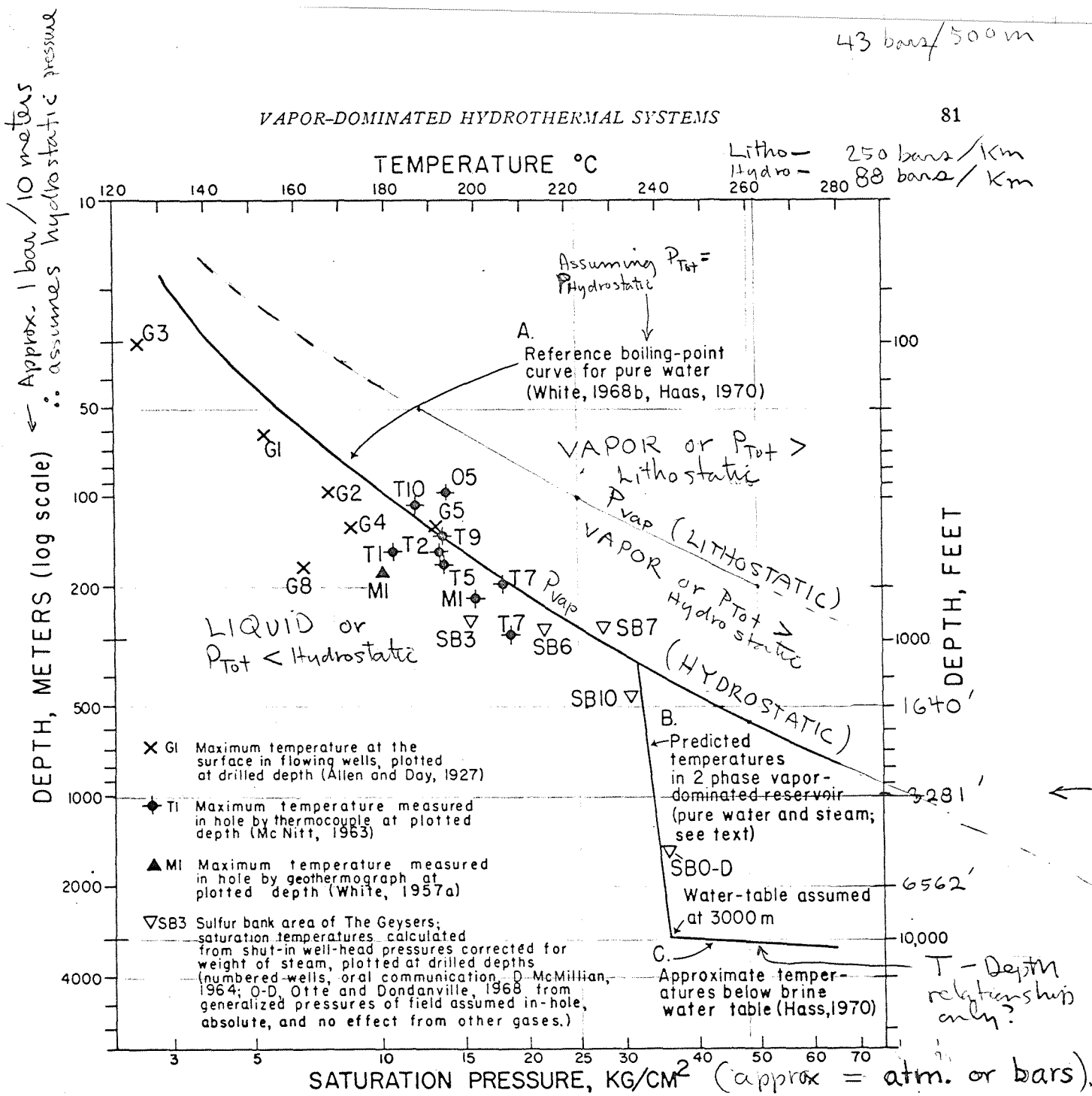


FIG. 1. Measured and calculated temperatures from The Geysers, Calif., with some theoretical curves. The reference boiling-point curve for pure water, curve A, differs in shape from its usual representation because of the logarithmic scale for depth. Note that curves B and C are temperature-deficient and pressure-deficient relative to curve A.

very low discharge, totaling little more than 100 lpm (Allen and Day, 1927). Most of the springs are strongly acidic (pH from 2 to 3). The few neutral springs (Table 1, anal. 1) have chloride contents of less than 2 ppm, similar to local rain water. A careful search of the creek that flows through the area was made on the chance that undetected chloride springs might be seeping into the creek (White, 1957a, p. 1651). However, throughout an area of at least 30 square miles surrounding The Geysers, the

surface and ground waters are no higher in chloride than normal cold streams.

Chloride contents have not been included in reports on natural springs associated with the original vapor-dominated Larderello fields, but available descriptions of spring activity, dominated by mud pots and fumaroles, suggest the presence of sulfate waters low in chloride. However, present springs are not low in pH (R. Cataldi, written commun., 1970), perhaps because of the neutralizing action of abun-

0°C = 273.18°K
 $T_c = 374.15^\circ\text{C} = 647.33$

H₂O
 $\Delta H_{\text{vap.}} = 539.6 \text{ cal/gm}$
 @ 1 bar & 100°C

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374.15°C

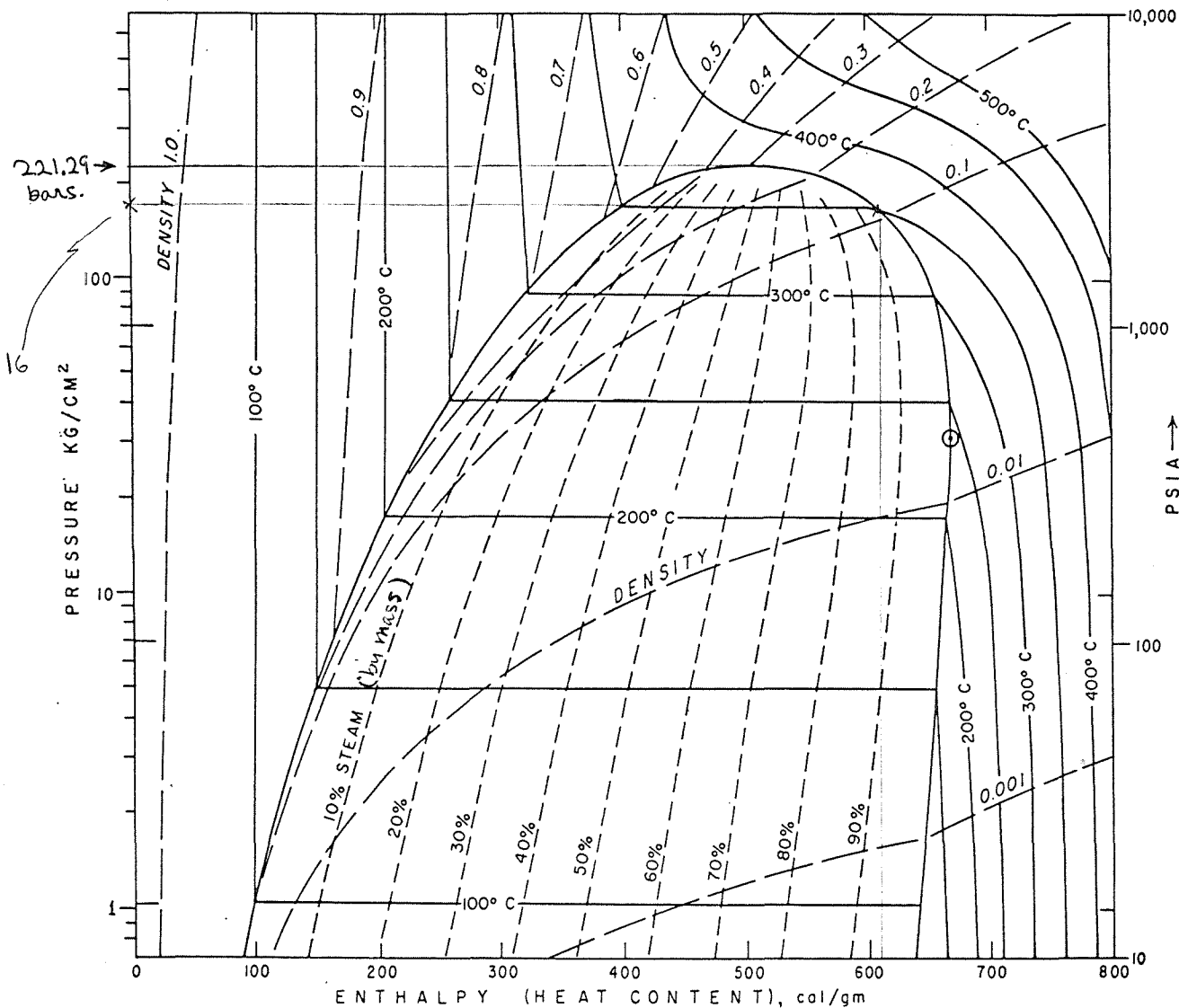


FIG. 2. Pressure-enthalpy diagram for pure water and vapor, showing contours of equal temperature, density, and mass proportions of steam to water (computed from Keenan and Keyes, 1936). Open circle indicates maximum enthalpy of saturated steam, 670 cal/gm at 236°C and 31.8 kg/cm².

dant NH₃ absorbed from the gases. Some springs and wells of the Carboli area just south of the vapor-dominated fields (Cataldi and others, 1969) contain some chloride (42.6 ppm, Table 1). Although this Cl content is not notably high, it is consistent with the abundant water and old travertine which suggest that Carboli is a hot-water system.

In general where surface springs are all low in chloride and subsurface thermal waters are similarly low (< 20 ppm) a vapor-dominated system is indicated. The Cl content of steam is normally less than 1 ppm, but near-surface waters involved in condensation of the steam commonly contain a few ppm of Cl because, with little or no discharge, Cl can be selectively concentrated.

Typical wells at Larderello (Burgassi, 1964) produce dry or slightly superheated steam with 1 to 5 percent of CO₂ and other gases (Table 2, anal. 2). Liquid water evidently occurs in some noncommercial wells on the borders of the fields. Shut-in well-head pressures in typical steam wells tend to increase with depth up to a maximum of about 32 kg/cm² (Penta, 1959; Burgassi, 1964). Increased productivity reported at greater depths evidently is not due to significantly higher initial pressures. Ferrara and others (1963) list the temperatures of two Larderello wells as 251°C, but all other cited wells are 240°C or lower (depths not given).

Typical wells at The Geysers also produce dry or superheated steam containing gases similar to those

in the Larderello field (Table 2, anal. 1). Pressures up to about 35 kg/cm² were measured in the deeper wells (500 psi, Otte and Dondanville, 1968), but whether pressures were at the well-head or in-hole, and gage or absolute were not specified.

Figure 1 shows the maximum temperatures measured or calculated for individual shallow wells in The Geysers field. For a variety of reasons each point is individually unreliable and is probably not identical with the original ground temperature at its plotted depth. Nevertheless, temperatures of shallow wells (< 350 m) do show a rather close relationship to curve A, the reference boiling-point curve for hydrostatic pressure of pure water. A few points plot above this curve, indicating ^{continuing} pressures above hydrostatic but below lithostatic.

All of the early shallow wells at The Geysers were drilled in or near fumaroles, hot springs, and hydrothermally altered ground that provided evidence of surface discharge of thermal fluids. Figure 1 suggests, and our model (to be discussed) assumes, that liquid water condensed from rising steam fills much of the pore spaces; this condensed water provides a major buffering control over temperatures and pressures in the zone of upflowing fluids. McNitt (1963) concluded from other data that a near-surface zone is water saturated; we support his general conclusions but disagree on the nature of the evidence.

Although available data are scanty, temperatures at The Geysers increase irregularly with depth, probably along or near the hydrostatic boiling-point curve, and temperatures near 236° C (and pressures near 32 kg/cm²) are attained, with only slight additional increases approximately along curve B of figure 1 to explored depths. In the Sulphur Bank area of The Geysers (Otte and Dondanville, 1968), about 1½ km west-northwest of the original field, wells range from 450 m to more than 2,000 m in depth and are remarkably uniform in temperature (close to 240° C) and in pressure (about 35 kg/cm²), as shown in Figure 1. Otte and Dondanville state that "the fluid exists in the reservoir as superheated steam," but the reported temperatures and pressures indicate approximate saturation. No specific data for individual wells are available.

No data have been published to indicate that wells in the central parts of any vapor-dominated field have penetrated a deep water-saturated zone or a water table. In such a penetration, in-hole pressures should increase downward through the water-filled parts of shut-in wells instead of remaining near 32 kg/cm². This evidently does occur in parts of the Italian fields (R. Cataldi, written commun., 1970), but detailed relationships are not yet available. The expected temperature-depth relationships below the

Table 3.--Pressures and temperatures in a two-phase

reservoir in which steam is the continuous phase.

Top of reservoir assumed to be 236°C 31.8 kg/cm²,

and 360 m deep (from hydrostatic boiling-point curve).

Depth meters	Pressure, kg/cm ² (bottom hole)	Temperature °C
360	31.8	236.0
500	32.0	236.1
1,000	33.5	239.0
1,500	34.3	240.3
2,000	35.1	241.6

Note: Second line of column 3 should read 236.3.

deep water table are shown in Figure 1, curve C. This curve has an increasing slope with depth and all points on it are also deficient in pressure with respect to external water pressures, probably to depths of 2,000 m or more below the water table.

James (1968) noted that initial temperatures and pressures of the Larderello steam fields were close to the temperature (236° C) and pressure (31.8 kg/cm²) of saturated steam of maximum enthalpy (670 cal/gm; indicated on Fig. 2). James reasoned that enthalpies up to this maximum can be obtained in undisturbed steam reservoirs by evaporation at a subsurface water table. Higher temperatures (and pressures) can exist below but not *at* the water table. He reasoned that if saturated steam at 350° C and 468.7 kg/cm², for example, with an enthalpy of only 612 cal/gm (Fig. 2), formed deep in a system and rose up to levels of lower hydrostatic pressure, part of the steam would increase in enthalpy as it continued to rise while the rest would condense to liquid water and remain behind. For a pure water system, this separation of liquid from vapor continues until the pressure at maximum enthalpy is attained.

The enthalpy of saturated steam near its maximum, however, is not very sensitive to changes in temperature and pressure (Fig. 2). James suggested that the top of a natural vapor-dominated reservoir is likely to have a temperature near 236° C and a pressure near 31.8 kg/cm² but that, because of the weight of steam in a deep reservoir, the temperature near a boiling water table may be as much as 240° C at a pressure near 34 kg/cm². Table 3 shows expected depth-related variations in temperature and pressure of a pure water system in a homogeneous, vapor-dominated reservoir.

Why?
Because it would remain saturated but would physically separate from the condensate.
∴ would remain 100% sat. steam & would approach max. H.

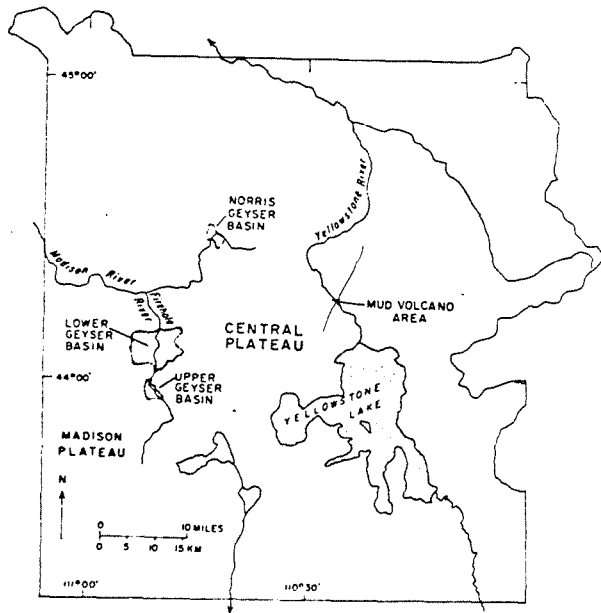


Fig. 3. Index map of Yellowstone National Park, Wyoming, showing location of Mud Volcano area and the major geysers basins.

The scanty available data suggest that temperatures and pressures may exceed the limits suggested by James because of the effects of dissolved salts and the partial pressures of other gases. In addition, although

[REDACTED], we see no fundamental reason why the available heat supply may not form somewhat more steam than can escape at these pressures through available channels. In this paper we shall assume James' suggested range in temperatures and pressures as the most probable, but we emphasize that more precise data are essential in understanding the detailed characteristics of these systems.

Recorded temperatures of the vapor-dominated reservoirs are significantly lower than in some hot-water fields, which range up to 388° C (Mercado, 1969). The Carboli field on the southern edge of the Larderello steam fields is notable in being the only described field in the Larderello region that produces more water than steam by mass and thus is a hot-water system. Its maximum temperature is about 300° C (Cataldi and others, 1969), which clearly exceeds all temperatures reported from the vapor-dominated areas.

The Mud Volcano Area, Yellowstone Park

General Setting.—The Mud Volcano area is located along the Yellowstone River about 8 km north of Yellowstone Lake (Fig. 3). Bedrock of the area

is rhyolitic ash-flow tuffs erupted approximately 600,000 years ago (R. L. Christiansen and J. D. Obradovich, 1969, written commun.). Glacial gravels and sands of Pinedale age (about 25,000 to 12,000 years B.P.) mantle the bedrock except near the center of the area.

Thermal activity in the Mud Volcano area consists almost entirely of vigorously bubbling mud pots, acid-sulfate springs, and steam vents concentrated on north-northeast lineaments. Total discharge is only about 80 lpm (Allen and Day, 1935, p. 58) from an area of 2½ km². There are no chloride-rich springs like those of the major geyser basins, even along the Yellowstone River, which is the local base level for the water table of the area. Instead, acid-sulfate and nearly neutral bicarbonate-sulfate springs occur along the river (anals. 5 and 6, Table 1). A little silica is being deposited by evaporation from algal mats at two of these nearly neutral springs, and opal-cemented Holocene alluvium is common along the riverbanks. Although none of the present springs has enough silica to deposit hard sinter from flowing water on the surface (generally requiring at least 240 ppm SiO₂), three small areas of old sinter occur as much as 3 m above river level. This indicates that sometime in the past 12,000 years silica-rich water, presumably also rich in chloride, discharged at the surface in the Mud Volcano area.

Acid-sulfate springs similar in discharge and chemistry to the Mud Volcano springs occur locally where H₂S is abundant in high ground of the major Yellowstone geyser areas (anal. 9, Table 1). However, in contrast to drill hole Y-11 in the Mud Volcano area (anal. 7, Table 1), all drill holes in the geyser basins tapped water rich in chloride and similar to waters from the geysers and the principal flowing springs (anal. 8, Table 1).

Y-11 was drilled by the U. S. Geological Survey at the north end of the Mud Volcano area, 75 m north-northeast of Old Sulphur Cauldron. Figure 4 shows the locations of the hole and the "tree line" inside of which trees do not grow because temperatures are too high. Also shown are two heat-flow contours mapped by snowfall calorimetry (White, 1969). The 900 µcal/cm² sec (microcalories per sq cm per second) contour is probably within 20 percent of the existing total conductive and convective heat flow. This heat flow is about 600 times the world-wide average conductive heat flow of the earth (Lee and Uyeda, 1965). The 5,000 µcal contour is less precisely located, but total heat flow obviously increases rapidly southeast from Y-11 drill hole.

Near-surface Ground Temperatures.—Relationships between heat flow, depth, and temperature determined in shallow auger holes near Y-11 clarify some principles of major significance to the vapor

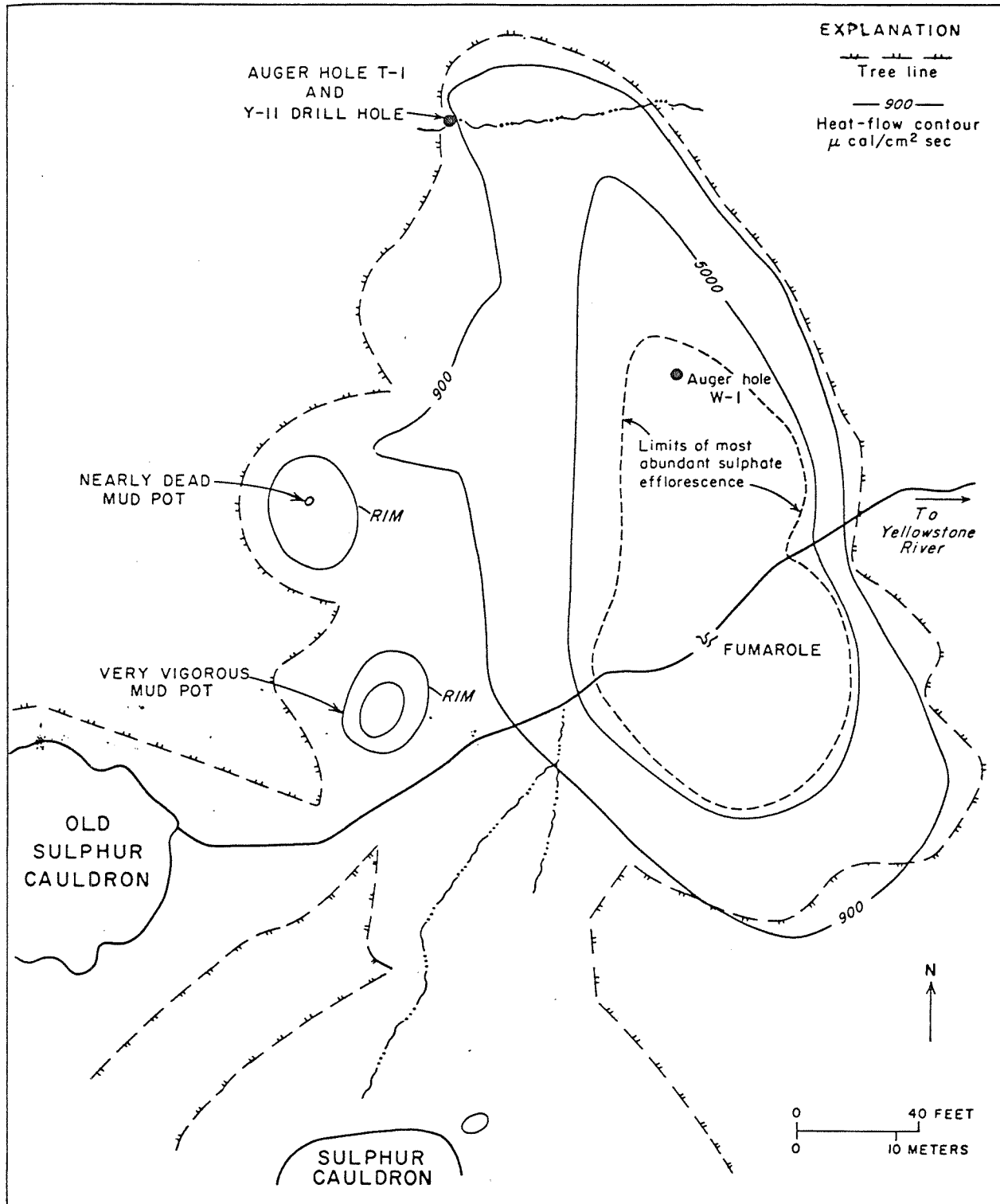


FIG. 4. Sulphur Cauldron area, north end of Mud Volcano area, showing location of Y-11 drill hole relative to heat flow and other features.

$P_{vap} = X P_{atm}$
actual *pure*

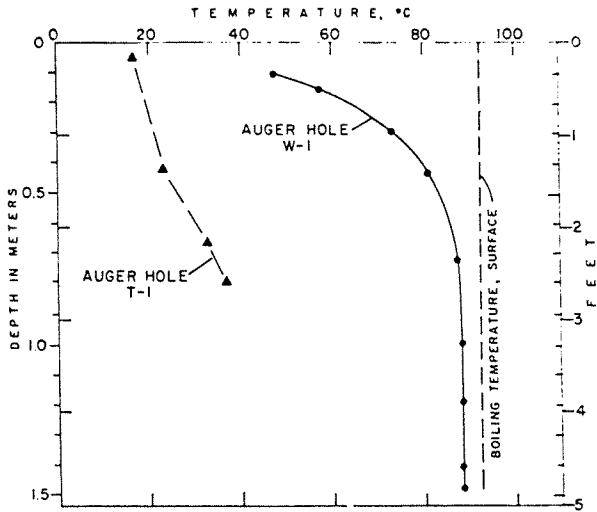


FIG. 5. Temperature-depth curves in shallow auger holes in ground with dispersed upflow of steam and other gases.

dominated systems. Hole T-1 (Fig. 5) was augered on the Y-11 site just prior to drilling, and hole W-1 was augered 35 m to the southeast (Fig. 4). The near-surface temperature at any given depth increases abruptly to the southeast, correlating with increasing heat flow.

Temperatures in W-1 increased rapidly with depth to about 3/4 m, where they leveled off at 88.2° C. From 1.0 to 1.55 m there was no temperature change. Consequently, heat cannot be transferred by conduction through this interval, and all heat that flows out

at the surface must be transferred in steam and other gases through the no-gradient zone. Total heat flow at the surface of W-1 auger hole has not been measured by snowfall calorimetry, but extrapolation of data on Figure 4 suggests a heat flow of perhaps 10,000 $\mu\text{cal}/\text{cm}^2 \text{ sec}$.

The leveling off of temperatures in W-1 at 4.1° C below the boiling temperature of pure water (92.3° C at this altitude) is due to the high content of CO₂, H₂S, and other gases in the rising vapor. The vapor pressure of water at 88.2° C is 491 mm of Hg, but the atmospheric pressure averages about 572 mm of Hg. Thus 14 percent of the total vapor pressure results from the partial pressures of other gases. At a depth where the temperature is 85° C, 25 percent of the total pressure is due to residual gases (143 mm of total Hg pressure); similarly, 50 percent consists of other gases at 75° C, 90 percent at 40° C, and 97.7 percent at 15° C.

The depth at which the temperatures level off is dependent on the heat flux from below, the thermal conductivity of the soil, the air-ground interface temperature, and the amount and nature of precipitation of the preceding few days or weeks. If the rate of upflow of steam increases sufficiently, a surface fumarole is produced. If, in contrast, the rate of upflow decreases, complete condensation occurs at a greater depth appropriate to the thermal conductivity heat flow, and surface temperature.

In the steam-gas mixture in W-1 auger hole, no steam condenses between a depth of 1.0 m and the bottom of the hole because of the absence of a tem-

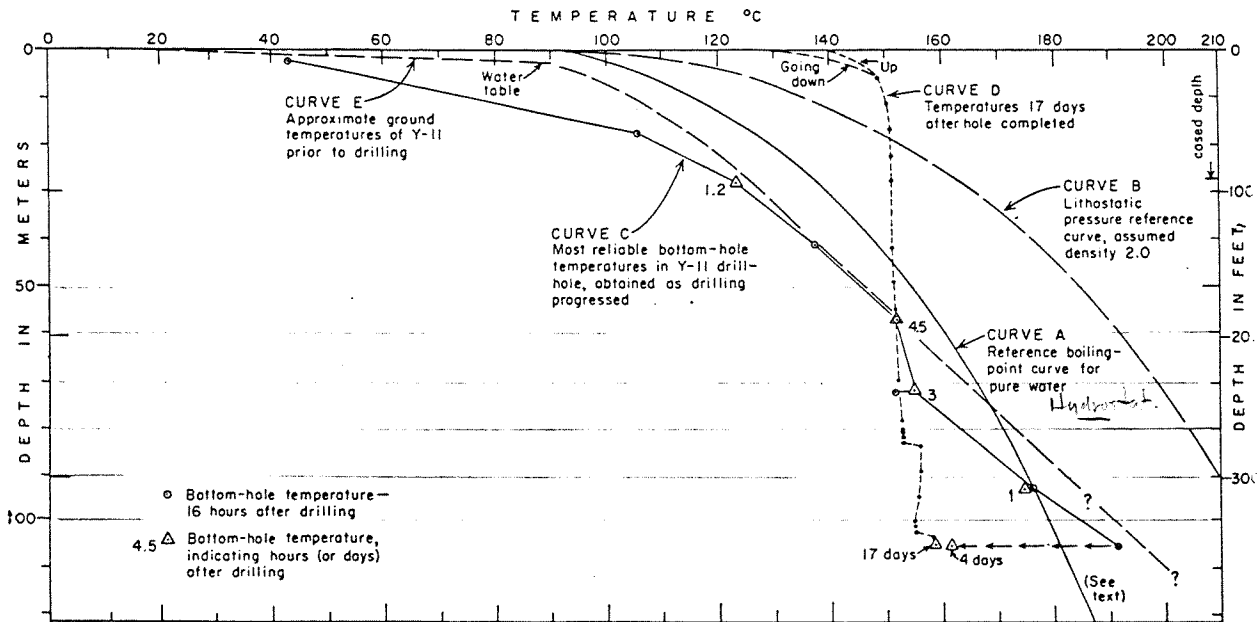


FIG. 6. Temperatures in Y-11 (Sulphur Cauldron) drill hole, Mud Volcano area, Yellowstone Park, Wyoming.

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 May

dissolved solids (Non-vol. solutes) would offset, in part at least, the effects of the volatile gases.

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 1 m
 pera

Table 4.--Temperatures, pressures, and other data from Y-11 (Sulphur Cauldron) drill hole, Mud Volcano area, Yellowstone National Park, Wyoming

Underlined data considered most reliable

Date and time 1968	Observation depth, m	Temperature, °C	Depth to water m	Total pressure kg/cm ²	Comments
May 15 8:35A	2.0	<u>43.0</u> ?			Drilled to 6.1 m, set 4 in. casing, and cemented on May 14; on cement at 2.0 m, temp. probably minimum.
12:10	8.3	(36.0)	2.3		80 min. after circulation ceased; good water level 1 hr. after circulation ceased.
3:25	18.1	87.0		0.85	Drilled 18.3 m; pressure all gas.
16 8:15A		104.4	2.3	1.17	Do.
8:23	<u>17.3</u>	106.4		<u>105.4</u>	1.08
1:25P	27.7	<u>123.3</u>		2.05	1.2 hrs. since circulation; could have erupted; drilled, set 27.4 m 3 in. casing and cemented.
17 8:00A	13.7	73.0			On cement; temperature probably minimum.
4:45P	(41.8)	(116.1)	5.2		Drilled 41.8 m; lost circulation 37.2-41.8 m.
18 7:50A		137.0	4.9		Gas pressure from outside rods, 1.56 kg/cm ² temperature at water level inside rods 73°C.
8:08A	11.4	137.2		<u>137.1</u>	
1:00P	(57.0)	(120.3)	4.7		Lost circulation 41.8-57.0 m; temp. 1-1/2 hrs. after circulation.
19 10:40A		152.4	4.6		23 hrs. after circulation.
11:00	56.9	151.4		<u>151.9</u>	
20 7:54A	57.0	151.4			45 hrs. after circulation.
8:05	57.0	153.4			
8:12		150.9		<u>151.9</u>	
8:22		151.9	0.3		
2:00P	72.2	<u>154.5</u>		0.8	3 hrs. after circulation, pressure fluctuating; water discharged outside of rods; drilled 72.3 m.
21 7:45A		151.1	72.2	3.5-3.6	Max pressure; note temp. decrease since May 20. Drilled to 93.6 m; erupted after pulling core; nearly all steam after much initial water.
7:57		152.0		<u>151.6</u>	
12:20P		<u>174.3</u>		5.8 then to 4.1	Temp. 40 min. after eruption; some water with steam at 4.1 kg/cm ² .
3:45P	93.4	174.7		7.8 down to 5.9	Leaking steam at 7.8 kg/cm ² , then down to 5.9 kg/cm ² , some water.
3:50				4.4	Pressure on side valve, outside drill rods.
22 8:00A		175.7		10.1 down to 8.0	Leaking vapor only.
8:12	<u>93.4</u>	176.1			Vapor and a little water; drilling increasingly difficult >100 m. Violent eruption at 105.7 m, initially much water (drill water?), then mostly steam.
23 8:15A		191.6	105.7	12.7 to 11.2	Leaking vapor only. Drill rods in hole a few feet off bottom; exact depth not noted.
8:30A		191.1			
11:00A				5.0	Rods pulled, pumping cold water down outside rods throughout; pressure with open hole >27.4 m. Hang up at about 33.6 m, erupted to clear--powerful steam eruption but little water.
27 Not noted	105.5	161.1		5.3	Note major permanent changes in temperature and pressure after rods out of hole.
		161.4		5.4	
June 10 10:00A	-105	<u>158.6</u>		5.3 to 5.4	Thermistor temperature series plotted on fig. 6. Temperature generally steady and reproducible down to 84.1 m, fluctuating somewhat at greater depths, up to 4°C at bottom (maximum is plotted).
16	96.3			5.0	R. O. Fournier attempted to sample; filled with vapor to existing bottom.
28	39.0			4.6	Blocked; no access to greater depths; no water to 39.0 m.
Sept. 1969 21	28.0	<u>131.7</u>		3.2 to 3.5	Attempting thermistor series; initial temperature at top 75°C, increasing to 107°C with leakage of gas. Hole filled with gas to cave at 28.0 m just below casing; thermistor wedged and lost. Erupted gas, mud, and water, and collected water sample. Pumped in 5 sacks of cement at pressures up to 11.3 kg/cm ² .

perature gradient. As steam rises above a depth of 1 m, however, a little starts to condense as a temperature gradient first becomes evident. The gra-

dient increases upward as the surface is approached, so more water vapor can condense. The residual gases are progressively concentrated upward as H₂O

is condensed, the velocity of upflow consequently decreases, and a correspondingly smaller proportion of the total heat is transported by water vapor. Convective transport of heat at the air-ground interface must be largely in the residual gases, but water vapor, even though a minor constituent, is still a significant transporter of heat because of its high heat of vaporization (588 cal/gm at 15° C), relative to heat content of other gases.

The water vapor that condenses between 1.0 m and the surface at W-1 percolates downward against the flow of steam. The ground is unsaturated with liquid at the bottom of the auger hole and probably to the local water table (2.3 m in Y-11 drill hole). Below the water table at W-1, pressures must exceed atmospheric, and temperatures probably rise along or near the hydrostatic boiling point curve of Figure 6.

The near-surface temperature gradient in auger hole W-1 of Figure 5 is much higher than in T-1, as we should expect from the heat-flow contours of Figure 4. Projection of the T-1 gradient downward to the water table at 2.3 m suggests that temperatures were slightly below boiling at this depth. It appears that only a little water vapor and other gases were rising at the Y-11 site prior to drilling, and most heat was being transferred from the water table to the ground surface by conduction.

Physical Measurements Made During Drilling of Y-11.—Data from Y-11 are summarized in Table 4, and bottom-hole temperatures are plotted in Figure 6. The bottom-hole temperatures considered to be most reliable are connected by a solid line. Much effort was made to obtain reliable data from Y-11 as drilling progressed, in part because of the paucity of such data from the large commercial vapor-dominated systems. Because of the high cost of drilling and other factors, available data from the commercial systems are entirely restricted to completed wells, and almost no data are obtained at shallow and intermediate depths as drilling progresses.

In the recent holes drilled in Yellowstone National Park, temperatures measured at each temporary bottom, just before resumption of drilling (generally after overnight shut-down of about 16 hours), provided reasonable approximations of pre-drilling ground temperatures; they are far superior to temperature profiles measured in completed holes (White, Fournier, Muffler, and Truesdell, unpublished data). Measured bottom-hole temperatures in Y-11, however, are less reliable than in the other holes but are considered to be within a few degrees of original ground temperature. At depths less than 27.4 m, rapid drilling plus the setting of two strings of casing prevented acquisition of reliable data. From 37.2 to 79.3 m, all drill water was lost into the

ground, and at greater depths only about 50 percent returned to the surface. Despite the apparent high permeability and loss of drill water, however, the temperature of 151.9° C at 57.0 m depth is probably reliable because it was repeated on successive days with no disturbance by drilling.

A temperature profile made in the open hole 17 days after completion is shown on Figure 6 (curve D). It differs greatly from the temperature profile obtained as drilling progressed. The temperatures from 12.2 to 83.5 m were almost constant, rising only 2° or so, to 153° C at 83.5 m. At greater depths, rapid fluctuations of 1° to 5° were observed. These fluctuations were not due to instrumental defects and were far too large and too rapid to be caused by only a vapor phase; coexistence of steam and water is thus indicated from 83.5 to 103.7 m. The pressure of saturated steam at 153° C is 5.3 kg/cm², which is very close to the measured well-head pressure, 5.4 kg/cm². The temperature of 158.2° C at 103.7 m, however, is not consistent with the well-head pressure, unless liquid water was present near the bottom of the hole. From drill records, we conclude that water was probably entering the hole from depths as shallow as 58 m or less, while an upward flow of steam dominated the central part of the casing. Detailed relationships that existed during the thermistor measurements between 83.5 and 103.7 m cannot be deciphered completely. Evidently some steam was flowing in near 84.2 and 103.7 m. Water seeping down from higher levels did not accumulate extensively but was either forced out into permeable walls or was evaporated by the higher temperature steam. At shallow depths in the hole, horizontal and vertical temperature gradients were so high that most water vapor condensed and residual gases were concentrated, as in auger hole W-1. The condensed water trickled down the walls of the casing.

On several occasions during the drilling of Y-11, we were unable to prevent the hole from erupting for short intervals. The eruptions differed notably, however, from those in holes in the hot-water systems of the geyser basins. In drill holes in permeable rocks, with adequate water supply, and a temperature of 160° C, for example, only 11 percent of the total liquid water vaporizes to steam when erupted (at constant enthalpy) to atmospheric pressure (Fig. 2). The remaining 89 percent of the erupted mass is liquid; the large content of liquid water produces effects that are similar to those of the early stages of geyser eruptions. During an eruption of Y-11, however, the local supply of liquid water was soon nearly exhausted and steam became completely dominant. We estimated that the steam was associated with less than 10 percent of liquid water

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by weight. Although at no time did the hole discharge dry steam free of liquid water, we are confident that a dry discharge would have occurred if the eruption had been permitted to continue or if the hole had been cased a little deeper. (The hole was uncased below 27.4 m, and the bottom-hole temperatures indicate an original dominance of liquid water in pore spaces to depths of about 73 m; curves C and E, Fig. 6.)

The pressure of 12.7 kg/cm² measured in the drill rods on May 23 at the greatest drilled depth represents the approximate total pressure at the drill bit, assuming vapor-filled drill rods raised the usual 3 to 4½ m above bottom (1 to 1½ lengths of drill rods), and neglecting the weight of the vapor. If 3.7 m off bottom is assumed, with liquid water filling the hole below the rods, the calculated bottom-hole pressure was about 13.1 kg/cm² (with a possible range from about 12.8 to 13.6 kg/cm²). The pressure at the bottom of an open hole 105.8 m deep and filled with water everywhere just at boiling should be 10.5 kg/cm². Thus, the excess pressure above hydrostatic was about 2.6 kg/cm² or 25 percent. The fact that temperatures and pressures are higher than those of a simple hydrostatic control is important and must be consistent with any satisfactory general model of the vapor-dominated systems.

Liquid-dominated and Vapor-dominated Parts of the System.—In Y-11 drill hole, water-saturated ground evidently extended from the water table at 2.3 m down to a depth of about 73 m. At 72.2 m, the bottom-hole temperature measured 3 hours after drilling ceased was 154.5° C; 18 hours later it had dropped 3° C. We believe that this change was due to the cooling effect of drill water continuing to drain down the hole and into channels that had formerly been dominated by vapor. The pre-drilling ground temperature probably was not attained at this drilled depth and was probably about 165° C (Fig. 6, curve E); flow of water down the hole prevented a normal temperature recovery.

The hole was definitely in vapor-dominated ground at a depth of 93.4 m. At this depth an unanticipated eruption through the drill rods first discharged abundant drill water and then changed rapidly to wet steam with only traces of liquid water. Such a change in behavior is not particularly significant in tight rocks of a hot-water system when the water available for immediate eruption is exhausted; the behavior is similar to that of a geyser as it changes from its main eruptive phase to a steam phase (White, 1967a). However, permeability was so high at all depths below 37 m in Y-11 that little or no drill water returned to the surface. Lack of permeability clearly does not explain the observed eruptive behavior; a limited supply of *available liquid*

water provides the only reasonable alternative. If all lost drill water had remained in nearby permeable ground, the eruption likewise could not have been so nearly dry. The drill water must have percolated down former vapor-filled channels to become unavailable in supporting the eruption.

Forty-six days after completion of the hole, measurements made by an in-hole sampling device (Fournier and Truesdell, 1970) demonstrated that the hole was filled with vapor to 96.4 m, where caving had occurred. Presumably all drill water was then exhausted and all inflowing pore water from higher levels either evaporated completely or escaped downward through former vapor-filled channels.

From these data we can conclude that vapor pressure in the hot core of the system below about 76 m is now significantly above hydrostatic pressure (Fig. 6). Some vapor is being forced upward and outward into the cooler walls. The excess driving pressure above hydrostatic presumably is dispersed in overcoming the frictional resistance to flow of vapor along narrow channelways, which become increasingly clogged upward and outward with liquid water condensed from steam; some of the gases other than steam dissolve in this liquid condensate. If many large free-flowing channels vented to the surface as fumaroles and mud volcanoes, the high vapor pressures in excess of hydrostatic obviously could not be maintained.

Another factor that may be of major importance in impeding the escape of vapor is the formation of montmorillonite and kaolinite, which are the dominant alteration products in rocks and fracture fillings of Y-11 drill core from about 15 to 58 m. Montmorillonite and kaolinite also occur sporadically at greater depths but are generally less abundant than other hydrothermal minerals and unaltered rock silicates. The condensed steam is saturated with CO₂ and other gases from the rising vapor. This carbonated water, represented by analyses 6 and 7 of Table 1, is highly effective in altering feldspars and other silicates to clay minerals, and in leaching cations from the rocks. Pyrite is also relatively abundant through the same general interval from 18 to 61 m, but is sporadic at greater depths. Much sulfide from the rising H₂S evidently dissolves in the condensate and becomes fixed, combining with Fe of the rocks.

The hot vapor-dominated core of the system evidently is not sharply separated by a single fluid interface from the cooler liquid-dominated walls. We conclude that, in the core of the system, the largest fractures and open spaces are mostly or entirely filled with vapor but open spaces of similar dimensions in the margins of the system are largely filled with liquid water, except for dispersed clays

with the
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and vapor bubbles that sporadically rise through the water.

General Model of Vapor-dominated Geothermal Systems

A vapor-dominated geothermal system must normally develop from water-saturated rocks. This statement may be unconvincing for young volcanic rocks (how do we know that such rocks were ever water-saturated?) but is irrefutable for old marine sediments that are now far below the regional water table, as in Tuscany and The Geysers. A new regime is initiated with the introduction of a local potent source of heat at depth (probably a body of magma). Much heat is transferred via conduction and circulating water into surrounding rocks that

have some permeability. Because of thermal expansion and resulting decrease in density of the heated water, a hot-water convection system is then initiated. Most rocks seem to be sufficiently permeable to persist as hosts for hot-water systems; the rate of flow of water remains high enough and the supply of conducted heat below the circulation system remains low enough for most of the water flowing through the system to remain liquid. Near-surface temperatures in the hotter systems, however, are high enough for some boiling to occur as the water rises to intersect the boiling point curve (A of Fig. 6). The depth where boiling first occurs in the rising water depends mainly on the temperature of the water.

Many hot-water systems are to a major extent self-regulating. With more heat flow, the upflowing

This paleogeotherm of a vapor-dominated system could be the result of a hot water sys.

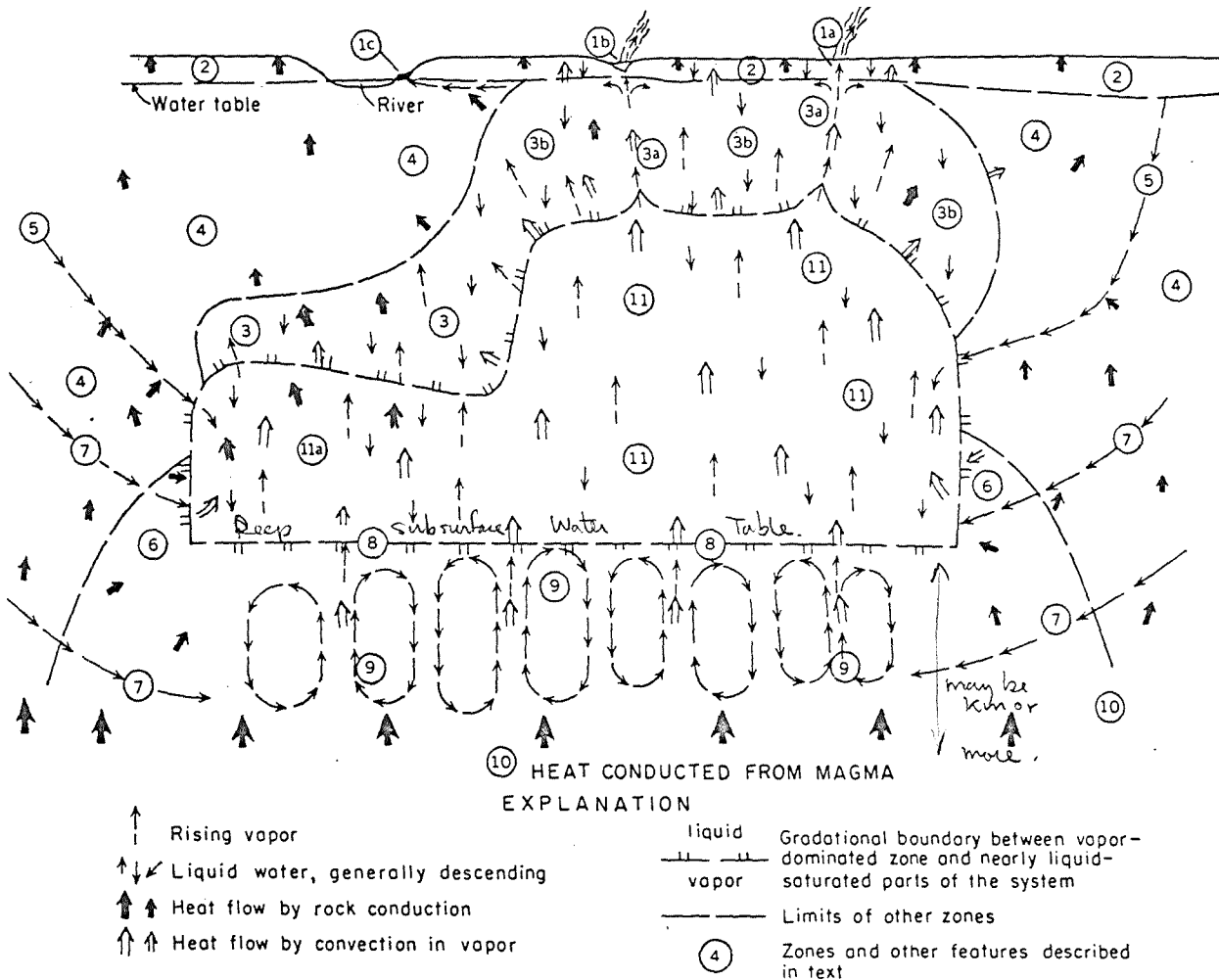


FIG. 7. Model of dynamic vapor-dominated geothermal reservoir surrounded by water-saturated ground. The most significant parts of the model, inward and downward by number, are: 4) zone of conductive heat flow; 3) zone of condensation of steam (conductive and convective heat flow equally important); 11) main vapor-dominated reservoir, with convective upflow of heat in steam in larger channels, and downflow of condensate in small pores and fractures (surface tension effects); 9) deep zone of convective heat transfer, probably in brine; 10) deep zone of conductive heat flow (too hot for open fractures to be maintained). Other features are discussed in text.

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water becomes hotter and lower in density and viscosity; the pressure drive for recharge increases, and the increased rate of through-flow removes most of the additional heat. This self-regulation, however, may be limited by insufficient permeability.

With sufficiently potent heat supply or, for any reason, a decreasing rate of recharge of water, a hot-water system of limited permeability may start to boil off more water than can be replaced by inflow. A *vapor-dominated system* then starts to form. Direct evidence for the assumed initial dominance of liquid water is lacking for the major vapor-dominated systems. Hot-spring sinter constitutes the clearest evidence, and is so interpreted for the Mud Volcano system. However, sinter is deposited only from very hot water that flows so rapidly to the surface that little SiO_2 precipitates en route. In addition, the early hot-water stage of these systems of high heat flow and low permeability is likely to have been brief and their thin sinter deposits (if any) are likely to be destroyed by erosion.

Figure 7 is our tentative general model of a well-developed vapor-dominated system. Different parts, discussed below, are keyed by number on the figure.

(1) Fluids that discharge at the surface provide much of the evidence for a vapor-dominated system. Fumaroles (1a) are generally at temperatures near surface boiling or somewhat lower. High-chloride springs are completely absent; associated springs and mud pots are generally acid, high in sulfate, and low in discharge (such as 1b of Fig. 7, and anal. 2 and 5, Table 1), and they deposit little if any sinter. Surrounding ground may be bleached and lacking in vegetation. Some springs not so strongly influenced by oxidation of H_2S (or containing enough NH_3 absorbed from gases) are nearly neutral in pH and are dominated by bicarbonate and sulfate without much chloride (1c, Fig. 7, and anal. 1 and 6, Table 1).

(2) Zone 2 lies between the ground surface and the water table.⁵ Where hot enough, steam and other gases rise above the water table, as in auger hole W-1 of Figure 5. At the water table heat transfer is nearly all convective, but as the temperature gradient increases upward and water vapor condenses, near-surface heat transfer becomes largely conductive.

(3) Zone 3 inhibits the free escape of rising vapor. The zone is nearly saturated with liquid water derived largely from condensing steam rich in CO_2 . Montmorillonite and kaolinite form by reaction of this

CO_2 -saturated condensate with rock silicates. Clay minerals and condensed water clog most pore spaces and channels, impeding but in many places not prohibiting the escape of residual uncondensed gases. Temperatures in this zone may be similar to those along the hydrostatic reference curve A of Figure 6. Near major channels of upflowing steam (3a, Fig. 7), temperatures and pressures are somewhat above hydrostatic, and conductive heat flow and condensation of steam are consequently high; at least part of the condensate is swept upward to the water table or to surface springs, mud pots, and mud volcanoes. A crude steady-state rate of upflow is determined by pressure gradients, dimensions of the channels, strength of wallrocks, and impedance provided by condensate and suspended clays. Other parts of zone 3 (3b, Fig. 7) are dominated by downflowing condensate and some surface water, with temperatures that are likely to be somewhat lower than those along reference curve A of Figure 6. As temperature gradients in general increase outward and upward through zone 3, more of the heat of vaporization in the rising steam can be transferred by conduction, so water vapor is continuously condensing and the rate of mass flow of vapor therefore decreases upward. A part of the heat in rising vapor is transferred through local horizontal gradients to heat the downward-percolating condensate, which must absorb heat as it descends into hotter ground. The dashed line bounding the outer part of zone 3 marks the gradation in mode of heat transfer from dominantly convective to dominantly conductive.

The lower ^{outer} limit or "pinch-out" of zone 3 is at a depth where the hydrostatic pressure of water in the reservoir margins exceeds the total vapor pressure of steam and gases in the reservoir. Below this depth, vapor can no longer effectively penetrate ^{outward along} the reservoir margin. ?

Wells drilled into parts of zone 3 may produce liquid dominantly, but if drilled and cased into deeper parts they probably yield wet steam and some water when first produced (as in Y-11 drill hole). If an uncased section of hole intercepts channels of upflowing steam and zones of cooler downflowing condensate, the temperature and pressure of the steam will commonly dominate the hole. This occurred in Y-11 below 72 m.

(4) Zone 4 is characterized mainly by conductive heat flow, with heat being supplied from condensing steam within zone 3. Wells bottomed in zone 4 may fill with water, and may erupt hot water and some steam, but discharge rates are likely to be low and the wells noncommercial.

(5) Representative channels of intermediate-level recharge are deep enough at points of entry for hydrostatic pressure to exceed the vapor pressure of

⁵ In sands and gravels the water table is easily recognized. In clays, however, the water table is poorly defined, but we consider it to be the level at which water is maintained in a shallow open hole. The zone of saturation can rise as much as 10 m above this level, owing to surface tension in the clays. Hydrostatic pressure increases downward only below the water table as defined in the open hole.

about 31 to 35 kg/cm² in the main reservoir (zone 11).

Channels of inflow tend to be enlarged by solution of SiO₂ as the inflowing water is heated by conduction (indicated by heat-flow arrows in Fig. 7). Channels are diminished, however, by deposition of CaCO₃ and CaSO₄, which are rather unusual in decreasing in solubility with increasing temperature (see, for example, Holland, 1967). In all rocks with recharge waters relatively high in CaCO₃ and CaSO₄, channel permeabilities are especially likely to decrease rather than increase with time. These considerations may be important in understanding Larderello, which involves anhydrite-bearing limestone and shales, and The Geysers, where mafic lavas and serpentine are associated with graywacke and shale.

(6) Zone 6 consists of reservoir margins where temperatures decrease toward the reservoir. The depth of the top of zone 6 is not easily predicted. If there were no convective heat flow, the depth would be near that of the 240° C isotherm of the original conductive gradient from the surface to the magma chamber. If 600° C is assumed at 4 km, for example, and the rocks are homogeneous, 240° C would be at 1.6 km depth. The development and downward penetration of the main vapor-dominated reservoir as excess pore water is vaporized result in extensive convective modifications of temperature that greatly change the relationships. Convective cooling from downflowing meteoric water increases this depth, and a shallower intrusion at higher temperature decreases it. These reservoir margins contain channels of inflowing water at pressures that are close to hydrostatic and much greater than ~33 kg/cm² of the reservoir. Sharp pressure and temperature gradients decreasing toward the reservoir must therefore exist in zone 6. In contrast to zone 3, heat is transmitted through zone 6 by conduction (and inflowing H₂O) to the reservoir. The temperatures of zone 6 grade downward into, and are maintained by conduction from zone 10.

(7) Channels of inflowing water are narrowed by precipitation of calcite and anhydrite as zone 6 is approached; clogging of channels by these minerals of decreasing solubility may be offset entirely or in part by solution of quartz, which increases in solubility as long as the liquid water continues to rise in temperature. At the outer edge of zone 6, however, pressures and temperatures in the recharge channels attain their maxima; with further flow toward the reservoir, boiling commences and temperature declines as the pressure drops to that of the reservoir. The fluid in these channels is now a two-phase mixture of steam and water. Specific resistance to flow (resistance per unit of mass) of steam is much

greater than that of liquid water, and specific flow resistance of a two-phase mixture is greater than a linear combination would indicate (I. G. Donaldson and Gunnar Bodvarsson, oral commun., 1970). Because of evaporative concentration by boiling and because of decreasing temperature, quartz and other minerals are now deposited, further impeding the flow of the two-phase mixture. The result of all of these processes is to decrease the rate of recharge through the deeper channels.

(8) The deep subsurface water table recedes as long as the heat supply is sufficient for net loss of liquid water and vapor from the system to exceed net inflow (water table shown in Fig. 7 is horizontal, but it may be very irregular in detail). As mentioned above, recharge tends to decrease with time as resistance to flow of H₂O through individual channels increases. As the water table recedes and liquid water in the reservoir is largely replaced by vapor at nearly constant pressure throughout the reservoir, the driving pressure on the deeper channels of inflow increases, offsetting in part the increasing impedances. A crude steady state may be attained in some systems, especially as rate of heat flow eventually starts to decline.

(9) With time, if not initially, the water boiling below the deep water table becomes a brine as recharging water boils off and as dissolved substances of low volatility are residually concentrated. Vapor from brine is superheated with respect to pure water at the same pressure. Steam boiling from 25 percent NaCl brine at 35 kg/cm², for example, is superheated by about 12° C with respect to saturated steam and pure water (254° vs 242° C, Haas, 1970). The critical temperature of a salt solution increases above that of pure water (374° C) as salinity increases; that of a 1 percent NaCl solution is about 384° C (Sourirajan and Kennedy, 1962, p. 134); that of a 10 percent solution is about 480° C; and that of a 25 percent solution is about 675° C. Thus, brine can be a very effective agent for convective transfer of heat and dissolved matter at temperatures much above 374° C. Note that Figure 7 has no vertical scale; the depth of zone 9 may be 1,000 m or more, and through all or most of this depth, pressures are lower than hydrostatic pressures outside the system (Fig. 1, curve C, increases downward in slope).

(10) Conductive heat flow from the magma predominates deep under the reservoir where rock plasticity due to increasing temperature prevents the maintenance of open channels. On the outer margins of zone 10 where convective disturbance is not so severe, conductive heat flow predominates to higher levels than under zone 9, grading upward without distinct boundaries into zones 6 and 4. The amount of convective circulation may eventually de-

Important factor in the economics because could affect life span of utilization of the resource

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crease beneath the vapor-dominated reservoir by decreased permeability from deposition of minerals, and possibly as a stable salinity gradient becomes established.

11) The main vapor-dominated reservoir contains liquid water and vapor coexisting, except possibly in major channels of steam discharge and locally just above the brine water table. Steam and other gases rise in the largest channels where resistance to flow is lowest. Steam starts to condense on the outer borders of the reservoir and continues to condense from all vapor escaping into zone 3, where temperatures decrease outward and provide a thermal gradient for conductive transfer of the heat of vaporization of steam. The condensate from zone 3 percolates down into the reservoir, favoring narrow channels and pore spaces between mineral grains because of surface tension and the lower specific resistance to flow of liquid water relative to steam.

Edwin Roedder (personal commun., 1970) has suggested that our model for vapor-dominated systems is similar in many respects to recently-developed remarkable devices that have been called "heat pipes" (Eastman, 1968). These devices may be "several thousands of times more efficient in transporting heat than the best metallic conductors." They consist of a closed chamber with inside walls lined by a capillary structure or wick, and saturated with a volatile fluid. Heat is transferred by vapor from the hotter to the cooler end, where the vapor condenses. The liquid condensate returns by capillary action to the evaporator section; temperature gradients in the pipe may be extremely low. The top end may be the hotter, with capillary return of liquid (to some limited height) being opposed by gravity. Our natural "heat pipes" are not completely closed systems, and their depth has no theoretical limit because gravity assists rather than opposes the return flow of condensate.

Parts of the subsurface reservoir such as 11a of Figure 7 may be isolated from direct outflow of vapor and may be representative of parts of the Larderello and The Geysers systems that have no apparent direct discharge in fumaroles. Pressures throughout the reservoir are controlled primarily by the total vapor pressure at the boiling water table, modified by frictional resistance to the upward flow of vapor and by the weight of the vapor. Near the top of the reservoir the vapor may be greatly enriched in CO₂, H₂S, and other gases that are not flushed out of the system as actively as near the top of the main reservoir (11). Much water vapor condenses below the boundary of the vapor-dominated reservoir near 11a. In contrast to the flushed part of the main reservoir, significant thermal gradients exist in the poorly flushed parts. Consequently,

Table 5.--Saturation temperatures of water calculated for ideal steam-gas mixtures at constant vapor pressure, 31.8 kg/cm².

Percent steam by vol.	Percent other gases by vol.	Pressure, kg/cm ²	Saturation temp., °C
100	0	31.8	236
99	1	31.5	235.5
98	2	31.2	234.9
95	5	30.2	233.1
90	10	28.6	230.1
80	20	25.4	223.7
70	30	22.3	216.9
50	50	15.9	200.1
30	70	9.5	176.8
10	90	3.2	134.7
5	95	1.6	113.0
1	99	0.3	68.0

some steam can condense and other gases are residually concentrated. Pressure of the remaining water vapor requires lower saturation temperatures, as shown in Table 5. This table suggests that temperatures in isolated parts of the reservoir differ little from 236° C until the residual gases are enriched above 5 percent. With higher residual gas contents, temperature gradients and conductive heat flow increase.

The above-described relationships may explain the relatively high pressures and low temperatures of the vapor-dominated fields of Bagnore and Piancastagnaio near Monte Amiata (Burgassi and others, 1965; Cataldi, 1967). Initial pressures were 22 to 40 kg/cm² and gas contents of the vapor were as high as 96 percent, but reported temperatures did not exceed about 150° C (Burgassi, 1964; Burgassi and others, 1965; Cataldi, 1967). Pressures and gas contents of the vapor decreased rapidly with production.

Similar reasoning indicates that high contents of gas in vapor coexisting with liquid water at a temperature near that of the maximum enthalpy of steam can result in total vapor pressure significantly above 31.8 kg/cm² at 236° C (Table 6). These data indicate that, as contents of other gases increase in the vapor phase at constant temperature of liquid and vapor, total pressures must increase. The least

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as gas contents increase upward the saturation temp. (temp. of condensation) decreases.

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Table 6.--Total vapor pressures of steam-gas mixtures coexisting with liquid water at 236°C

Vol. percent gas in vapor	H ₂ O pressure kg/cm ²	Pressure other gases, kg/cm ²	Total pressure, kg/cm ²
0	31.8	0	31.8
1	31.8	0.3	32.1
2	31.8	0.7	32.4
5	31.8	1.7	33.5
10	31.8	3.5	35.3
20	31.8	7.9	39.8
50	31.8	31.8	63.6

Note: Pressures 32.4 and 39.8 in column 4 should read 32.5 and 39.7 respectively.

actively flushed extensions of The Geysers field that have recently been discovered are likely to have higher gas contents and initial pressures than the original field. *what is P. not controlled by P_{vapor} @ boiling water table?*

Table 6 also suggests a possible triggering mechanism for some hydrothermal explosions and phreatic eruptions (Muffler and other, 1970) in gas-rich hot-spring and volcanic systems where escape of vapor and flushing of residual gases are inhibited by barriers of low permeability. Local accumulations of gas-rich vapor can attain pressures that exceed hydrostatic and perhaps even lithostatic, finally resulting in rupture and explosive eruption.

The tentative model described above has additional support from thermodynamic calculations and comparison of actual production data with production predicted on the basis of our model (manuscript in preparation). We are hopeful that the model will prove to be of value in predicting the behavior of individual wells, in detecting interference between wells, in detecting inhomogeneities within the reservoir, in calculating reserves of steam in the original vapor-dominated reservoir, and in detecting a major influence by increased boiling below the water table as a result of declining reservoir pressures.

Speculations Relating Vapor-dominated Systems and Ore Deposits

Some mercury deposits may have formed in the upper parts of vapor-dominated systems. We also suggest, more tentatively, that porphyry copper de-

posits may have formed in the deep brine zones hypothesized to underlie vapor-dominated reservoirs.

Mercury Deposits.—Many mercury deposits appear to have formed near the surface in relatively recent time. Furthermore, mercury deposits occur on the periphery of two active vapor-dominated geothermal systems: The Geysers in California and Monte Amiata in Italy (White, 1967b; Dickson and Tunell, 1968). Recent geothermal exploration for extensions of The Geysers field disclosed dry steam 2½ km to the west under the Buckman mercury mines. Other wells yield dry steam near Anderson Springs, only 1½ km from the Big Chief and Big Injun mercury mines (White, 1967b), which are 10 km southeast of the original steam field. A number of other mercury mines in the district are within 3 km of steam wells.

Vapor-dominated systems of high gas content, previously discussed, have recently been discovered from 3 to 10 km south and southwest of the major Monte Amiata mercury mine (Burgassi and others, 1965; Cataldi, 1967), the largest Italian mercury deposit. No vapor-dominated reservoir has been found to prove a genetic relation to the mercury deposits, although abnormally high temperatures (63°C at 440 m depth) and notable concentrations of CO₂ and H₂S characterize these Italian deposits (White, 1967b). Dall'Aglio and others (1966) have shown that mercury occurs in anomalous amounts (> 1 ppm) in stream sediments in and around the Lardello-Monte Amiata fields. The anomalies in some stream drainages may be related to specific mercury deposits, but many clearly are not. These widespread anomalies instead seem more directly related to the geothermal fields and their broad anomalies in temperature gradient (Burgassi and others, 1965; Fig. 7).

Krauskopf (1964) has emphasized the high volatility of mercury, which provides an attractive mechanism for separating this metal from most other metals. The vapor-dominated geothermal systems, as we now understand them, provide a mechanism for shallow, moderately high temperature vapor-phase separation of mercury from other metals. Mercury is known to occur in vapor from The Geysers steam field (White, 1967b, p. 590, and unpub. data), and large mercury anomalies have been found in Yellowstone Park in mudpots of the Mud Volcano area and elsewhere, that are maintained by steam flow and condensation (W. W. Vaughn, U. S. Geol. Survey, written commun., 1969). Especially attractive is the possibility that Hg and H₂S dissolve in the steam condensate of zone 3 of our model (Fig. 7), precipitating as HgS as temperature decreases and as the pH of the condensate increases from reaction with silicates.

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We do not claim that *all* mercury deposits form in this way. The Sulphur Bank and Abbott mines east of The Geysers, for example, are associated with discharging thermal chloride waters that may be, respectively, metamorphic and connate waters being forced out of their source rocks by lithostatic pressure (White, 1957b, 1967b). During peak mineralization at high temperatures, similar water was almost certainly being discharged, perhaps with more abundant vapor than now.

Porphyry Copper Deposits.—The possibility that porphyry copper deposits may be forming in the zone of boiling brine below vapor-dominated systems (zone 9 of Fig. 7) should be tested in these systems by looking for copper minerals in core and cuttings from the deepest drill holes. The model provides attractive possibilities for explaining many aspects of these deposits:

1. Recent isotope studies (Sheppard, Nielsen, and Taylor, 1969) demonstrate that water of meteoric origin probably is dominant over water of other origins during ^{late alteration only?} mineralization stages.

2. Temperatures of filling of fluid inclusions are most commonly above 250° C and exceptionally range up to 725° C (Edwin Roedder, oral and written commun.). The salinities of many inclusions are exceedingly high, probably ranging up to 60 percent of total fluid by weight. However, many inclusions are largely vapor, probably indicating boiling of the saline fluid at the time of entrapment.

3. Fluid relationships and the geologic setting of Copper Canyon, Nevada, are considered to be generally similar to porphyry copper deposits (J. T. Nash, written commun., 1970). Extensive fluid-inclusions studies by Nash and Theodore (1970) demonstrate that a) temperatures are most commonly in the range of 315° to 375° C; b) salinities of the ore fluids are commonly in the order of 40 percent (or higher, if CaCl₂ is abundant), with highest salinities in and near the porphyry intrusion and with lower salinities (2 to 15 percent) in peripheral gold-bearing deposits; c) vapor bubbles were trapped in many inclusions, demonstrating the prevalence of boiling or near-boiling conditions. The copper deposits are largely dispersed in the intruded rocks adjacent to the porphyry, and thus are within the spectrum of deposits that have been called porphyry copper deposits (Lowell and Guilbert, 1970).

4. High-salinity brines can develop from residual concentration of dilute (or saline) recharge water, providing a satisfactory system for transferring heat, metals, sulfur and CO₂ from the large magma body that presumably underlies the small multiple porphyry intrusions of most deposits. The critical temperature of water increases with salinity; with sufficient con-

tents of alkali and calcium chlorides, water can remain liquid at temperatures as high as those of the magma body. Copper and other metals could be derived from the local porphyries, a larger underlying magma chamber, and from surrounding rocks.

5. The return flow of condensate through the vapor-dominated reservoir is relatively dilute, but is normally saturated in SiO₂ (with respect to quartz, 440 ppm at 240° C, Fournier and Rowe, 1966). Reevaporation of this water may account for much of the abundant hydrothermal quartz of porphyry copper deposits.

6. Condensate from the discharge areas of vapor-dominated systems is high in sulfate. Some and perhaps much of this condensate may drain downward to the deep water table and account for the abundant anhydrite of many porphyry copper deposits.

7. The most commonly quoted range in depth for the tops of porphyry copper deposits is from 1,000 to 3,000 meters (Lowell and Guilbert, 1970). The shallower depths seem too low for attaining the indicated temperatures and salinities, but may be possible in a brine below a shallow vapor-dominated reservoir (Fig. 1, curve C, can be at shallower as well as greater than plotted depth).

8. If porphyry copper deposits were indeed formed at depths of 1,000 to 3,000 meters, if most of the water of the ore fluids is of surface origin, as indicated by isotopes, and if near-magmatic temperatures and excess heat flow were maintained close to the surface for thousands of years, some type of hydrothermal activity *must* have characterized the then-existing ground surface. Hot-water systems are numerically far more abundant than vapor-dominated systems, and may be the surface expression of some kinds of ore generation (White, 1967b, 1968a), but dissolved salts are *dispersed* by discharging water, and extreme salinities are not ordinarily attained. The highest salinity yet known in active hot-water systems is about 25 percent, characterizing both the Salton Sea and the Red Sea geothermal brines (White, 1968a). Chemical evidence indicates strongly that the high salinities of these two systems result from the solution of NaCl-rich evaporites. We doubt that evaporites are also involved in the generation of all porphyry copper deposits; some other mechanism for attaining extreme salinity is indicated. Our proposed mechanism for residual concentration of salts by boiling below vapor-dominated systems is a feasible and attractive possibility.

9. The postulated water below a vapor-dominated reservoir may be characterized by high positive temperature and salinity gradients extending downward from the deep water table (Fig. 1), thereby providing a favorable environment for upward transport and

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deposition of copper sulfides and pyrite. Temperatures in the water-dominated zone must increase toward the source of heat, presumably an igneous intrusion; actual gradients are highly dependent on the extent of convection in this zone. Formation of vapor bubbles probably occurs largely near the base of penetration of water of the system, where temperatures are highest relative to pressure. Salinity is thereby increased by residual concentration near the base, where permeability is low enough to inhibit convection. On the other hand at higher levels near the deep water table, dissolved salts are being diluted by three processes: (a) *condensation* of dilute water from steam bubbles rising in the brine, as pressures decrease to about 34 kg/cm², as discussed above; (b) downward percolation of *condensate* of steam from the upper margins of the vapor-dominated reservoir; and (c) entry of *new water* recharging the system; this water is likely to be considerably more dilute than the average deep water.

Porphyry copper deposits should be reexamined with consideration of these speculations on temperatures and salinities. If temperatures and salinities do increase sharply downward, our model may provide a new understanding of mode of transport and deposition of the ore minerals. Both decreasing temperature and decreasing salinity upward should favor precipitation of copper sulfides because of the decreasing stability of copper chloride complexes. Introduction of the ore metals may normally occur during a late stage in the total activity after very high salinities have been attained from residual concentration by boiling, and perhaps after the deepest permeable fractures (zone 9 of Fig. 7) have extended downward into a partly cooled major magma chamber.

Porphyry copper deposits should also be examined to determine whether the primary deposits were limited in upward development by a subsurface water table (8 of Fig. 7). Copper and other base metals have low volatilities and could not be transferred into the vapor-dominated reservoir. Pyrite and cinnabar are likely to be characteristic of the zone of condensation (zone 3), and pyrite can also form within the reservoir (zone 11) by reactions involving H₂S and Fe of the rocks. However, pyrite is likely to be much more abundant below the brine water table. Thus, where the original upper limit of copper mineralization and the level of the former brine water table are exposed in the present topography, the water table may be indicated by an anomalous upward *decrease* in supergene oxidation where pyrite was initially so scarce.

U. S. GEOLOGICAL SURVEY,
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REFERENCES

- Allen, E. T., and Day, A. L., 1927, Steam wells and other thermal activity at "The Geysers," California: Carnegie Inst. Washington Pub. 378, 106 p.
- , 1935, Hot springs of the Yellowstone National Park: Carnegie Inst. Washington Pub. 466, 525 p.
- Baba, Kenzo, 1968, On the temperature and pressure measurements of the geothermal well Matsukawa No. 4: Japan Geol. Survey Bull., v. 19, no. 11, p. 725-728.
- Banwell, C. J., Cooper, E. R., Thompson, G. E. K., and McCree, K. J., 1957, Physics of the New Zealand thermal area: New Zealand Dept. Sci. and Indus. Research Bull. 123, 109 p.
- Barnes, H. L., and Czamanske, G. K., 1967, Solubilities and transport of ore minerals, in Barnes, H. L., Editor, Geochemistry of Hydrothermal Ore Deposits: New York, Holt, Rinehart, and Winston, Inc., p. 334-381.
- Bodvarsson, Gunnar, 1964a, Physical characteristics of natural heat resources in Iceland, in Geothermal Energy I: United Nations Conf. New Sources Energy, Rome, 1961, Proc., v. 2, p. 82-90.
- , 1964b, Utilization of geothermal energy for heating purposes and combined schemes involving power generation, heating, and/or by-products, in Geothermal Energy II: United Nations Conf. New Sources Energy, Rome, 1961, Proc., v. 3, p. 429-436.
- , 1970, Evaluation of geothermal prospects and the objectives of geothermal exploration: Geoexploration, v. 8, p. 7-17.
- Burgassi, Renato, 1964, Prospecting of geothermal fields and exploration necessary for their adequate exploitation performed in the various regions of Italy, in Geothermal Energy I: United Nations Conf. New Sources Energy, Rome, 1961, Proc., v. 2, p. 117-133.
- , Cataldi, R., Mouton, J., and Scandellari, F., 1965, Prospezione delle anomalie geotermiche e sua applicazione alla regione Amiatina: l'Industria Mineraria, v. 16, p. 1-15.
- Cataldi, R., 1967, Remarks on the geothermal research in the region of Monte Amiata (Tuscany-Italy): Bull. Volcanol., v. 30, p. 243-270.
- , Ferrara, G. C., Stefani, G., and Tongiorgi, E., 1969, Contribution to the knowledge of the geothermal field of Larderello (Tuscany, Italy). Remarks on the Carboli area: Bull. Volcanol., v. 33, no. 1, p. 1-27.
- Craig, H., 1963, The isotopic geochemistry of water and carbon in geothermal areas, in Nuclear geology on geothermal areas, Spoleto: Consiglio Nazionale delle Ricerche, Laboratorio di geologia nucleare, Pisa, p. 17-53.
- , 1966, Superheated steam and mineral-water interactions in geothermal areas: Am. Geophys. Union Trans., v. 47, p. 204-205.
- , Boato, Giovanni, and White, D. E., 1956, Isotopic geochemistry of thermal waters: Natl. Acad. Sci., no. 400, p. 29-38.
- Dall'Aglio, M., DaRoit, R., Orlandi, C., and Tonani, F., 1966, Prospezione geochemical del mercurio—distribuzione del mercurio nelle alluvioni della Toscana: Industria Mineraria, v. 17, p. 391-398.
- Dickson, F. W., and Tunell, George, 1968, Mercury and antimony deposits associated with active hot springs in the western United States, in Ridge, J. D., Editor, Ore deposits of the United States 1933-1967 (Graton-Sales volume), v. 2: New York, Am. Inst. Mining Metall. Petroleum Engineers, Inc., p. 1673-1701.
- Eastman, G. Yale, 1968, The heat pipe: Scientific American, v. 218, p. 38-46.
- Elder, J. W., 1965, Physical processes in geothermal areas, in W. H. K. Lee, Editor, Terrestrial heat flow: Am. Geophys. Union Mon., ser. 8, p. 211-239.
- Ellis, A. J., and Mahon, W. A. J., 1964, Natural hydrothermal systems and experimental hot-water/rock interactions: Geochim. et Cosmochim. Acta, v. 28, p. 1323-1357.
- , 1967, Natural hydrothermal systems and experimental hot water/rock interactions (Pt. II): Geochim. et Cosmochim. Acta, v. 31, no. 4, p. 519-538.
- Facca, G. technol. p. 143-144, 1968, 1969, canol., v. Fenner, C. Park: J. Ferrara, C. isotopic steam j. geology. Naziona. cleare, 1 Fournier, ground: hot spri. no. 9, p. —, and down-he wells: J. Haas, J. L. thermal pressure Hayakawa Volcano. Helgeson, teristics. Sci., v. 2. Holland, 1 deposits, thermal Winston. James, Ru power s. Technol. Keenan, J. Properti. 89 p. Krauskopf. compound Lee, W. F. data, in Am. Geol. Lowell, J. alteration Econ. G. Marinelli, areas of McNitt, J. thermal j. ogy Spec Mercado, S. M-20, C. v. 80, p. Muffler, L. Hydrothe Park: G.

- Facca, Giancarlo, and Tonani, Franco, 1964, Theory and technology of a geothermal field: *Bull. Volcanol.*, v. 27, p. 143-189.
- , 1967, The self-sealing geothermal field: *Bull. Volcanol.*, v. 30, p. 271-273.
- Fenner, C. N., 1936, Bore-hole investigations in Yellowstone Park: *Jour. Geol.*, v. 44, no. 2, pt. 2, p. 225-315.
- Ferrara, G. C., Ferrara, G., and Gontfanti, R., 1963, Carbon isotopic composition of carbon dioxide and methane from steam jets of Tuscany, in Tongiorgi, Editor, Nuclear geology on geothermal areas, Spoleto 1963: Pisa, Consiglio Nazionale Delle Ricerche, Laboratorio di Geologia Nucleare, p. 277-284.
- Fournier, R. O., and Rowe, J. J., 1966, Estimation of underground temperatures from the silica content of water from hot springs and wet-steam wells: *Am. Jour. Sci.*, v. 264, no. 9, p. 685-697.
- , and Truesdell, A. H., 1970, A device for measuring down-hole pressures and sampling fluids from geothermal wells: *Jour. Sci. Instruments* [in press].
- Haas, J. L., Jr., 1970, The effect of salinity on the maximum thermal gradient of a hydrothermal system of hydrostatic pressure and near boiling: *ECON. GEOL.*, in press.
- Hayakawa, M., 1969, Anomalous heat flow in Japan: *Bull. Volcanol.*, v. 33, no. 1, p. 57-68.
- Helgeson, H. C., 1968, Geologic and thermodynamic characteristics of the Salton Sea geothermal system: *Am. Jour. Sci.*, v. 266, p. 129-166.
- Holland, H. D., 1967, Gangue minerals in hydrothermal deposits, in H. L. Barnes, Editor, *Geochemistry of Hydrothermal Ore Deposits*: New York, Holt, Rinehart, and Winston, Inc., p. 382-436.
- James, Russell, 1968, Wairakei and Larderello; geothermal power systems compared: *New Zealand Jour. Sci. and Technology*, v. 11, p. 706-719.
- Keenan, J. H., and Keyes, F. G., 1936, *Thermodynamic Properties of Steam*: New York, John Wiley and Sons, 89 p.
- Krauskopf, K. B., 1964, The possible role of volatile metal compounds in ore genesis: *ECON. GEOL.*, v. 59, p. 22-45.
- Lee, W. H. K., and Uyeda, S., 1965, Review of heat flow data, in W. H. K. Lee, Editor, *Terrestrial heat flow*: *Am. Geophys. Union, Geophys. Mon.* 8, p. 87-190.
- Lowell, J. D., and Gilbert, J. M., 1970, Lateral and vertical alteration-mineralization zoning in porphyry ore deposits: *ECON. GEOL.*, v. 65, p. 373-408.
- Marinelli, G., 1969, Some geological data on the geothermal areas of Tuscany: *Bull. Volcanol.*, v. 33, no. 1, p. 319-334.
- McNitt, J. R., 1963, Exploration and development of geothermal power in California: *Calif. Div. Mines and Geology Spec. Rept.* 75, 45 p.
- ~~Mercado, Sergio~~ 1969, Chemical changes in geothermal well M-20, Cerro Prieto, Mexico: *Geol. Soc. America Bull.*, v. 80, p. 2623-2629.
- Muffler, L. J. P., White, D. E., and Truesdell, A. H., 1971, Hydrothermal explosion craters in Yellowstone National Park: *Geol. Soc. America Bull.* [in press].
- Nakamura, Hisayoshi, and Sumi, Kiyoshi, 1967, Geological study of Matsukawa geothermal area, northeast Japan: *Japan Geol. Survey Bull.*, v. 18, no. 2, p. 58-72.
- Otte, Carel, and Dondanville, R. F., 1968, Geothermal developments in The Geysers area, California [abs.]: *Am. Assoc. Petroleum Geologists Bull.*, v. 52, p. 575.
- Penta, Francesco, 1959, *Sulle Origini del Vapore Acqueo Naturale e Sull' attuale Stato delle Relative Ricerche (Ricerche per "forze Endogene")*: *La Ricerca Scientifica*, v. 29°, num. 12, p. 2521-2536.
- Saito, Masatsugu, 1964, Known geothermal fields in Japan, in *Geothermal Energy I: United Nations Conf. New Sources Energy*, Rome, 1961, Proc., v. 2, p. 367-373.
- Schoen, Robert, and Ehrlich, G. G., 1968, Bacterial origin of sulfuric acid in sulfurous hot springs: *Internat. Geol. Cong.*, 23d, Prague, v. 17, p. 171-178.
- Sheppard, S. M. F., Nielsen, R. L., and Taylor, H. P., Jr., 1969, Oxygen and hydrogen isotope ratios of clay minerals from porphyry copper deposits: *ECON. GEOL.*, v. 64, p. 755-777.
- Sourirajan, S., and Kennedy, G. C., 1962, The system H₂O-NaCl at elevated temperatures and pressures: *Am. Jour. Sci.*, v. 260, p. 115-141.
- White, D. E., 1957a, Thermal waters of volcanic origin: *Geol. Soc. America Bull.*, v. 68, no. 12, p. 1637-1658.
- , 1957b, Magmatic, connate, and metamorphic waters: *Geol. Soc. America Bull.*, v. 68, p. 1659-1682.
- , 1964, Preliminary evaluation of geothermal area by geochemistry, geology, and shallow drilling, in *Geothermal Energy I: United Nations Conf. New Sources Energy*, Rome, 1961, Proc., v. 2, p. 402-408.
- , 1967a, Some principles of geyser activity, mainly from Steamboat Springs, Nevada: *Am. Jour. Sci.*, v. 265, p. 641-648.
- , 1967b, Mercury and base-metal deposits with associated thermal and mineral waters, in Barnes, H. L., Editor, *Geochemistry of Hydrothermal Ore Deposits*: New York, Holt, Rinehart, and Winston, Inc., p. 575-631.
- , 1968a, Environments of generation of base-metal ore deposits: *ECON. GEOL.*, v. 63, no. 4, p. 301-335.
- , 1968b, Hydrology, activity, and heat flow of the Steamboat Springs thermal system, Washoe County, Nevada: *U. S. Geol. Survey Prof. Paper* 458-C, p. C1-C109.
- , 1969, Rapid heat-flow surveying of geothermal areas, utilizing individual snowfalls as calorimeters: *Jour. Geophys. Research*, v. 74, no. 22, p. 5191-5201.
- , Brannock, W. W., and Murata, K. J., 1956, Silica in hot-spring waters: *Geochim. et Cosmochim. Acta*, v. 10, p. 27-59.
- , Hem, J. D., and Waring, G. A., 1963, Chemical compositions of subsurface waters, in *Data of geochemistry*: *U. S. Geol. Survey Prof. Paper* 440-F, 67 p.
- Wilson, S. H., 1955, Chemical investigations, in Grange, L. I., Editor, *Geothermal steam for power in New Zealand*, chap. 4: *New Zealand Dept. Sci. and Indus. Research Bull.* 117, p. 27-42.